

GEOLOGY

and Economic Minerals of Canada

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GEOLOGY AND ECONOMIC MINERALS OF CANADA

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REPORT No. 1

GEOLOGY AND ECONOMIC MINERALS OF CANADA

R. J. W. Douglas,
Scientific Editor

DEPARTMENT OF
ENERGY, MINES AND RESOURCES
CANADA

PREFACE

This is the fifth edition of *Geology and Economic Minerals of Canada,* a general review of our present knowledge of the geology of Canada. It has been possible for the first time to treat many different aspects of the geology in considerable detail on a nation-wide basis, and to attempt a tectonic synthesis of the four major geological regions — the Canadian Shield, and the bordering Phanerozoic platforms and orogens of southeastern, western, and arctic Canada. The accompanying folio contains a geotectonic correlation chart for each of those regions and eight maps at 1:5 million scale dealing with the geology, tectonics, mineral deposits, glacial features, physiography, isotopic ages, and magnetic and gravity anomalies. In the text, factual data are concisely summarized and economic and scientific interpretation and conclusions are clearly separated.

Each of the five editions represents a report of progress, but still the need and challenge to acquire more information remain. Geological knowledge is fundamental to the evaluation of the nation's potential mineral resources and to the search for mineral deposits, both important factors in Canada's general economic growth and in the development of our vast northland. Familiarity with geological structure is essential to engineering construction, groundwater supply, management of urban and rural land resources — both renewable and non-renewable — and for the conservation of man's environment. This book is a storehouse of cultural knowledge on the physical make-up of Canada, and will serve as a reference text for all scientists.

Y. O. FORTIER,

Director, Geological Survey of Canada

OTTAWA, August 14, 1968

COVER Malak

ENDLEAVES

Front: Spheroidal andesite, eastern Hudson Bay (F. C. Taylor, 125219)

Back: Inco's reduction works at Copper Cliff, Ontario (George Hunter)

CONTENTS

CH. I INTRODUCTION
by R. J. W. Douglas, 1.

CH. II PHYSIOGRAPHIC SUBDIVISIONS OF CANADA
by H. S. Bostock, 9.

CH. III THE CANADIAN MINERAL INDUSTRY
by R. B. Toombs, 31.

CH. IV GEOLOGY OF THE CANADIAN SHIELD
by C. H. Stockwell, J. C. McGlynn, R. F. Emslie, B. V. Sanford, A. W. Norris, J. A. Donaldson, W. F. Fahrig, and K. L. Currie, 43.

CH. V ECONOMIC MINERALS OF THE CANADIAN SHIELD
by A. H. Lang, A. M. Goodwin, R. Mulligan, D. R. E. Whitmore, G. A. Gross, R. W. Boyle, A. G. Johnston, J. A. Chamberlain, and E. R. Rose, 151.

CH. VI GEOLOGY OF SOUTHEASTERN CANADA
by W. H. Poole, B. V. Sanford, H. Williams, and D. G. Kelley, 227.

CH. VII ECONOMIC MINERALS OF SOUTHEASTERN CANADA
by E. R. Rose, B. V. Sanford, and P. A. Hacquebard, 305.

CH. VIII GEOLOGY OF WESTERN CANADA
by R. J. W. Douglas, H. Gabrielse, J. O. Wheeler, D. F. Stott, and H. R. Belyea, 365.

CH. IX ECONOMIC MINERALS OF WESTERN CANADA
by H. W. Little, H. R. Belyea, D. F. Stott, B. A. Latour, and R. J. W. Douglas, 489.

CH. X GEOLOGY OF THE ARCTIC ARCHIPELAGO
by R. Thorsteinsson and E. T. Tozer, 547.

CH. XI BIOCHRONOLOGY: STANDARD OF PHANEROZOIC TIME
by E. W. Bamber, T. E. Bolton, M. J. Copeland, L. M. Cumming, H. Frebold, W. H. Fritz, J. A. Jeletzky, D. C. McGregor, D. J. McLaren, B. S. Norford, A. W. Norris, G. W. Sinclair, E. T. Tozer, and F. J. E. Wagner, 591.

CH. XII QUATERNARY GEOLOGY
by V. K. Prest, 675.

CH. XIII GROUNDWATER GEOLOGY
by I. C. Brown, 765.

PHOTO CREDITS, *792.*

GUIDE TO ILLUSTRATIONS, *793.*

INDEX, *801.*

I. Introduction

Development of Resources 2

Geological and Geographic Subdivisions 4

Scope of the Book 5

Acknowledgments 6

References 6

Appendix 7



1

Introduction

R.J.W. Douglas

Since the fourth edition of *Geology and Economic Minerals of Canada* was published in 1957 our knowledge of the geology of Canada has increased greatly. Heretofore, geological information was meagre for many parts of the country, although detailed but isolated information was available for the mineral producing areas and more accessible regions. A large part of the new knowledge has been gained in the course of systematic geological mapping; by 1957 nearly 75 per cent of Canada was still unmapped geologically whereas in 1968 only 15 per cent remained unmapped at reconnaissance scales. The important element of continuity has been established by the initial mapping, so that it is now possible to assess and present a broad range of data in both geological and tectonic terms of reference. Of the total geological research in Canada, however, only a part has been directed towards the regional type of investigation; continuing studies have been made of many matters of economic and scientific interest, not only by the Geological Survey of Canada but also by provincial organizations, the mining and petroleum industries, consultants, and the universities (*see* Appendix).

The increased knowledge has resulted from a larger number of geologists active in geological studies, general improvements in accessibility, greater use of helicopters and other aircraft, and expansion in the exploration and production programs of the petroleum and mining industries. Federal and provincial organizations, whose prime function is to publish, have acquired much of this additional knowledge, but also industry has been freer in permitting staff to publish data of scientific value and data no longer considered to have competitive economic value. Moreover, money is available to university professors to engage in geological investigations and they too are contributing even more than ever before to the literature. Grants are made by the Geological Survey of Canada on the advice and recommendation of the National Advisory Committee on Research in the Geological Sciences and also by the National Research Council. Whole new fields of endeavour and extensions to the old established fields of geology have been developed in relatively recent times, and more and more other scientists, particularly physicists and chemists, are bringing their own techniques and ideas to bear on geological problems. As a consequence, it is no longer practicable to summarize all the facts pertaining to the geology of Canada, or to attempt to portray them on a single map. Instead, we must be selective in preparing a text, and in presenting certain facets in visual form on several maps.

The objective of this edition is to provide both the geologist and non-geologist with a brief and handy reference to new and important facts, concepts, and interpretations of the geology and mineral resources of Canada. For the Canadian geologist, it is hoped that the book will serve as a ready source of information on regions not particularly familiar to him and stimulate the development of new ideas involving tectonics and metallogenesis. For geologists from other countries and for students the book is intended to serve as a reference, the starting point for an understanding of the geology of Canada and an introduction to its literature.

DEVELOPMENT OF RESOURCES

Canada's economic growth is dependent to a very large degree on the discovery of new mineral deposits and continual renewal of reserves. Mineral production represents more than 7 per cent of the Gross National Product, and more than 40 per cent of the total primary

production. It is the largest component of our export trade, accounting for 30 per cent of the total. The value of the total mineral production in 1968 was $4,735,000,000; of this amount metals represented 53 per cent, fuels 28 per cent, non-metallic minerals 10 per cent, and structural materials 9 per cent. About 60 per cent of the output was exported, the greater part to the United States. In the last decade gains have been made in almost all sectors of the industry. Prominent among these advances have been increases in the production of iron ore, potash, natural gas, and sulphur. The increase in iron ore production has resulted from the revolutionary advances made in beneficiation processes and steel-making technology, and from the construction of the St. Lawrence Seaway which permits ore from Quebec and Labrador to be shipped into the interior. The potash deposits deep beneath the plains of southern Saskatchewan have been reached by the novel engineering technique of freezing the bedrock aquifers that were penetrated during the sinking of the shafts. The crude oil and natural gas industries have experienced rapid growth, natural gas now ranking sixth in the mineral industry, and at the same time production of sulphur as a byproduct has increased to the extent that Canada is the world's second largest producer.

Exploration and development of the mineral resources of Canada are carried out principally by private industry, the exceptions being Eldorado Mining and Refining Company, Société québécoise d'exploitation minière, and Panarctic Oils Ltd. Federal and provincial government agencies do not engage directly in prospecting or exploration, although some valuable mineral deposits have been developed as a result of their investigations. Within the bounds of the provinces, mineral rights are vested in the provincial governments except for the lands granted some years ago when land titles included both surface and mineral rights. The provincial governments enact legislation to regulate exploration, development, and conservation of mineral and other natural resources. Federal departments formulate and administer regulations governing the mineral industry in Yukon and Northwest Territories,

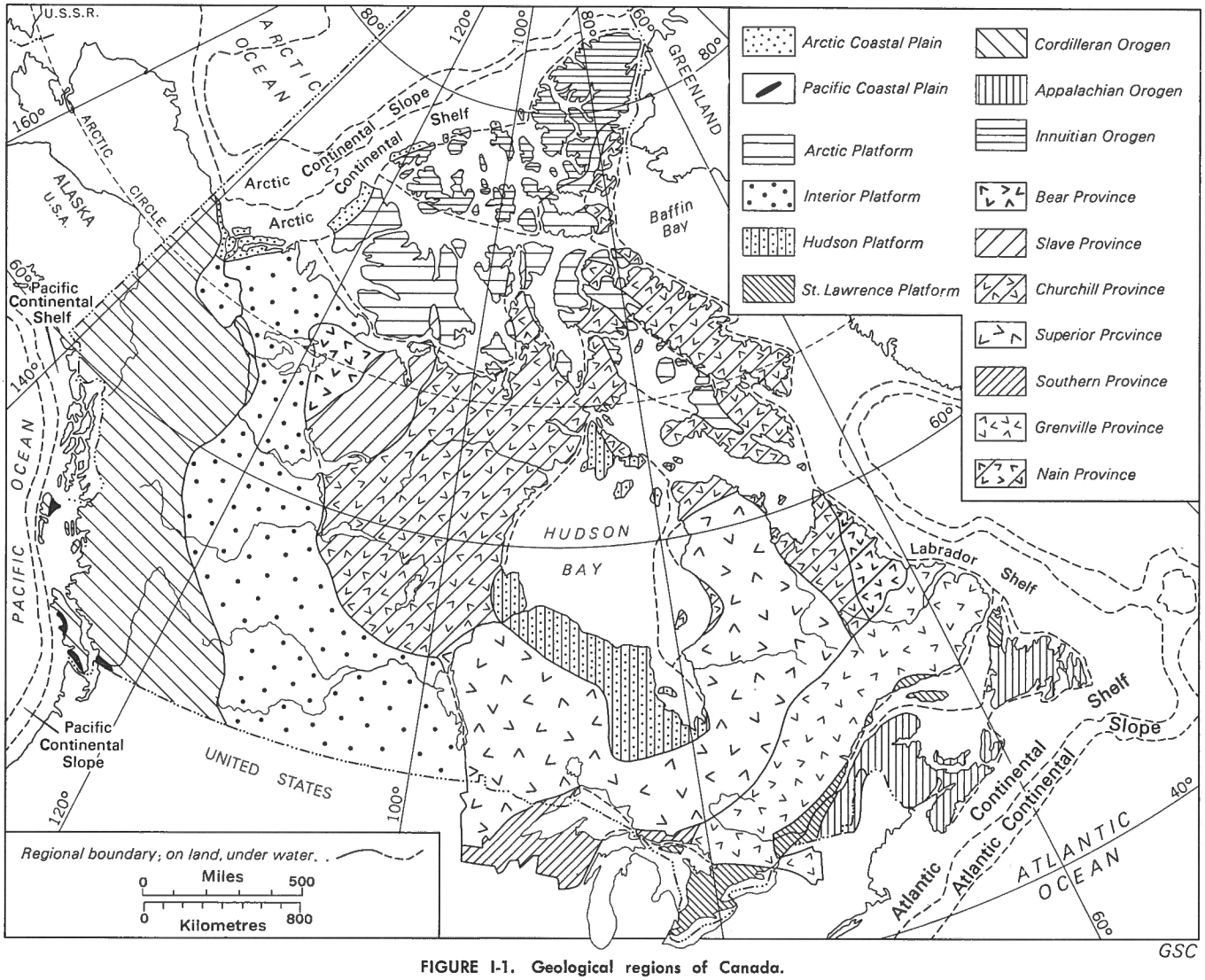

FIGURE I-1. Geological regions of Canada.

offshore, and in certain lands within the provinces such as Indian lands.

The Department of Energy, Mines and Resources is the principal federal government department dealing with the scientific and technical aspects of the nation's mineral, water, and energy resources (*see* Appendix). Three branches are concerned with geology — the Geological Survey of Canada, Marine Sciences, and the Earth Physics Branches; other branches are concerned with hydrogeology, mineral development, and other matters. The Earth Physics Branch contributes mainly in the fields of seismology and gravity; the Marine Sciences Branch investigates the sediments of the ocean floor; the Geological Survey of Canada conducts investigations designed to add to the understanding of the geology of Canada, to contribute to the advancement of knowledge in the earth sciences, and to provide data needed for the discovery and exploitation of the nation's mineral deposits.

GEOLOGICAL AND GEOGRAPHIC SUBDIVISIONS

Canada (Fig. I-1)[1] has a land area of 3,852,000 square miles of which 292,000 square miles is covered by freshwater lakes (Table I-1); about 2,964,000 square miles is mainland, and 596,000 square miles is islands, situated mainly in the Arctic. In common with other maritime nations, Canada exercises sovereign rights over adjacent regions under the sea for the purpose of exploring them and exploiting their natural resources. Within the confines of the extremities of the land areas of Canada is some 858,000 square miles of enclosed marine waters. The submarine area of the continental shelves bordering the Atlantic, Arctic, and Pacific coasts is about 523,000 square miles, and of the continental slopes is 563,000 square miles. The area, then, within the limits of the continental slope surrounding Canada totals about 5,526,000 square miles, or 2.9 per cent of the surface of the globe.

Geologically, the central part of Canada contains the Precambrian rocks of the Canadian Shield, the outcropping part of the North American craton. The several structural provinces comprising the Shield are composed of rocks laid down during the different Precambrian eras, and rocks, mainly crystalline, that were produced or affected by the orogenies near the close of the Archean eon and of the Proterozoic eras. These provinces are Superior, Slave, Bear, Churchill, Nain, Grenville, and Southern Provinces. The craton is bordered by three geosynclines composed mainly of Phanerozoic and late Precambrian rocks. They have been deformed at various times, and at present constitute the Appalachian, Innuitian, and Cordilleran Orogens. Between the Canadian Shield and the orogens

[1] All figures, tables, and plates are given in terms of consecutive arabic numbers preceded by a roman numeral which refers to the chapter. The exception is in Chapter XI — Biochronology, where the customary usage of plate and figures by paleontologists is used.

TABLE I-1 *Area of principal geological and geographic elements of Canada*

	Square Miles
Land areas	3,852,000
Land	3,560,000
Freshwater lakes	292,000
Mainland	2,964,000
Islands	596,000
Enclosed marine waters	858,000
Queen Elizabeth Islands	84,000
Southern arctic islands	217,000
Foxe Basin	83,000
Hudson Strait	64,000
Hudson and James Bays	320,000
Gulf of St. Lawrence	90,000
Continental Shelves	523,000
Atlantic	225,000
Labrador	125,000
Arctic	134,000
Pacific	39,000
Continental Slopes	563,000
Atlantic	124,000
Labrador	56,000
Baffin	219,000
Arctic	148,000
Pacific	16,000
Canadian Shield Provinces	2,146,000
Superior	586,000
Slave	87,000
Bear	47,000
Churchill	
western mainland	554,000
eastern mainland and islands[1]	449,000
Southern[2]	89,000
Nain	43,000
Grenville[2]	291,000
Orogens	1,046,000
Appalachian[1]	191,000
Innuitian[1]	230,000
Cordilleran[2]	625,000
Platforms	1,601,000
St. Lawrence[1]	110,000
Arctic[1]	408,000
Interior	701,000
Hudson[1]	382,000
Coastal Plains	27,500
Arctic	25,000
Pacific	2,500

[1] Includes areas under marine waters
[2] Includes contiguous parts in U.S.A.

are the St. Lawrence, Arctic, and Interior Platforms, which are the parts of the craton that subsided slightly in the past and are now covered by thin veneers of Phanerozoic sedimentary rocks. In the central part of the Canadian Shield, and underlying most of Hudson Bay, is the Hudson Platform. Bordering the oceans are the Atlantic, Arctic, and Pacific Continental Shelves which are composed mainly of little deformed Mesozoic and Cenozoic strata. The Arctic shelf has a broad, nearly co-extensive Coastal Plain; the Pacific Coastal Plain is small. The continental shelf and slope together constitute a depositional geological province, the outer margin of which marks the edge of the continental crust and the transition to oceanic crust.

SCOPE OF THE BOOK

Aspects of the geology and mineral deposits of Canada are treated either by major geological regions — the Canadian Shield, Southeastern Canada, Western Canada, and the Arctic Archipelago — or by individual subjects covering the whole of the country — physiography, mineral resources, biochronology, Pleistocene geology, and groundwater. The accompanying folio includes a new series of maps at 1:5 million scale showing the geology, tectonics, glacial features, physiography, mineral deposits, magnetic anomalies, and gravity. A geotectonic correlation chart is also presented for each of the four major geological regions, an innovation that reduces to visual form the sequence of rocks and tectonic events that ultimately produced the geology of each local sector as we know it today.

On the Geological Map of Canada (1250A) the four principal geological parameters — age, facies, structure, and degree of metamorphism — are used to typify and unify the geology of major sectors of the country and to show the fundamental geological features of the Precambrian geosynclinal belts of the Canadian Shield and the bordering Phanerozoic geosynclines in a consistent way (Douglas, 1969). Rocks are categorized by age according to two standards—depositional and orogenic. Although orogenic ages are, in large part, customarily considered to be determined by radiometric methods, the geological age designation of the orogeny is assigned only those rocks demonstrably igneous, intrusive, or metasomatized. Metamorphism does not change the geological age of sedimentary, volcanic, and basic igneous rocks unless they have been so reconstituted that their original nature can no longer be determined. The same principle is applied to granites and other plutonic rocks; they are considered to be metamorphosed or remetamorphosed unless it can be shown that they have been reintruded, remelted, or metasomatically replaced.

On the Tectonic Map of Canada (1251A) the geology is portrayed in relation to the age of the last major orogeny that the rocks have undergone. The rocks are categorized according to both their tectonic position within the geosynclines and craton and to the stages of geosynclinal evolution. In the course of preparation of the map considerable progress has been made in time classification of Precambrian rocks and in establishing structural provinces and subprovinces in the Canadian Shield (Stockwell, 1964). Much of this progress has been due to the steadily increasing numbers of isotopic age determinations available. Approximately 1,500 age determinations of rocks and minerals have been made in the laboratories of the Geological Survey of Canada; most are by K-Ar methods on biotite, muscovite, hornblende, and the whole rock. About the same number have been made on rocks from Canada at Canadian universities and at several universities and other organizations in the United States. Selected data, categorized according to method, material, geological environment, and laboratory, are shown on the Isotopic Age Map of Canada (1256A).

Mineral deposits, produced or being produced, are set forth on Map 1252A, Mineral Deposits of Canada. The metallic and non-metallic deposits are classified according to type and geological environment; only those with cumulative production of more than 36,000 tons are shown. The fuels are classified according to commodity; if designated a field, petroleum and natural gas deposits are shown regardless of size, and the age of their producing horizons is indicated. The bulk of the developed mineral deposits are in the southern part of the country close to the main population centres. This is partly a function of the geology, but in large part is related to the high costs of exploration, development, and marketing of mineral products from the less accessible regions of Canada where only those deposits of exceptional quality, grade, and size can now be produced economically (Lang and Douglas, 1959).

For the past 20 years aerosurveys of the total intensity of the earth's magnetic field have been carried out, in large part jointly by federal and provincial departments. The results are being published at a scale of one inch to one mile. These surveys are flown with a half-mile spacing of flight-lines and a mean terrain clearance of 1,000 feet. The sensitivities of the magnetometers used vary from 3 to 5 gammas. Some of these data, together with information from private companies, reduced to 1:5 million scale, are on the Magnetic Anomaly Map of Canada (1255A). To the end of 1967, aeromagnetic maps for about 1,250,000 square miles, one third of Canada, have been published. The large-scale maps, mainly of the Canadian Shield, are used by industry in exploration for mineral deposits and by the geologists of the Survey during their geological investigations or in the subsequent interpretation of their data. On the small-scale compilation many of the details are suppressed, but aspects of the larger features of the surface geology and structures deep in the crust are evident.

The intensity of the earth's gravity field has been systematically determined for a large part of Canada by the Gravity Division of the Earth Physics Branch, and the anomalies are shown on the Bouger Gravity Anomaly Map of Canada (1257A). Spacing varies from a grid of one determination per 6 to 7 miles in remote regions to more closely spaced determinations in the more accessible areas

or where the values are anomalous. Most of the larger gravity features are now known but not all have been related to geological features.

In some chapters certain subjects are treated on a Canada-wide basis. In Chapter II, Physiographic Subdivisions of Canada, the distinctive features of each major division are described, and for the first time formal physiographic subdivisions of the Canadian Shield are defined. The limits of the units are outlined on Map 1254A. This map is also useful for geographic reference as it has the maximum number of place names and physical features. The importance of mineral production to the economy is described in Chapter III, The Canadian Mineral Industry, together with data on Canada's position relative to other mineral producing countries of the world, and some other economic comparisons. The composition of the typical and most useful faunas used in correlating and establishing the age of the Phanerozoic rocks of Canada are given in Chapter XI, Biochronology: The Standard of Phanerozoic Time. Some of these accounts also include comments on the paleogeographic significance of the faunal realms and evolutionary trends. In recent years much knowledge has been obtained on the Pleistocene of Canada not only of the relationships and extent of the northernmost continental ice sheets and glacier complexes, but also, particularly in the more southerly regions, of more precise stratigraphy, correlation of events by radiocarbon dating, and the details of the history of deglaciation from the Wisconsin maximum to the present-day glaciers and ice fields. These data are presented in Chapter XII, Quaternary Geology, and portrayed on Glacial Map of Canada (1253A). A brief account of the hydrogeological regions of Canada is given in Chapter XIII, Groundwater, together with descriptions of some of the types of current investigations of the groundwater resources of Canada, whose appraisal, on a national scale, has just recently commenced.

Chapters IV to X deal with the geology and economic geology of the four major regions — the Canadian Shield, Southeastern Canada, Western Canada, and the Arctic Archipelago. In the geological chapters, the approach is basically tectonic, an ordering of the data temporally and regionally. For the Canadian Shield, the Precambrian rocks are described according to the main structural provinces and their tectonic evolution by era. For the parts of Canada composed mainly of Phanerozoic rocks, the geology of the adjacent platformal and geosynclinal regions are brought together. To this are added the as yet rather meagre data on the geology of the bordering continental shelves and coastal plains. The tectonic history for each period is summarized in general terms that can be appraised without familiarity with the rock units themselves or recourse to stratigraphic nomenclature. They are interpretive and speculative to some extent. The main body of the texts of the four geological chapters is descriptive and essentially factual, but of necessity selective. Structures and facies are shown in their present geographic positions as regional palinspastic bases cannot yet be prepared. On the geotectonic correlation chart for each of the four major geological regions of Canada all correlations are in time, three degrees of certainty being indicated. The time standard to which the correlation is made is not indicated, however. The sectors are of unequal size. They vary in geographic extent to embrace an area appropriate to the particular sequence of rock units and events. They are generalizations, as it is not practicable to show all possible geological relationships for each sector.

In the economic geology chapters — treating the Canadian Shield, Southeastern Canada, and Western Canada — the mineral deposits are grouped and described in sections on each commodity whether metal, mineral, or fuel. Generalizations are expressed in the introductory statements to the degree that seems warranted without engaging in undue speculation, and serve also to place the deposits in their natural setting or geological habitat. Descriptions themselves are intended to be factual and are limited to geological, mineralogical, and genetic features and to the factors controlling formation of the deposits. Read in conjunction with the chapters on geology it is hoped that other ideas concerning mineral associations will be suggested, particularly to exploration geologists.

ACKNOWLEDGMENTS

Acknowledgment of data used is not generally in accordance with the custom followed in the preparation of scientific papers because it is not practical to cite all sources of factual data and concepts or to list all available publications. The authors are indebted to all those who have contributed to knowledge of the geology of Canada whether or not their work is listed in the references. We are especially indebted to colleagues who have generously contributed unpublished data and interpretations, participated in discussions, and engaged in critical reading of the manuscripts. The partial reference system is intended primarily to facilitate the interested reader's transition from this book to the geological literature where he can get additional information and references, and only secondarily to extend credit to individual contributors. Personal communications and data from unpublished theses are acknowledged simply by use of the individual's name. Bibliographic lists of all reports and maps by provincial and federal organizations are available on request (*see* Appendix), and are also available from some of the publishing geological societies.

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APPENDIX

SOURCES OF GEOLOGICAL INFORMATION

Federal Departments

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Ottawa 4, Ontario.

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Calgary 44, Alberta.

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Mines Branch
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Inland Waters Branch
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Policy and Planning Branch
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Northern Administration Branch
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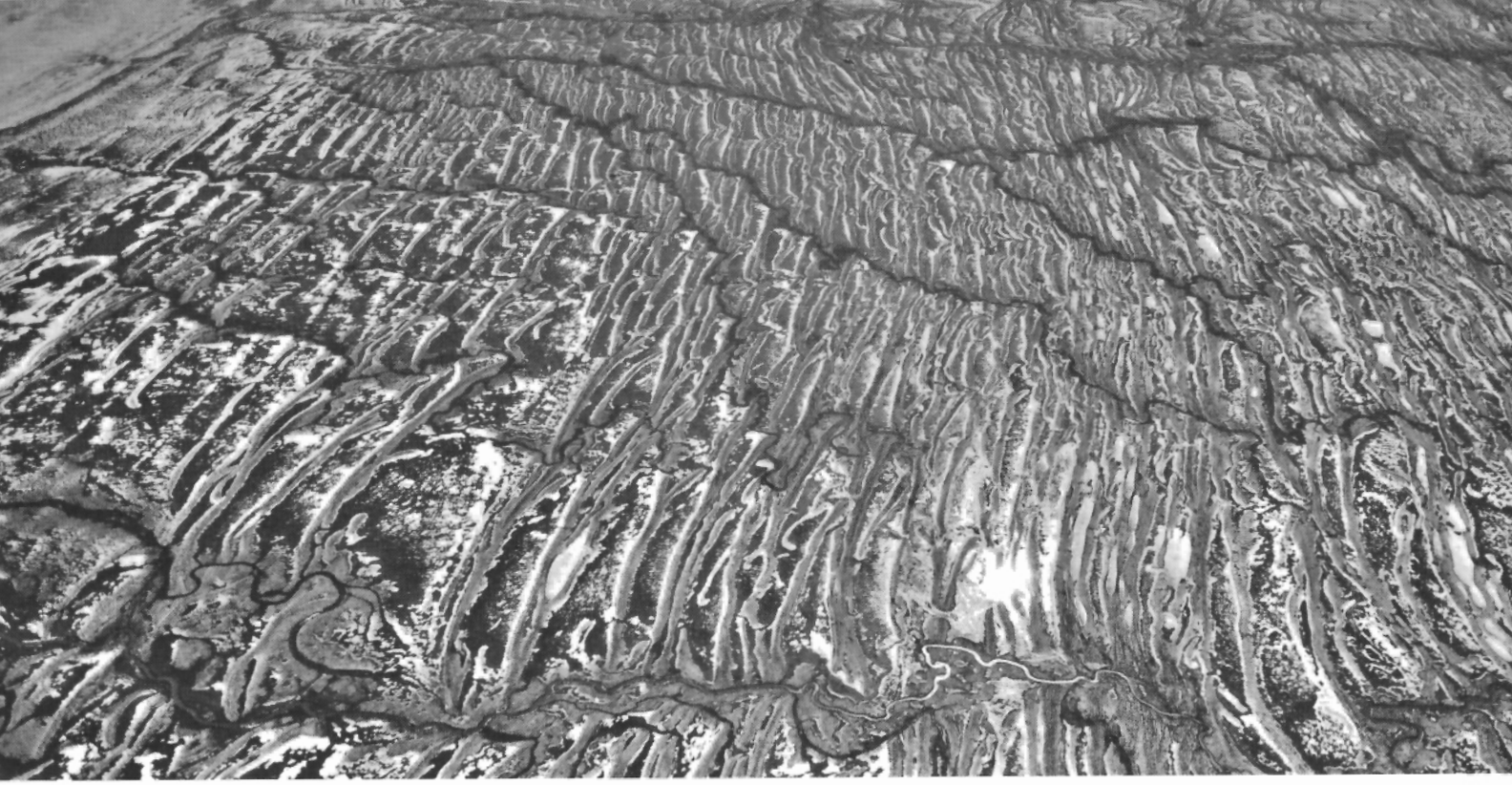
II. Physiographic Subdivisions of Canada

Introduction 10

The Shield 10

The Borderlands 17

Selected References 30



INTRODUCTION

Physiographically and geologically Canada is composed of two great parts: a core of old, massive, Precambrian crystalline rocks forming the Shield, and a surrounding crescent of younger, mainly stratified, rocks forming the Borderlands. In simple terms, the surface of the Precambrian core resembles that of an inverted dinner plate with its flat, slightly depressed centre and an outward shelving rim terminated by a steep edge. The younger rocks of the Borderlands surround the northern, western, and southern parts of the plate as segments of two concentric rings. The inner ring comprises a chain of lowlands, plains, and plateaux of generally flat lying sedimentary rocks overlying the gently shelving rim; the outer ring is formed of discontinuous areas of mountains and plateaux in which the younger rocks are deformed. On the northeast rim of the plate the Borderlands are absent; there the edge of the Shield is tilted upward and drops abruptly into the sea.

Part of the boundary between Canada and adjoining territories of the United States is natural, marked by lakes, rivers, or divides; much is, however, drawn at parallels or meridians. Canada has immensely long and irregular coastlines that embrace great bays, inlets, and channels within the general confines of her headlands and islands. Parts of these coastal waters are shallow and cover the continental shelves along the Atlantic, Pacific, and Arctic coasts.

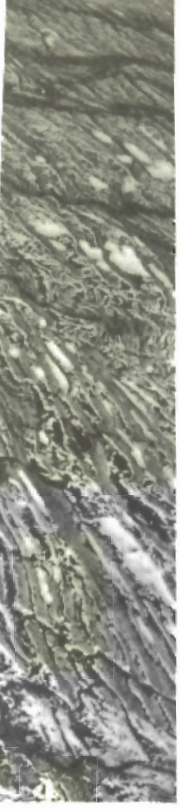
Physiographically the Shield and Borderlands are divided into regions or provinces each comprising many smaller units. In the Borderlands physiographic boundaries are to a great extent indicated by distinctive changes in the topography and geology. For the Shield no such single controlling factor is evident, although the ages of orogeny form a basis for geological subdivision (Stockwell, 1965). Despite this, there are features that enable a physiographic division of the Shield into major and minor units to be made.

THE SHIELD

General Features

Viewed from some prominent summit the landscape presents an even, monotonous skyline interrupted by rounded or flat-topped monadnocks and ranges of hills. The smooth horizon is evidence of an old, almost all-subduing erosion surface; this is the most outstanding characteristic of the Shield and gives a similarity to the terrain whether it is seen in Labrador or the Northwest Territories. Only in scattered areas such as the Davis and Labrador Highlands, around Wager Bay, and Mealy Mountains is the relief mountainous, but even in these places remnants of an old erosion surface form the summits. Over the greater part of the Shield local relief is 200 to 300 feet. The greatest relief of more than 5,000 feet is found in the Davis and Labrador Highlands.

Many long periods of erosion have had a great influence in levelling the broad expanses of the Shield and reducing them to a nearly featureless surface even though parts are geologically distinct. Such features as



Raised beaches on Hudson Bay Lowland.

II

Physiographic Subdivisions of Canada

H.S. Bostock

the East Arm Hills are no higher than the crystalline uplands on either side of them. The eastern edge of Kaniapiskau Plateau is a marked scarp that overlooks the sinuous ridges of the Labrador Hills, although some of their summits stand as high as the plateau surface.

Drainage

The surface of the Shield varies in detail, but as a result of glaciation a great proportion of it is taken up by water as lakes, ponds, and swamps. Although the percentage of area covered by water has not been accurately measured, it is known to exceed 25 per cent in many large regions; in some parts, such as the Thelon Plain or areas of Pleistocene lakes and marine submergence, the percentage is relatively small.

The drainage pattern varies widely and as yet no preglacial system has been deciphered even for the main streams. With a few exceptions such as Leaf River, the main streams flow in the direction of general slope of the land surface. The details of the tributary drainage, however, are disrupted by the local effects of glaciation including the tilting of the land surface, the grouping of the bedrock, and the deposition of surficial materials. Stream channels commonly follow elements of bedrock structure such as persistent fracture zones, major joint patterns, or the traces of soft strata; in places they are diverted across former divides dammed by glacial debris.

In many areas glacial movement transverse to bedrock topography and scattered deposition of glacial drift have formed a mosaic of ponds interspersed with small ridges and hummocks amongst which streams flow along tortuous courses. In contrast to this heterogeneous drainage, where drift deposits are scarce the streams show a trellis drainage pattern, developed to various degrees, in which the main lines follow long straight lineaments, major fracture zones, or the dominant joint systems in the massive rocks. Differences in the resistance to erosion of the strata or gneisses may also impress a drainage pattern guided by the sinuous twists of the folds with S-shaped lakes or concentric or radiating valleys. In other localities depositional features of glaciation may have been the overruling influence, imposing a strong parallelism upon the drainage in drumlin fields, the swells and troughs of drumlinoid features, or the irregularly parallel, congruent system of ridges and hollows of ribbed moraines. Indeed, in no other region of the world has glaciation more clearly placed its stamp upon the drainage pattern.

A comparison of the drainage patterns on each side of the boundary between the Shield and the Borderlands commonly reveals decided differences despite the over-all masking effect of glaciation. On the Shield the drainage generally shows a multitude of lakes and streams, their pattern controlled by the Precambrian bedrock; whereas on the adjacent Borderlands lakes are less numerous and often display bulbous outlines, and the Paleozoic bedrock imposes different stream patterns.

Old Peneplains

The general surface of the Precambrian rocks of the Shield dips at low angles under the bordering Paleozoic strata of the Arctic and Interior Plains, the southwest part of the St. Lawrence region, and particularly around

Hudson Bay Lowland and Foxe Plain. In some areas drilling has shown that the surface is undulating but in general it shelves gradually to great depth beneath the rocks of the Borderlands. Where the character of the Precambrian surface is well exhibited along the border its hills and valleys continue beneath the overlying Paleozoic strata. The strata fill hollows on the Precambrian surface, encroach on the ridges, partly mantle them, and leave only the summits projecting as inliers along the continuations of the ridges. The present surface of the Shield, at least in many areas near its borders, appears to be an exhumed pre-Paleozoic erosion surface.

Outliers of Paleozoic strata are widely scattered in the Shield (some lying far within its borders) generally in depressions or grabens. Some are in valleys such as that of Lake Timiskaming (Hume, 1925) where the Paleozoic strata were deposited on a relief of more than 200 feet, or in deep basins such as that of Manicouagan Lake (Rose, 1955) where several outliers are preserved at 1,200 feet or more below the surrounding Precambrian upland surface without evidence of tectonic movement to bring them to their present position. From the distribution of the Paleozoic outliers it appears that the Shield was peneplaned and the valley system cut in the bedrock before the Paleozoic. It seems that the Precambrian surface has been little changed since the removal of the Paleozoic cover, and that glaciation has only superficially modified its character (Ambrose, 1964).

Although the surface of the Shield seen today is mainly an exhumed surface once buried under Paleozoic strata many areas of Proterozoic sedimentary rocks show similar relationships to the surface of the crystalline Archean rocks beneath them (Ambrose, 1964). The surface of the Shield has had an ancient and complex history indeed. First there was the development and partial dissection of a peneplain on the Archean rocks in pre-Proterozoic time, then the burial and exhumation of this surface once or perhaps twice during the

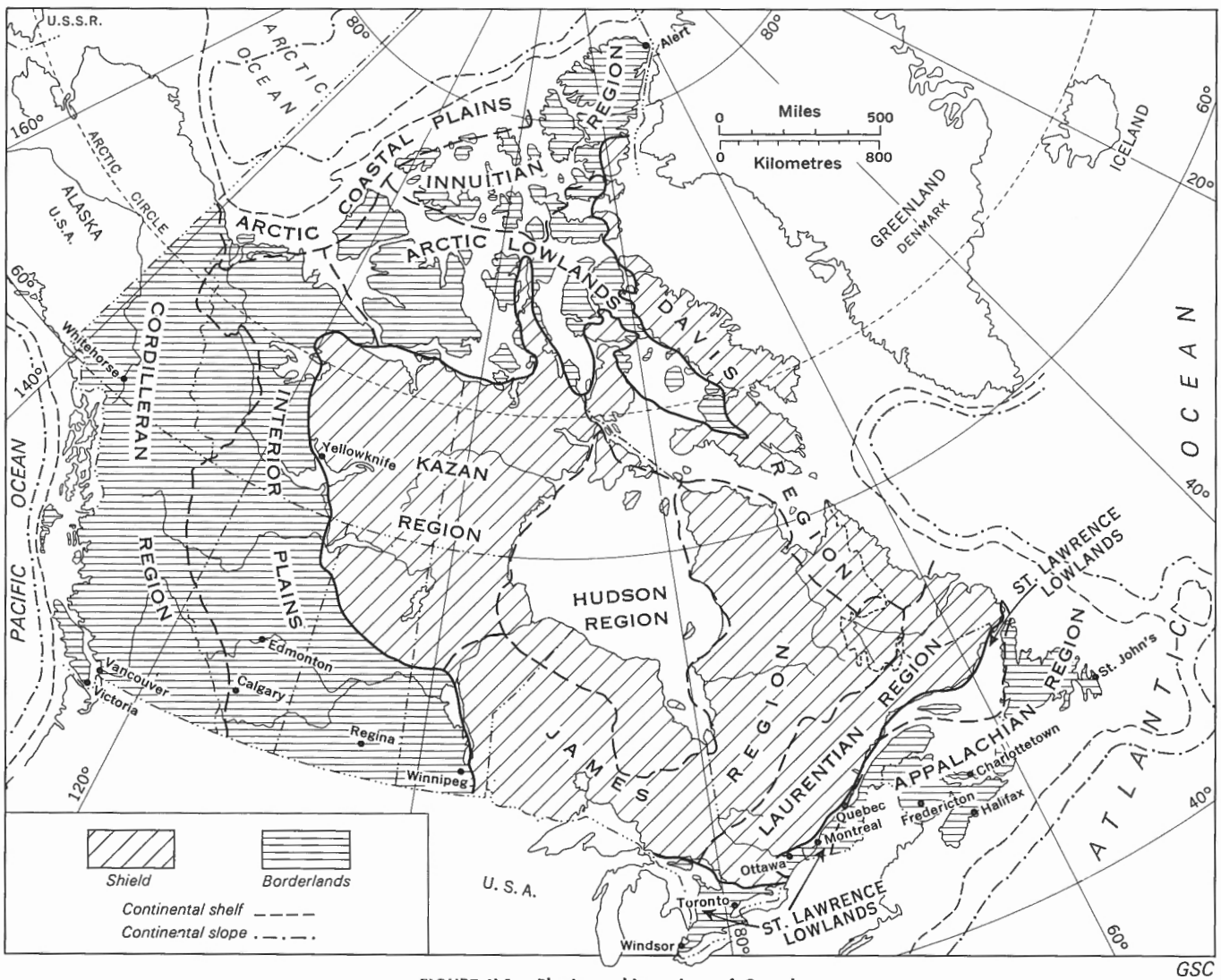


FIGURE II-1. Physiographic regions of Canada.

Proterozoic, followed by the exhuming of these several surfaces from beneath the cover of Paleozoic strata, and finally the scouring of the surface during the surges of Pleistocene glaciation.

Subdivisions of the Shield

Despite the general uniformity of its terrain, geological structure and evidence of orogenies show that the Shield is composed of different parts that were once probably as distinct physiographically as the Cordilleran Region and Interior Plains are today. The ancient boundaries are still geologically evident and in places are physiographically distinct. Accordingly, some geological boundaries transect terrain that forms a single physiographic unit.

Four types of terrain in the Shield are relatively easy to distinguish as physiographic units and coincide with particular geological characteristics: (1) plains, formed on areas of unmetamorphosed rocks more or less flat lying, such as the sandstone and conglomerate terrains of the Thelon, Athabasca, and Cobalt Plains or the terrain of flat-lying gabbro sills of Nipigon Plain; (2) hills, composed of low-grade metamorphic rocks, generally tilted or gently folded sediments or sills, such as the Bathurst, East Arm, Penokean, and Labrador Hills; (3) mountains, composed of particularly resistant massive rocks, such as the anorthosites of the Mealy Mountains which stand well above the surrounding terrain; (4) highlands, formed of broad uplifted areas of deeply incised massive crystalline rocks, such as the Labrador Highlands or the glacier-capped Davis Highlands. Other units are distinguished by less obvious characteristics, and in some places the boundaries are arbitrarily drawn to divide uniform but unwieldy units.

Kazan Region

This region consists of great areas of massive rocks that form broad sloping uplands, plateaux, and lowlands — a rather expressionless terrane. Included in its northwestern part are hills formed of down-faulted and folded stratified rocks, and in its southern and central parts there are some sandy plains of flat-lying strata.

The largest unit of the region is *Kazan Upland,* which covers a wide expanse of rolling lake-spattered country, typical in general aspect of so much of the Shield. North of the Athabasca Plain the highest points are 1,500 to 1,900 feet in elevation with relief rarely more than 200 or 300 feet. The upland slopes gently and drains to Hudson Bay except along its western border where it drains into the Mackenzie River system. Towards Hudson Bay, but about 150 miles from the coast, the relief increases to 600 feet and the higher summits become separated and conspicuous. As much as 100 miles from the sea the upland was covered by the postglacial marine overlap that reached as high as 675 feet above present sea level (Lee, 1959). The area of overlap forms a coastal plain of little relief mantled largely by reworked drift deposits that obscure nearly all the bedrock.

Athabasca Plain forms a large embayment into the western part of Kazan Upland. The plain consists of nearly flat lying sandstones and has a rolling hummocky, lake-spotted, forested surface that rises from 900 feet elevation on the west to 2,000 feet on the east and has a local relief of 200 to 300 feet.

Thelon Plain lies in the central part of the region and includes areas of nearly flat lying sandstones and volcanic rocks that are characteristically expressed on the surface by sandy flats sparsely covered by vegetation.

Southwest of Thelon Plain, *East Arm Hills* are formed of down-faulted and folded, differentially eroded sediments and gabbro sills. On the north side the resistant sills dip southerly forming broad cuestas as much as 900 feet above Great Slave Lake, the surface of which is about 500 feet in elevation. Most intervening valleys are flooded by arms of Great Slave and other lakes. On the south side the hills are narrow and lower, ending abruptly at the long, straight, high, fault scarp bordering Kazan Upland.

In *Back Lowland,* east of Bathurst Hills near Thelon Plain, some upland areas are more than 1,000 feet in elevation. The lowland surface, however, is generally lower than that of Kazan Upland and Wager Plateau which border it on the south and east, particularly between Back and Ellice Rivers. Much is typical of the treeless Shield, but near the coast the surface is mantled by mud and silt of the postglacial marine overlap. Eskers are a typical feature of the inland areas.

The surface of *Wager Plateau* rises gradually from Chesterfield Inlet and Wager Bay to 2,000 feet elevation where it is deeply dissected. From Wager Bay northward the surface declines and loses its identity in low patches alternating with rugged areas of 400 to 700 feet elevation.

Boothia Plateau is underlain by crystalline gneiss forming a narrow north-trending prong of the Shield partly covered by outliers of Paleozoic strata. In the south its surface merges with Wager Plateau 2,500 feet above sea level; northward the surface declines as that of the adjacent lowlands rise.

Bathurst Hills (Bird and Bird, 1961) are composed of down-faulted, folded sediments and sills that lie within and extend south from Bathurst Inlet between higher upland areas of massive rocks. The softer strata have been eroded and in many places lie submerged beneath bays and channels, leaving the harder members, generally the sills, as long cuestas more than 1,000 feet in elevation. Marine silts and reworked deposits from the marine overlap mantle some of the lower parts. Southwestern Bathurst Hills form a prong of ridges termed the Peacock Hills. Some are rugged and reach more than 2,000 feet elevation standing as much as 600 feet above nearby lakes.

Bear–Slave Upland consists mainly of massive rocks though some stratified rocks lie in the northeastern part. Generally the peneplain surface is typical of the treeless,

bare-rock parts of the Shield. Numerous lakes fill the hollows, and rounded rocky hills, exhibiting a relief of a few hundred feet, are characteristic. Many monadnocks exceed 1,000 feet in relief, their summits reaching above 1,600 feet elevation.

The main part of the *Coronation Hills* area is low, extending along Rae and Richardson Rivers eastward into Coronation Gulf. The hills are formed of gently northward dipping sediments intruded by sills and dykes, the latter showing as islands in the Gulf. The hills and ridges rise more than 800 feet above sea level. In the southwest the ridges become more closely spaced until they merge to form the broad, relatively smooth topped uplands of the Coppermine Hills with altitudes of 2,000 feet. Southwest of Dismal Lakes massive rocks protrude through the sediments to form large rounded hills.

Davis Region

Davis Region (Dunbar and Greenaway, 1956) extends from Ellesmere Island south and southeastward to northern Labrador. It is relatively narrow at each end but widens in the middle where crystalline Precambrian rocks surround flat-lying Paleozoic strata of Foxe Plain, a part of Hudson Region. The general aspect is one of a broad, gently warped, old erosion surface, shallowly etched by erosion along joint systems and zones of weakness. Surficial deposits are few except locally near Foxe Plain. Along the eastern coast relief is generally high. The region is divided into a northern or insular part and a southern or mainland part.

Davis Highlands (Ives and Andrews, 1963) are an elevated belt of deeply dissected crystalline rocks (Pl. II-1) that extend along the northeastern flank of Baffin and Bylot Islands and thence north over eastern Devon Island and Ellesmere Island as far as Bache Peninsula. Along this distance of some 1,200 miles the highlands are mountainous with remnants of an old erosion surface forming the summits or marked by a general accordance of peaks 5,000 feet or more above sea level. The old surface slopes gently westward in the north and southwestward in the south. Long arms of the sea penetrate the land as sounds or fiords, and on Baffin Island some cut through the highland into Baffin Upland. The highest parts are commonly crowned by ice caps; the largest is Penny Ice Cap along the border with Baffin Upland, at an elevation of 6,000 to 7,000 feet.

Baffin Coastal Lowland borders the Davis Highlands on eastern Baffin Island, extending from Henry Kater Peninsula to Eglinton Fiord. Its greatest width is about 25 miles, but generally it occurs as isolated, narrow strips at the ends of the peninsulas or islands.

Baffin Upland is also dominated by the high upland surface present on Davis Highlands. The old erosion surface slopes generally southwestward from 3,000 feet elevation near the highland and Barnes Ice Cap to near sea level around Foxe Plain. The border between Davis



PLATE II-1. Fiords and glaciers, Davis Highlands, Baffin Island.

Highlands and Baffin Upland is drawn where the dissection is shallower, the old erosion surface becomes dominant, and the drainage begins to flow southwestward.

Hall Upland on Hall Peninsula is separated from Baffin Upland by Cumberland Sound; the surfaces of these two uplands are similar. Hall Upland reaches 3,800 feet elevation on the northeast side and slopes southwestward into Frobisher Bay. South of the bay *Frobisher Upland* rises abruptly from the sea to 3,000 feet, then slopes southward into Hudson Strait. A similar but lower and more irregular terrain extends westward to Foxe Peninsula. The southwest part, with maximum elevations of about 1,200 feet, is included in the Frobisher Upland. The massive rocks that underlie all these uplands apparently continue beneath the sea and project as isolated areas of the Frobisher Upland in the rugged plateaux of Salisbury and Nottingham Islands.

The mainland part of *Melville Plateau* is largely a featureless smooth upland 1,500 to 2,000 feet above sea level, but with some rugged areas along its west border. On Southampton Island the upland surface is lower at the northwestern and eastern ends, but near the centre much reaches 2,000 feet elevation.

Labrador Highlands and *George Plateau* together resemble the highlands and uplands of Baffin Island. They exhibit parts of an old erosion surface that is cut on massive rocks and slopes inland. The highlands are deeply incised by valleys and fiords along the sea coast; they include several ranges such as Torngat Mountains, the highest and most rugged area of the eastern mainland of Canada. These ranges commonly reach 3,500 feet above sea level and occasionally 5,000 feet. The old erosion surface occurs as undulating, generally westerly sloping remnants around which glaciation has sculptured cirques, rugged peaks, and deep U-shaped valleys and fiords where the cliffs drop precipitously into the sea. In the north some deep valleys traverse the highlands and connect with others in George Plateau, whose streams drain into Ungava Bay. Inland, the old surface becomes more continuous and somewhat less undulating. It forms the main features of George Plateau sloping towards Ungava Bay and Whale Lowland. The highest parts of the plateau are generally between 1,000 and 2,000 feet elevation, but are somewhat higher near Labrador Highlands. Much is covered by areas of drumlinized drift, particularly the lower parts.

Whale Lowland is a broad, irregular, drift-covered area drained by Whale River. Scattered hilltops range from 2,000 feet elevation in the south to 800 feet in the north. Its boundaries, determined largely by the limits of drift cover, are arbitrary.

Hudson Region

Hudson Region includes Hudson Bay Lowland, Hudson Bay itself, Southampton Plain, the Belcher Islands, and Richmond Hills. It is composed of the partly submerged blanket of unmetamorphosed, mainly flat lying Paleozoic and Proterozoic strata that form the main central depression on the surface of the Shield.

Hudson Bay Lowland is a low, swampy, marshy plain with subdued glacial features and a belt of raised beaches that border the bay *(Frontispiece).* Paleozoic strata mantle the massive or stratified Precambrian rocks, and slope gently northeastward and eastward into Hudson and James Bays. Elevations of 200 to 400 feet are reached along the edge of the Severn and Abitibi Uplands; part of its south border is marked by an escarpment overlooking Abitibi Upland. A most conspicuous feature is Sutton Ridge, an inlier of Precambrian strata that rises to 600 feet elevation, about 500 feet above the surrounding surface.

Southampton Plain (Dunbar and Greenaway, 1956), formed on flat Paleozoic strata, is rather irregular but well drained. It is generally less than 300 feet in elevation and is bordered by two higher outliers of Precambrian rocks of Melville Plateau: that on Southampton Island reaching 500 feet and the other at the north end of Coats Island, 700 feet.

Belcher Islands, ranges of hills projecting through the waters of Hudson Bay, are formed from the resistant strata of a thick section of Proterozoic sedimentary and volcanic rocks elaborately folded into long curved hairpin-shaped structures. The higher summits reach about 400 feet above the sea and are truncated by an old erosion surface. *Richmond Hills,* including the Nastapoka Islands and other islands to the north, are mainly a group of east-facing cuestas of Proterozoic sedimentary and volcanic rocks.

Within the main area of *Hudson Bay* the average depth of water is 100 metres, but in the north-central region it reaches 230 metres. The general bathymetric configuration of Hudson Bay is saucer-shaped, particularly in the peripheral areas. Long ridges and valleys disrupt the pattern, however, and in some places, the more continuous, sinuous trends resemble a submerged river system. Echogram soundings clearly show steep-walled valleys as much as 30 metres high that extend beyond present estuaries. The floor of one valley, which is 80 miles west of the Ottawa Islands and trends northerly, lies 330 metres below sea level, almost 200 metres below the level of the adjacent sea floor. It appears to owe its configuration to subaerial erosion along a structural trend.

Other linear trends comprise topographic highs, the most prominent of which is the northerly trending ridge parallel to and 75 miles west of the deep trough mentioned above. The highest part of this arcuate ridge, the Central Shoals, lies at a depth of about 30 metres and may be bedrock controlled. Other smaller ridges are apparent and are thought to be submerged interfluvial areas. In northeastern Hudson Bay adjacent to Digges Islands, a steep-walled, linear trough about 490 metres deep is

thought to have resulted from subaerial erosion and glaciation along a fault.

James Region

The greater part of James Region exhibits the characteristic features of the Shield that are apparent in the major uplands and plateaux of Kazan Region but are less apparent in Davis Region. It is divided into a number of units mainly on arbitrary grounds, but also includes small distinct units such as Nipigon Plain and the Labrador Hills. The physiography of northern Quebec and Labrador has been described by Hare (1959) from a study of airphotographs. His map has been used extensively in preparing this report, but his boundaries have been simplified and some names added.

Severn and *Abitibi Uplands* are similar; both are underlain by crystalline Archean rocks and both have broad rolling surfaces that rise gently from Hudson Bay Lowland in the north reaching about 1,500 feet in their higher parts near the southwest and southern borders. Most of the uplands lie between 900 and 1,200 feet elevation. Large areas in southwest Severn Upland and northern Abitibi Upland are mantled by glacial lake deposits.

Nipigon Plain is formed on nearly flat lying Proterozoic gabbro sills and sediments surrounding Lake Nipigon. It lies at an elevation of 900 feet and is deeply dissected so that remnants of the resistant sills stand as scattered hills. *Port Arthur Hills* are also formed of Proterozoic sills and sediments tilted to the south forming cuestas and ridges some of which extend as promontories into Lake Superior. They reach summits of 1,500 feet and in places rise 800 feet above the lake and adjacent valleys. Michipicoten Island is formed of stratified rocks and forms a link with the *Penokean Hills,* which are composed of folded sediments. Most summits of the latter are 800 to 1,000 feet in elevation, but two points reach at least 1,500 feet, giving a maximum relief of about 900 feet. *Cobalt Plain* is composed of flat-lying clastic sediments with ridges and hills formed by gabbro sills or inliers of Archean crystalline rocks.

The surface of *Eastmain Lowland* joins those of Abitibi Upland and Larch Plateau. It is generally lower than either of these units, however, and large areas are covered by the Pleistocene marine overlap, particularly near the border of James Bay. Lakes are few and muskeg prevails. As a whole the plain has a gently westward sloping surface, broken here and there by hills. In the highest part northwest of Lake Mistassini, lakes are 1,200 feet above sea level.

Larch Plateau has an undulating surface with elevations of 500 to 1,500 feet (Stevenson, 1965). Bedrock is widely exposed and drift is thin over most of its surface. Areas on the west side were covered by the postglacial marine overlap, however, and much of the low-lying southeast part is cloaked with drumlinized till. In the south drainage is westward to Hudson Bay, but farther north a large area is drained to Ungava Bay by Leaf and Larch Rivers. Leaf River starts within 30 miles of Hudson Bay at an elevation of about 500 feet; as it flows northeastward the hills in the heart of the plateau rise from 600 to more than 1,000 feet.

Povungnituk Hills include the Cape Smith and Povungnituk Ranges. Their rocks are largely Proterozoic volcanic rocks and gabbro sills similar to those of Richmond and Labrador Hills. The folded strata form a series of east–west ridges and valleys with a relatively high relief in the western part but of more subdued character in the east where the summit surface of the hills virtually merges with the Larch and Sugluk Plateaux.

Sugluk Plateau at the northern tip of Ungava Peninsula is separated from Larch Plateau by Povungnituk Hills. In places along the north coast the plateau drops precipitously as much as 1,700 feet into the sea. To the west it slopes gently and was covered by the postglacial marine overlap. Inland its slightly undulating surface reaches elevations of 1,900 feet.

The general aspect of *Lake Plateau* is that of a rolling plain with numerous lakes (Hare, 1959); isolated hills of bedrock (some of rugged character) stand about 500 feet above the general surface. In the northwest the elevations are between 1,500 and 3,000 feet. The southern part is generally higher with less relief and ranges from 2,500 to 3,000 feet, whereas to the northeast the elevations are generally lower, between 1,400 and 2,100 feet.

Kaniapiskau Plateau forms the core of Lake Plateau and is composed of rugged hills of massive rocks in which fracture systems are conspicuously etched. Parts reach elevations of more than 3,000 feet. Its eastern border forms an escarpment overlooking Labrador Hills but elsewhere its surface merges with that of Lake Plateau.

The *Mistassini Hills* consist of subparallel hills, ridges, and valleys; the valleys are partly filled with the waters of Lake Mistassini. Summits are more than 3,500 feet in elevation in the Otish Mountains, and stand 2,500 feet above Lake Mistassini. Most ridges are north-facing cuestas formed of sedimentary rocks and gabbro sills and generally have a subdued aspect, particularly around the lake.

Labrador Hills cover the main part of the belt of folded Precambrian sedimentary and volcanic rocks along the east side of Larch and Kaniapiskau Plateaux. It is a belt of sinuous ridges and valleys formed by downwarped folded and faulted strata. The summit elevations range from 2,400 feet in the south and central parts to 1,200 feet in the northern part. In the northern end the

relief is subdued but towards Koksoak River it is bolder and as much as 300 feet. Farther northward again the belt of hills becomes narrow and merges with the adjacent plateau areas. There the valleys are generally floored with drift, tills, drumlins, and sand.

Laurentian Region

This region comprises uplands and highlands that rise abruptly above the St. Lawrence Lowlands along its southeast border. The strong relief at the border is partly caused by fault scarps, in contrast to the western and northern margins of the Shield where the peneplain surfaces dip gently under the overlapping Paleozoic strata of the Interior Plains and Arctic Lowlands.

Laurentian Highlands rise abruptly above the St. Lawrence Lowlands, or the water of the St. Lawrence River and Gulf. The southeastern edge is deeply, and in places widely, incised by several large rivers that gather volume in the high undulating interior and drop rapidly into the river or the gulf. As a result, the southeastern border has a mountainous appearance, the relief commonly at 1,000 to 2,000 feet. In many parts of the interior the surface is also mountainous, but the evenness of summit levels is commonly apparent. This old erosion surface has a relief of 1,000 to 1,500 feet, somewhat higher than that of the adjacent Abitibi Upland to the north. Much of the upland surface lies at about 2,000 feet elevation but here and there broad areas have summits at 3,000 feet, and in a few parts at 4,000 feet. Laurentian Highlands extend south of Ottawa River and include the spur of hills formed of Precambrian rocks that cross St. Lawrence River to form the Thousand Islands and join with Adirondack Mountains of the United States.

Mecatina Plateau is a large area of varied terrain — a rough undulating region with deeply dissected margins. Its surface rises rapidly from the coast to elevations of 700 to 1,200 feet and thence slopes gently to about 1,800 feet along its inland borders. Much of the western and inland parts is drift covered.

Northwest of Mecatina Plateau and bordering on James and Davis Regions is an area of broken terrain composed of upland and plateau units with intervening lower areas. *Hamilton Upland* is a composite unit of three bold, rugged areas of hills, each as much as 2,500 feet in elevation, that stand 500 to 1,000 feet above the surrounding lake and drift-covered plateau surface. Southeast of the upland, drift-covered areas form *Hamilton Plateau*. *Melville Plain,* which contains Lake Melville lying practically at sea level, is an irregular lowland, mainly about 500 feet in elevation; a few hills reach 1,000 feet. It is much dissected by river valleys entering from the west. *Mealy Mountains* rise abruptly above the south side of the Melville Plain and reach their highest elevation of 3,700 feet in the east. An old erosion surface truncates their summits and slopes downward to the west. Cirques are reported in the highest part (Hare, 1959).

THE BORDERLANDS

The Borderlands, which contain a great variety of physiographic provinces, are divided into easily recognized subdivisions, many with long established names. They include the Innuitian Region, Arctic Coastal Plain, and Arctic Lowlands in the north; the Interior Plains and Cordilleran Region of western Canada; and the St. Lawrence Lowlands and Appalachian Region in the east.

Innuitian Region

This is a region of varied topography — on the average more rugged than that of the surrounding provinces — developed from the thick assemblages of deformed sedimentary rocks and minor intrusions. The region is roughly triangular, covers an area of about 210,000 square miles, and forms part of the outer ring of the Borderlands between the Shield and the Arctic Ocean. According to E. F. Roots it is characterized by two curving belts of relatively rugged terrain separated by extensive and discontinuous tracts of more subdued topography. The landscape shows the effects of several interrupted or superimposed physiographic cycles, but most of the landforms are in an intermediate or comparatively youthful stage of development and reflect directly the rock composition, geological structure, and recent vertical movements of the Innuitian system.

The northernmost mountain belt comprises the *Grantland* and *Axel Heiberg Mountains,* which include the British Empire and United States Ranges of northwest Ellesmere Island and the mountainous central and western Axel Heiberg Island. They consist mainly of long ridges of folded Mesozoic and Paleozoic strata, with minor igneous intrusions. The local relief is as much as 4,000 feet, and the highest summits reach 8,200 feet above sea level. In central Axel Heiberg Island and northwestern Ellesmere Island the mountains are nearly buried by ice sheets through which the peaks project as rows of nunataks. On their northwestern flank the mountains pass rather abruptly into a narrow seaward-sloping plateau; on the east and southeast sides they pass with decreasing ruggedness into the elevated dissected edge of the Eureka Upland. At the south end of Axel Heiberg Island, however, the ridged terrain is apparently continuous with that of the inner mountain belt of Eureka Upland. The ranges and ridges are transected by numerous parallel-sided, steep-walled cross valleys, some straight and angular, and some smoothly curving. In northwest Ellesmere Island these make impressive fiords and trough-like valleys with glaciers at the head.

The inner belt of rugged topography extends the length of the Innuitian Region from northeast Ellesmere to the west end of Parry Islands, and is developed throughout on subparallel, gently curving folds in Paleozoic carbonates, shales, and sandstones of the Franklinian Geosyncline. The topography is on the whole less rugged than

that of the outer mountain belt; it is most mountainous in northeast Ellesmere Island where *Victoria and Albert Mountains* reach elevations above 6,500 feet and carry an extensive ice sheet and numerous glaciers. Well-developed valley and ridge topography, with ridge tops 3,500 to 4,000 feet in altitude and a local relief up to 2,500 feet, trends south to southwest across Ellesmere Island and then swings abruptly westward continuing, with interruptions, to Melville Island. The tops of the ridges become progressively broader, and the flat or undulating summit surfaces become the dominant feature of the landscape as the ridged upland passes into the little dissected Parry Plateau.

Eureka Upland comprises an elongated area of relatively subdued topography in central and western Ellesmere Island and eastern Axel Heiberg Island, and lies between the two major mountain belts of the Innuitian Region. In large part it has a rolling and ridged surface controlled by underlying folded strata, but with altitudes generally less than 3,000 feet. There are extensive areas of low dissected plateaux and gently rolling low uplands developed on soft late Mesozoic and Tertiary sandstone and shale. The uplands are cut by trench-like depressions, curving, sinuous, or branching in plan, forming great dendritic drainage systems with floors that extend gradually beneath the sea to depths of more than 300 metres in Nansen Sound and Norwegian Bay.

The terrain of *Parry Plateau* is strongly ridged on Bathurst Island. The ridges are broad, flat topped and straight sided, average less than 800 feet above sea level, and separate wide flat-floored longitudinal valleys transected by rugged, ravine-like cross valleys. Farther west on Melville Island where the plateau surface dominates the landscape, the average elevation increases to more than 1,000 feet. A few hills at 2,500 feet elevation in the western part of the island carry four small ice caps, and cliff-walled fiord-like bays and straits cut deeply into the uplifted plateau surfaces.

In the western Queen Elizabeth Islands, Parry Plateau is bounded on the north by *Sverdrup Lowland,* a region of low relief developed on a structural basin of generally soft, poorly consolidated, and little deformed Mesozoic rocks. For the most part the land surface is a rolling, scarped lowland less than 500 feet above sea level. Locally areas of low upland and dissected plateau reach 1,200 feet elevation and elsewhere extensive areas of plain lie only a few feet above sea level. Among the distinctive landforms are dissected domes and ring structures developed on gypsum anhydrite piercement diapirs and igneous intrusions. About half the total area of Sverdrup Lowland is covered by water; the sea floor is nearly flat and lies at depths of 200 to 500 metres over large areas, with a relatively abrupt rise to the island shores. This suggests an extensive period of erosion controlled by a wave-base level below that of the present seas.

Arctic Coastal Plain and Continental Shelf

Arctic Coastal Plain includes the coastal terrain along the shores of the Arctic Ocean from Meighen Island to Alaska. The plain borders and lies seaward of the Innuitian Region, Arctic Lowlands, Interior Plains, and Cordilleran Region. Although there is a general similarity of landscape throughout the province, each section has distinctive physiographic characteristics.

The *Island Coastal Plain,* the section extending from Meighen Island to Banks Island, is underlain by the Beaufort Formation of unconsolidated Tertiary or early Pleistocene sands and gravels. On Meighen Island this formation has been uplifted 600 feet and eroded to form a hilly terrain that carries an ice cap whose summit is more than 800 feet above sea level. On Ellef Ringnes, Borden, Brock, and Prince Patrick Islands the coastal plain is low, remarkably flat, and uniform, with a network of consequent streams (incised at most a few feet) draining seaward. The inland boundary of the coastal plain, as much as 100 feet above sea level, is marked in places by a low ridge or scarp facing seaward or inland. There are scattered remnants of the coastal plain surface in Sverdrup Lowland. On Banks Island the coastal plain is characterized by low rolling hills and well-organized drainage, with alluvial plains and low westward-sloping terraces bordering the master stream valleys that bear witness to several stages of erosion and uplift. Inland at an altitude of about 300 feet the plain merges with the Victoria Plain.

On the mainland the Arctic Coastal Plain is represented by Mackenzie Delta and Yukon Coastal Plain. *Mackenzie Delta* has a compound character and includes not only the delta of the present river but remnants of earlier deltas and complex fluviatile-marine features such as Cape Bathurst (Mackay, 1958). The delta plain is remarkable for its multitude of lakes and channels, and the older parts for the scores of pingoes, some of large size, which form the most outstanding features of the landscape.

Immediately west of the mouth of Mackenzie River, *Yukon Coastal Plain,* which lies considerably higher than the Mackenzie Delta, appears to be largely an erosion surface cut in bedrock mantled with a thin veneer of recent sediments. Towards the west, the plain decreases in elevation and is drift or lake covered along the coast to near Herschel Island. West of the island to the Alaskan border, the plain is formed of coalesced deltas and alluvial fans built by streams flowing from the British Mountains and of coastal lagoons.

Arctic Coastal Plain extends beneath sea level to merge with *Arctic Continental Shelf.* Offshore from Meighen, Ellef Ringnes, Borden, and Prince Patrick Islands the sea floor is shallow and slopes both seaward and towards the channels which separate the islands where the slope is steeper, more irregular, and reaches depths of about 300 metres. It is evident that the exposed part, or Arctic Coastal Plain, is but a part of an emerged sur-

face of deposition which, in the main, antedates the formation of the deep channels between the islands. Below about 300 metres the drowned headlands that extend seaward from each island group merge with the continental shelf proper. The continental shelf is a nearly plane surface (sloping seaward at about 12 feet per mile) whose inner part is smoothly continuous with the flat channel floors at a depth of 450 metres and whose outer edge, 80 to 120 miles from the headlands of the outermost islands, breaks sharply and cleanly at a depth of 750 metres to fall away as a relatively steep continental slope which descends smoothly to the Arctic Ocean basin. North of the mainland the continental shelf is shallower, and the general slope of Yukon Coastal Plain continues unchanged beneath sea level, reaching a depth of only 100 metres at a distance of 50 or more miles from shore. There the shelf breaks away rapidly to an irregular, relatively dissected continental slope that drops steeply to oceanic depths beneath Beaufort Sea.

Arctic Lowlands

The Arctic Lowlands are formed on the flat-lying or nearly flat Paleozoic and late Proterozoic sedimentary rocks, lying between the Shield and the Innuitian Region. They include Lancaster Plateau, Foxe Plain, Boothia Plain, Victoria Lowland, and Shaler Mountains.

In the northeast the surface of *Lancaster Plateau* slopes gently southward from about 2,500 feet on southern Ellesmere Island, across central Devon Island to average elevations of 1,000 to 2,000 feet on Somerset Island and Brodeur Peninsula of northwestern Baffin Island. Its general uniformity is continuous without apparent deflection across the wide channels of Jones and Lancaster Sounds and Prince Regent Inlet. Farther south the surface descends still lower until on both sides of the Gulf of Boothia it forms the surface of *Boothia Plain* reaching sea level and apparently dips beneath it at Committee Bay.

Foxe Plain is low, smooth, and underlain by Paleozoic strata. It forms a very shallow basin-like area on the old surface of the Precambrian rocks which is partly covered by very shallow seas. On Baffin Island the plain is divisible into three parts, the Great Plain of the Koukdjuak, Soper Highland north of Koukdjuak River, and Putnam Highland to the south. The last reaches 600 feet elevation. Conspicuous features of the Great Plain of the Koukdjuak are a broad belt of emerged beaches stretching north-south on the easterly part and numerous, circular lakes on the central part of the plain.

West of the narrow finger of crystalline rocks of the Shield forming Boothia Plateau, the east side of the *Victoria Lowland* appears to have the same tilted surface as that forming Lancaster Plateau and Boothia Plain, the general elevations decreasing to the southwest and south across Prince of Wales Island, the southwest coast of which is very low. Little is known of the water depths in M'Clintock Channel, but it may be that the channel floor is a part of the lowland surface below present sea level, for to the west the surface rises again to form the low east coast of Victoria Island ultimately reaching 2,500 feet elevation in the central part of Shaler Mountains in the interior of the island. *Shaler Mountains* are composed of later Proterozoic stratified rocks intruded by gabbro sills which form cuestas and are capped by flat-lying volcanic rocks. Farther west, the elevation of the general surface of Victoria Lowland decreases gently, and in central Banks Island it merges with Arctic Coastal Plain. On eastern Banks Island and throughout much of Victoria and Prince of Wales Islands, the smooth undulating ancient erosion surface of the lowland is covered by a variety of glacial deposits, with extensive areas of drumlinoid ridges that impart a characteristic grain to the minor topography. Belts of very thick, rough moraines form irregular hills on southwestern Victoria Island and on both sides of Prince of Wales Strait comprising the most rugged topography within the Victoria Lowland.

Interior Plains

Interior Plains are underlain by flat-lying later Proterozoic, Paleozoic, Mesozoic, and Tertiary strata, occupying the region between the Shield on the east and the mountains of the Cordilleran Region on the west. They join with the St. Lawrence Lowlands of eastern Canada through the United States of America and are separated from the Arctic Lowlands by Amundsen Gulf. The southern part is semi-arid grass land; the central part is tree-covered; the northern part is tundra. The region is divided into several units, those in the north being smaller and more varied than those in the south.

Horton and *Anderson Plains* form the Arctic slope where drainage is directly to the Arctic Ocean. Anderson Plain is covered by a sheet of glacial drift and outwash, the main feature which distinguishes it from the somewhat higher Horton Plain where bedrock is generally exposed. Horton Plain is underlain by nearly flat lying Paleozoic and late Proterozoic sediments and its surface reaches elevations of 1,200 to 2,000 feet, the higher parts being in the south. Much of the western part of the plain is rocky, but in the eastern part rolling areas of drift occur. With few exceptions the lakes are small and scattered. Throughout, as the streams gather size, they become entrenched 200 to 400 feet below the surface. In the north some late Proterozoic strata are slightly folded and faulted giving rise to a rolling surface of low scarps and scattered mesas. Melville Hills, which include Mount Hooker, 1,600 feet in elevation, form the divide between Horton River and the sea. Anderson Plain (Douglas, *et al.*, 1963; Mackay, 1958) is characteristically undulating and rises inland. Large parts of its higher levels are rocky, but extensive areas of outwash from an ice front that lay near its southeast border cover much of the surface and several entrenched run-off channels wind across the plain. On the north along the coast it includes the low, flat, lake-dotted

Capes Bathurst and Parry but inland it rises abruptly along an escarpment 600 to 1,000 feet in elevation. As elevations gradually increase southward the main stream valleys and lake-floored run-off channels lie at elevations of 500 to 700 feet whereas the surrounding surface is about 800 to 1,000 feet.

Peel Plain (Douglas, *et al.*, 1963) lies southwest of Mackenzie River. The southwest part is a broad, shallow hollow about 400 feet in elevation, some parts with innumerable small lakes and others without any. On the northeast the surface rises to Grandview Hills which reach 1,500 feet elevation at their eastern end. Arctic Red River is entrenched across Peel Plain, and Ontaratue River rising on its surface also follows a deeply incised course to Mackenzie River. The southwestern part of the plain is underlain by Mesozoic sediments whereas Paleozoic strata form the hills in the north and east.

Peel Plateau rises in steps between Peel Plain and Mackenzie Mountains. The surface of the first step truncates the upturned edges of the Paleozoic and Mesozoic strata and is, therefore, an erosion surface. Above, a second step or terrace rises in parts of the plateau and on top of this a group of undulating, rounded plateaux of nearly horizontal strata form a third step. These plateaux are separated from Mackenzie Mountains by a broad, shallow but poorly defined valley. In the southwest some areas of the plateau appear unglaciated but most of its surface is covered by thin drift deposits with small scattered lakes lying in the hollows. Remnants of one or more large, entrenched, meltwater channels wind across the plateau from the east to Snake and Peel Rivers.

Colville Hills (Douglas, *et al.*, 1963) embrace several ridges of Paleozoic strata that stand above the general level of the surrounding plains. The hills and ridges enclose hollows with several large lakes in a net-like pattern whose meshes are 10 miles or more across. The lower ground within the meshes lies at 800 to 1,000 feet elevation, whereas the rather sinuous ridges have summits up to 2,200 feet.

Great Bear Plain (Douglas, *et al.*, 1963) is largely composed of Mesozoic strata. Its rolling surface is generally below 1,000 feet elevation, but a number of small subcircular plateaux and hills such as the Scented Grass and Grizzly Bear Hills, rise above this, reaching elevations of about 1,500 feet.

The south-facing escarpment of Cartridge Mountains forms the southern boundary of Great Bear Plain and overlooks *Great Slave Plain* whose surface of Paleozoic strata has generally little relief. Low scarps of resistant carbonate strata and small shallow lakes are characteristic of much of the surface below 1,000 feet elevation. Its central part, however, contains Horn Plateau underlain by Cretaceous strata, an outlier of the Alberta Plateau to the south.

Alberta Plateau (Holland, 1964) is composed of Cretaceous sediments and consists of a ring of plateaux separated by wide valleys. The plateaux include Cameron Hills and Caribou Mountains. On the north these hills form a high disconnected escarpment with summits between 2,500 and 3,200 feet elevation overlooking Great Slave Plain. On the east they overlook the north end of Saskatchewan Plain. On the southeast they embrace Cheecham Hills, 2,500 feet in elevation, and from there the south border of the plateau extends along the flanks of hills on the north side of the Athabasca River to join the Foothills of the Rocky Mountains. A smooth upland surface is the main feature of all the hills. They are generally higher to the southwest at about 4,300 feet elevation. The two main valleys that occupy more than 50 per cent of the area are termed the *Fort Nelson* and *Peace River Lowlands.* The rivers and their main tributaries are more or less entrenched into the valleys which rise from less than 1,000 feet in the north and northeast to about 2,500 feet elevation in the west.

South of Athabasca River, *Alberta Plain* (Acton, *et al.*, 1960) also largely composed of Mesozoic sediments but with some of Tertiary age in the west and south stretches southeastward to the International Boundary. Although virtually the continuation of the Alberta Plateau, it has a more even surface, with a few widely separated groups of low hills such as the Neutral, Cypress, and Porcupine Hills. Much of the plain is about 2,500 feet in elevation, river valleys entrenched 200 to 400 feet, and hills rising to 3,500 feet or more. Cypress Hills reach 4,700 feet and parts of their summit are believed to have escaped glaciation.

The eastern edge of the Alberta Plain (Pl. II-2) referred to as the Missouri Coteau constitutes a step down to *Saskatchewan Plain.* The Coteau, well marked at the International Boundary by a line of low, rounded hills partly formed by Tertiary sediments, gradually becomes indefinite northwestward. East of the Coteau, Saskatchewan Plain, formed on Mesozoic sediments and lower and smoother than Alberta Plain, is gently undulating and rolling, its surface ranges from 1,500 to 2,600 feet above sea level and relief is about 300 feet in the hillier parts. Saskatchewan River is now entrenched in bedrock across the plain. By the end of the Pleistocene its waters had incised a broad valley along a southeasterly course now only occupied by a comparatively small stream, the Qu'Appelle River. In the northwest the Ile-à-la-Crosse Lowland (Acton, *et al.*, 1960) around Peter Pond Lake forms a detached area generally below the surface of southern Saskatchewan Plain but is here included with it.

On the east, Saskatchewan Plain is bordered by the Manitoba Escarpment overlooking *Manitoba Plain* which is underlain by Paleozoic sediments. Streams flowing eastward have cut deep valleys in the escarpment dividing it into a line of separate hills, the highest of which, Riding Mountain, stands nearly 1,000 feet above Manitoba Plain. The surface of the Manitoba Plain has an elevation of about 800 feet and is very gently undulating to flat. It is largely covered by lakes and includes most of Lake Winnipeg. In the southern part its features have been



PLATE II-2

Alberta Plain and Missouri Coteau, with glacial moraine, near Assiniboia, Saskatchewan.

smoothed over by the deposition of the clays and silts of glacial Lake Agassiz. The beaches of this lake, marking successively lower water levels, wind along the lower part of Manitoba Escarpment.

Cordilleran Region

General Features

The Cordillera in Canada is divided longitudinally into three great belts, the Eastern System, the Interior System, and the Western System (Bostock, 1946 and 1961; Holland, 1964). Each has its own characteristic geology and likewise its physiography. The Eastern System is composed almost entirely of folded sedimentary strata. The Interior System is made of folded sedimentary and volcanic strata, massive metamorphic rocks, all intruded by large or small bodies of igneous rocks, with local areas of flat-lying volcanic rocks. In the Western System the largest unit, the Coast Mountains, is composed mainly of plutonic rocks whereas other parts contain a mixture of folded volcanic and sedimentary strata intruded by scattered bodies of massive igneous rocks. The three great systems are further divided into areas and subdivided into mountains, ranges, plateaux, hills, valleys, trenches, basins, plains, etc.

The Cordillera is also divided transversely into a number of segments by east–west belts of relatively low terrain. The broad northern part of the Interior Plateau, the most southerly and also the main feature, separates the two main segments of high and mountainous terrain in British Columbia. At the sixtieth parallel an area of low terrain including Liard Plateau, Liard Plain, and Yukon Plateau separates the Rocky and Cassiar Mountains from the Mackenzie, Selwyn, and Ogilvie Mountains to the north. The latter group, in turn, is separated from the Richardson Mountains by Porcupine and Peel Plateaux. The British Mountains in northern Yukon Territory are separated from the Richardson Mountains by the broad saddle that connects Porcupine Plateau and Yukon Coastal Plain.

Old Erosion Surfaces

Probably the most notable feature of the physiography of the Cordillera is that remnants of an old erosion surface



PLATE II-3

Unglaciated part of Yukon Plateau and Yukon River near Dawson, Yukon Territory.

are nearly always evident. This surface may not be the same in all parts, and indeed many features suggest that there are surfaces of different ages. An old surface is most conspicuous in the plateau areas of the Interior System. It has been referred to as a peneplain but its relief is not sufficiently low over large areas to qualify as such. Generally the accordant summits give an impressively even skyline, but the surface is undulating with some higher rounded summits. Locally, particularly in northern Yukon Territory, upland ridges are flat topped, and their structures bevelled to a remarkable degree. On the west flank of Coast Mountains the surface is impressive where it rises from Hecate Strait, truncating the island hills, bridging the fiords, and bevelling the mountain summits. On the higher peaks of the inland ranges it has been destroyed by erosion.

The old surface is thought to have been formed largely in the beginning of the Tertiary or possibly in the late Cretaceous, and to have gone through a complicated subsequent history of uplift, warping, subsidence, and renewed uplift that is just beginning to be understood in a few parts of the Interior System (Lay, 1940 and 1941). In the early Tertiary the old surface of erosion was uplifted, the rivers entrenched, and sediments deposited in the valleys and hollows. Early Tertiary volcanic rocks, chiefly lavas, then filled the valleys. They were later down faulted, warped, and bevelled by erosion accompanying renewed uplift and dissection. Lava has been extruded here and there up to the present.

Unglaciated Areas

Large unglaciated areas cover most of northwestern Yukon (Map 1253A) and give an indication of the character of the plateaux of the Cordillera before Pleistocene glaciation. There the topography is sculptured by the normal processes of subaerial erosion at northern latitudes. The classic area is that around Klondike River in Yukon Plateau where relief is 2,000 to 3,000 feet, and the metamorphic rocks are much the same as in the adjoining glaciated areas. The unglaciated plateau area is composed of long ridges spreading from a central divide, separated by closely spaced small creeks that merge into larger streams,

the valleys gradually widening. The gradients are even and uninterrupted by rapids or lakes. The valleys are V-shaped with sides flaring upwards and diminishing in steepness until the broad, rounded ridges of the upland are reached. These may be broken in places by castle-like outcrops on or near their summits. The general drainage is dendritic among smaller streams, but a few larger streams are straight, following some linear structural feature. Persistent bedrock terraces along many of the larger valleys are mantled by variable thicknesses of gravel, sand, and soil, and follow more gentle gradients than the present streams. In places stream reversal and capture are apparent. Lakes are generally lacking; a few ponds occur due to ice pushing on the valley flats, to damming of the rivers by lava flows, and to landslides; oxbow lakes occur along the larger rivers.

Glacial Features

Glacial ice covered the Cordillera with the exception of the area referred to above and some minor areas. In many ranges only the highest peaks projected above the ice, which left its imprint on the topography as both erosional and depositional features. Conspicuous in the mountains are cirques, arêtes, and horns, and along the valley sides ice-margin channels have been cut in bedrock, kame terraces, and moraines. On the valley floors are a host of features mainly formed of surficial materials: pitted till, outwash plains and terraces, drumlin fields, simple and compound eskers, lake terraces, and all manner of fluvial-glacial features.

Great U-shaped valleys furrow the country, most impressive where filled with lakes or inlets of the sea. A belt of long lakes stretches along the western or dry side of the Interior System from Kluane Lake in Yukon Territory to Taseko Lakes in British Columbia. Another zone lies on the eastern side, on the relatively wet western slopes of the mountains from Mayo Lake in the north to Shuswap and Kootenay Lakes in the south. It is, however, in the fiords of Coast Mountains that these glacially scoured valleys are most spectacular. Cut in a terrain formed almost wholly of hard crystalline rocks the fiords follow a pattern mainly controlled by fracture and joint systems on a grand scale. Many are several tens of miles long and in cross-section are deeply U-shaped as much as 2 miles wide, with a relief of 6,000 feet in places. Their floors may sometimes lie at more than 600 metres below sea level, rising seaward to a sill of small rock islands.

Trenches

Four great linear valleys, each at least 300 miles long, trend northwesterly through the Cordilleran Region, some of which delineate major physiographic and geological provinces. These are the Southern and Northern Rocky Mountain Trenches of British Columbia and the Tintina and Shakwak Trenches in the Yukon, which are impressive features where narrow, straight and steep-walled. The floors vary in width from less than a mile to more than 15 miles and are drained by several rivers whose low divides lie in the trenches. The floors are remarkably uniform in elevation rising gradually from about 1,800 to 3,000 feet except for Tintina Trench, much of which is near 1,000 feet elevation. A number of similar, smaller, and shorter trench-like valleys occur in the Interior System, such as the sinuous, northerly trending Purcell Trench of southern British Columbia and Teslin Valley in Yukon Territory.

Volcanic Features

In the Interior and Western Systems constructional landforms have been produced by volcanic processes. The most abundant are the plains, plateaux, and mesas formed from extrusion of basic lava, the Plateau basalts. Most striking, however, are three great shield volcanoes in the northwestern part of Interior Plateau and a group of similar volcanoes, such as at Edziza Peak, in Stikine Plateau. Also in the Stikine Plateau are a number of tuyas — flat-topped volcanoes which erupted in the midst of the Pleistocene ice (Mathews, 1947). A large number of small volcanoes and cinder cones occur singly, in line, or in groups. These range in age from pre-Pleistocene to within a few hundred years of the present, and vary in height from a few tens to a few hundreds of feet.

Eastern System

Porcupine Mountain Area

In the light of recent studies in physiography and geology in northern Yukon Territory a re-grouping of the physiographic units is made here. The Porcupine Plateau and its subdivisions formerly in the Northern Plateau and Mountain Area are termed the Porcupine Mountain Area and placed in the northern part of the Eastern System whose units they resemble. The ranges heretofore called the Northern Ogilvie Mountains are also similar and included in the Eastern System. The British Mountains and Old Crow Range are retained in the Interior System constituting the Brooks Range Area.

Richardson Mountains comprise a group of ranges having rugged peaks up to 5,500 feet elevation in its northern core. The narrow, lower southern ranges characteristically exhibit smooth, rounded profiles. These mountains do not appear to have been glaciated, but a tongue of Laurentide ice extended through them from the east via McDougall Pass.

Porcupine Plateau is almost entirely unglaciated. The northernmost part, Old Crow Plain, has an elevation of about 900 feet and is underlain by unconsolidated sills and peat. Closely spaced lakes and ponds have an oriented, rectangular pattern apparently produced by the prevailing winds (Plate II-4). Old Crow River and its tributaries meander elaborately and are entrenched in youthful valleys into the plain. In Eagle Plain, in the southern part of the plateau, long, even-topped ridges have broad, gently



PLATE II-4

Old Crow Plain, northern Cordilleran Region, Yukon. View northward, Old Crow River and rectangular lakes.

rounded summits typical of unglaciated areas. Relief reaches a maximum of about 900 feet in the highest part at the south where some ridges are 2,500 feet in elevation. Around the perimeter of the plateau the surface rises, notably in the west where bordered by the *Porcupine Ranges*. These ranges are underlain by folded stratified rocks that form linear ridges, hills, and low mountains, separated by broad irregular valleys. In Keele and Nahoni Ranges the summits exhibit remnants of an old erosion surface.

Mackenzie Mountain Area

Taiga Ranges and *Wernecke Mountains* comprise the westernmost part of Mackenzie Mountain Area. Taiga Ranges are separated from the Porcupine Plateau and Ranges by the broad, irregular valleys of Peel and Ogilvie Rivers. They were only slightly glaciated and their highest peaks and ridges reach 7,000 feet elevation. Wernecke Mountains are formed of phyllite and massive, nearly horizontal carbonate carved by glaciation into very rugged massive mountains. They are divided into several ranges by broad northwesterly trending valleys.

The interior part of *Mackenzie Mountains,* the Backbone Ranges, is a sea of peaks and ridges reaching to 8,500 feet. Some contain small alpine glaciers, and have been sculptured by alpine and valley glaciers. The more easterly ranges, the Canyon Ranges, are lower, divided by wide valleys or cut by deep canyons. During the Pleistocene ice from the east penetrated the valleys but otherwise the mountains show only local alpine and valley glaciation. Mackenzie Mountains, in the precipitation shadow of the Selwyn Mountains to the west, are relatively dry. The timberline is low, and the mountains present an aspect of bare, rock-covered slopes with a few cliffs.

Mackenzie Plain is a broad, rolling, drift- and tree-covered plain lying between Mackenzie and Franklin Mountains into which Mackenzie River is entrenched for part of its course. *Franklin Mountains* form a series of linear relatively low ranges and ridges, largely of resistant carbonates, that reach elevations of about 5,000 feet. They were over-ridden by ice moving from the east.

Liard Plateau south of Mackenzie and Franklin Mountains lies between South Nahanni and Liard Rivers. It is a region of tree- and tundra-covered hills, mostly less than 4,500 feet in elevation, and underlain mainly by shale and sandstone. Many summits are flat because of the attitude of the strata but extensive remnants of former erosion surfaces are evident. The valleys are narrow and deeply incised.

Rocky Mountain Area

This area extends more than 850 miles from Liard River to the Forty-ninth Parallel, and includes the Rocky Mountain Foothills, which form the eastern front of the southern part of the Cordillera, and the Northern, and the Southern Rocky Mountains. Each is divisible into

numerous subparallel ranges. The *Foothills* generally rise abruptly above the Interior Plains and are mainly linear ridges and hills of Mesozoic sandstones with some mountainous parts. The *Northern Rocky Mountains* are broadest and highest around Mount Churchill (10,500 feet in elevation) in the Muskwa Ranges. There the terrain is as rugged as any other part of the Rockies, having been carved by glaciation from great sections of stratified rocks. A number of alpine glaciers occur around the higher peaks. The Rockies are narrow where crossed by Peace River. To the south the Hart Ranges of the *Southern Rocky Mountains* rise gradually and form a long saddle of relatively subdued mountains with summits 6,000 to 9,000 feet in elevation. The Continental Ranges are broader and elevations increase to that of the highest peak, Mount Robson, at 12,972 feet. A number of high peaks on the continental divide cluster around Columbia Icefield — the largest of many alpine glaciers in the Rockies. Southward, peaks up to more than 10,000 feet elevation occur at intervals or in groups along the mountain backbone to near Crowsnest Pass, where they are generally lower, rising again south of the Forty-ninth Parallel. Continental Ranges are characterized by linear ranges with great cliffs and precipitous faces of bare grey rock, sculptured by glaciation from thick sections of carbonate strata.

Interior System

Brooks Range Area

British Mountains rise close to the sea, abruptly from the narrow Yukon Coastal Plain on the north and become lower inland, towards Porcupine Plateau. They comprise a maze of V-shaped valleys fingering among branching ridges. Many streams have cut courses radially across the ridges, and the valley of the largest, the Firth River, completely traverses the mountains. Only small areas in the highest part near the Alaskan Boundary are rugged. There the summits reach about 6,000 feet elevation and seem to have remained unglaciated. Southwest of Old Crow Plain *Old Crow Range* is a subdued range of large rounded hills commonly crowned by castle-like outcrops of granitic rocks.

Northern Plateau and Mountain Area

Ogilvie Mountains lie between Tintina Trench and Taiga Valley, south of Taiga Ranges. They are composed of sedimentary strata intruded by granitic stocks. Their highest peaks and ridges reach 7,000 feet elevation. During the Pleistocene these mountains contained alpine and valley glaciers which coalesced into continuous ice fields in their eastern part.

Yukon Plateau is a major physiographic unit containing many subdivisions. It is gently basin-like, its central parts lying at about 4,000 feet elevation. The surface rises towards the borders of the plateau and also lies at higher elevations in the local mountain ranges, such as the Pelly Mountains. The plateau is drained by the vast headwaters of the Yukon River system. Near its headwaters the river is entrenched about 2,200 feet below the plateau surface which lies at about 4,400 feet elevation. It declines at grades of 3 or 4 feet per mile, with few rapids, to less than 900 feet at the Alaskan Boundary. During the Pleistocene ice encroached from the mountains on all sides except the northwest but did not cover the Klondike Plateau.

On the northeast and east Yukon Plateau is bordered by *Selwyn Mountains,* which were extensively glaciated and comprise many rugged ranges separated by broad valleys. A few alpine glaciers occur on the highest summits between 7,000 and 9,000 feet elevation. Mount Keele, at 9,700 feet elevation, is the most outstanding peak.

Hyland Plateau is formed of rolling hills rising to 4,000 feet elevation and broad valleys, some aligned with structural features. The whole plateau was glaciated; the drift deposits form a widespread mantle and clog the main valleys. Along the southern boundary with Liard Plain are four isolated mountain groups reaching above 5,000 feet elevation. *Liard Plain* is a low-lying area into which Liard River is entrenched. The plain is less than 3,000 feet in elevation and much is at 2,000 feet. It is mantled with glacial and outwash deposits, including a great compound esker.

Central Plateau and Mountain Area

Stikine Plateau forms the southern continuation of the Yukon Plateau. It notably has a number of late Tertiary and Pleistocene volcanoes, the most prominent of which are the great shield volcano of Level Mountain, a low, dome-shaped mass reaching 7,101 feet whose lavas cover more than 700 square miles, and Mount Edziza, a complex cone, 9,143 feet in elevation.

Cassiar and *Omineca Mountains* form a belt of massive crystalline rocks, whose highest peaks are about 8,000 feet in elevation and carry a number of alpine glaciers. *Skeena Mountains* to the south and west are somewhat lower in general elevation and are composed largely of folded stratified rocks of uniform resistance so that structure is an important factor in determining the positions of mountains and valleys. Southwest of Skeena Mountains the *Nass Basin* is an area of low relief with a gently rolling floor for the most part lying below 2,500 feet elevation. The basin is almost encircled by mountains, but large gaps connect it with Skeena River valley and the sea via Nass River valley.

Hazelton Mountains comprise a group of ranges traversed by Skeena River. They are largely underlain by folded sedimentary and volcanic rocks and isolated intrusions of igneous rocks in contrast to the almost wholly granitic terrain of Coast Mountains to the west and the sediments of Skeena Mountains to the east. The highest peaks range between 8,200 and 9,200 feet in elevation.

Southern Plateau and Mountain Area

The main unit is the *Interior Plateau* which extends half the length of British Columbia north of the Forty-

ninth Parallel. Three large shield volcanoes stand near the middle of the western side, and flat-lying Tertiary to Recent lava flows form the bedrock over large parts. The plateau surface generally lies between 4,000 and 5,000 feet in elevation, rising from Fraser Basin in the northeast where the general level is below 3,000 feet to above 6,000 feet in Columbia Mountains. *Fraser Basin* has a flat, gently rolling surface mainly covered by glacial drift into which the Fraser River system and other main streams are commonly incised more than 1,000 feet. The glacial deposits include well-developed drumlinoid features, pitted terraces, simple and compound eskers, and large areas have glacial lake deposits.

Columbia Mountains contain many peaks more than 10,000 feet in elevation and are generally as high as the Southern Rocky Mountains east of the Trench. They constitute the main area of the interior wet belt of British Columbia. They were ruggedly sculptured during the Pleistocene glaciation, and their higher parts still contain many glaciers. *Columbia Highlands* to the west are lower and less rugged.

Western System

Coast Mountain Area

This area includes the Coast Mountains and the Cascade Mountains. The former are composed of crystalline gneisses and granitic rocks in contrast to the Rocky Mountains which are made up of stratified rocks. The two great mountain groups also differ in that the summits of the Rockies are relatively blocky whereas the peaks and ridges of the Coast Mountains are commonly serrate with tooth-like pinnacles. Only Peace and Liard Rivers traverse the Rocky Mountains, whereas nine or more large streams cross the Coast Mountains.

Coast Mountains include three major divisions, the Boundary, the Kitimat, and the Pacific Ranges. *Boundary Ranges* extend from southwestern Yukon along the boundary of the Alaskan panhandle. Their summits form a belt of rugged peaks generally 7,000 to 10,000 feet in elevation capped by several large ice fields. These ranges are cut into several segments by great transverse valleys containing Taku, Stikine, and Iskut Rivers. In places relief along the sides of the valleys reaches 9,000 feet. Large glaciers move down tributaries from nearby ice fields to 500 feet above sea level; in Alaska several reach the sea. The *Kitimat Ranges,* their peaks rarely above 8,000 feet elevation, form a saddle between the higher Boundary and Pacific Ranges. Glaciers are few and small. *Pacific Ranges* form the main southern part of Coast Mountains. Many peaks are between 9,500 and 11,000 feet elevation; the highest, Mount Waddington, is 13,177 feet elevation. Like the Boundary Ranges, the higher peaks are surrounded by ice fields; many large glaciers extend to low elevations but none reaches the sea. They are also cut into segments by great transverse valleys and fiords or inlets, the largest being those of the Klinaklini and Homathko Rivers.

Cascade Mountains are separated from the southeast part of Coast Mountains by Fraser River Canyon and continue into the United States. They are composed of the Skagit, Hozameen, and Okanagan Ranges. The highest peaks reach 8,000 to 8,500 feet elevation. In Skagit and Hozameen Ranges the high ridges are serrate, whereas lower ones are rounded by glaciation. In the Okanagan Range, although deeply cut by cirques, the summits still show the rolling contours of an old erosion surface rising from the Interior Plateau.

Outer Mountain Area and Coastal Lowlands

The Outer Mountain Area includes the St. Elias Mountains and the Insular Mountains forming Queen Charlotte and Vancouver Islands. The Coastal Lowlands border the Outer and Coast Mountain Areas.

St. Elias Mountains are the highest mountains in Canada and with their great peaks ranging upwards to the 19,850 feet elevation of Mount Logan, they are one of the great mountain belts of the world. In the *Icefield Ranges,* the heart of the mountains near the Alaska boundary, the main peaks stand as somewhat isolated blocks separated by broad valleys filled by ice fields to 8,000 feet or more. As the limit of permanent snow is at about 7,000 feet, these mountains present great masses and expanses of ice and snow. Mount Logan is the highest point of a large block on which at least four others rise to more than 18,000 feet. This tableland in consideration of the elevation of the long saddles comprising Mount St. Elias at 18,008 feet, Mount Lucania at 17,150 feet, and Mount Steele at 16,439 feet, suggests the former presence of an even erosion surface. The ice fields are the source of many great valley glaciers that radiate to the sea along the Alaskan Coast and inland towards Shakwak Trench in Canada. The lesser peaks, many more than 10,000 feet, tend to form groups or ridges between the valley glaciers. Northeast of Icefield Ranges the lower *Kluane Ranges* with peaks up to 8,500 feet flank Shakwak Trench. These ranges are much less rugged than the Icefield Ranges and contain only small alpine glaciers.

Hecate and *Georgia Lowlands* bordering Coast Mountains are formed by an old erosion surface that emerges from the sea bevelling hills of crystalline gneissic and igneous rocks and rises gradually inland. The limit is arbitrarily drawn at about 2,000-foot contour. Even in its lower parts the lowland has been dissected by glaciers and bedrock is exposed. Inland much of the surface is covered by muskeg. *Georgia Lowland* includes an area of soft folded strata along the east and south sides of Vancouver Island where differential erosion has formed cuesta-like ridges below 2,000 feet elevation. It includes the islands of the Strait of Georgia and a fringe of the mainland. *Fraser Lowland* adjoining Fraser River is composed largely of unconsolidated material with some hard bedrock forming rolling hills up to 1,000 feet

elevation. Part is formed by the Fraser Delta.

Queen Charlotte Ranges are the backbone of Queen Charlotte Islands, and in the southeast their summits reach above 3,500 feet elevation. These ranges and Skidegate Plateau to the northeast were glaciated by a local ice sheet (Holland, 1964). *Queen Charlotte Lowland* is formed of gently dipping Tertiary strata covered by a veneer of glacial till and thick outwash deposits in its northeast part. Inland, it rises southwestward to between 500 and 1,000 feet above sea level.

Vancouver Island Ranges form the main body of the island and their higher peaks reach 7,000 feet elevation. They were sculptured by local alpine glaciers but some ice came from the mainland to the east carving deep U-shaped valleys. *Nahwitti Lowland* on the north end of Vancouver Island is a rolling lowland carved of slightly folded strata, generally lying below 1,000 feet elevation. On the west coast of Vancouver Island a narrow, broken lowland forms *Estevan Coastal Plain.* It is underlain mainly by gently dipping, soft Tertiary strata forming a low and flat plain, but parts are cut on relatively massive rocks forming low hills and reach about 250 feet above sea level.

Pacific Continental Shelf

Off the British Columbia coast the Pacific Continental Shelf is narrow and its surface drops abruptly to ocean depths along a nearly straight line that closely parallels the headlands of the outer islands. West of Queen Charlotte Islands the rim of the shelf hugs the shore and at its closest point, 4 miles from land, descends below 1,000 metres. Off Vancouver Island the edge of the shelf is generally 20 miles or more to sea, reaching a maximum of nearly 50 miles at the entrance of Juan de Fuca Strait and a minimum of about 6 miles off Brooks Peninsula.

North of Queen Charlotte Islands, at Dixon Entrance, a broad channel, generally 200 to 400 metres deep, extends towards Portland Canal and branches among the Alaskan Islands. The width of Dixon Entrance is believed to have been occupied by Pleistocene ice moving from the Coast Mountains and the Alaskan Islands. The channel contains a transverse bank as shallow as 40 metres in the middle part, near the outer end.

At the entrance of Queen Charlotte Sound the edge of the shelf has four short but deep embayments at oceanic depths. From these, three broad channels more than 200 metres deep extend landward to near the islands off the mainland coast. The channels are separated by two broad banks. The northern and smaller one is a little less than 125 metres deep, whereas much of the larger one is less than 40 metres deep on its landward part. The north channel projects into Hecate Strait and lies mainly on the east or mainland side thereby leaving a broad shallow area shelving gently easterly and stretching across most of the strait. In the north the shallows nearly reach the islands of the mainland coast so that Hecate Strait is no more than 20 metres deep for about 4 miles. The southern channel extends towards the straits between Vancouver Island and the mainland. The straits are narrow, surround many islands, and contain numerous deep holes and channels more than 200 metres deep. This is characteristic also of many channels and fiords along the coast. These conditions extend through to the Strait of Georgia, a large part of which is occupied by an irregularly shaped basin, at 400 metres or more in several places.

St. Lawrence Lowlands

The St. Lawrence Lowlands borders the Shield on the southeast extending from the west end of Lake Huron and the head of Lake Erie northeasterly to the Strait of Belle Isle. It comprises three units; the West St. Lawrence Lowland, the Central St. Lawrence Lowland, and the East St. Lawrence Lowland. The province is floored by unfolded Paleozoic strata and the west and central lowlands are plain-like areas. The East St. Lawrence Lowland, though formed of widely separate parts, maintains the same general plain-like form and structure. The whole province was affected by Pleistocene glaciation and its surficial deposits and features are common phenomena.

West St. Lawrence Lowland is broken into two parts by the Niagara Escarpment which extends from Niagara River in a sinuous westerly and northwesterly course to Bruce Peninsula and the Manitoulin Islands. The escarpment is a northeasterly facing cuesta and is the result of the resistance to erosion of Silurian carbonates that dip gently southwest. The escarpment attains an elevation of about 650 feet in the south and a relief of 250 to 300 feet. Along the south shore of Georgian Bay it attains an elevation of 1,700 feet, about 1,000 feet above the bay. The surface west of the escarpment slopes gradually southwestward through an area of rolling topography of low relief to the level of Lake Erie, 572 feet above sea level. East of the escarpment the land rises gently northward from Lake Ontario at 246 feet elevation, to a maximum height of 1,000 feet south of Georgian Bay. At several places east of Lake Simcoe the boundary with the Shield is marked by a low escarpment. The Precambrian rocks of the Frontenac Axis separate West and Central St. Lawrence Lowlands.

Central St. Lawrence Lowland includes the area between the Ottawa and St. Lawrence Rivers, straddles the St. Lawrence as far as Quebec City, and extends a short distance beyond on the north shore only. East of the St. Lawrence River the lowland is bounded by 'Logan's Line,' a fault zone that marks the northwestern border of the hilly Appalachian Region. North of Ottawa and St. Lawrence Rivers the flat-lying Paleozoic strata underlying the lowland are either faulted against or lie upon the crystalline rocks of the Shield. The land is rarely more than 500 feet above sea level, except for the seven Monteregian Hills formed of intrusive igneous rocks.

East St. Lawrence Lowland includes Anticosti Island, the Mingan Islands, and several small areas bordering the Gulf of St. Lawrence and Strait of Belle Isle, and the Newfoundland Coastal Lowland. Anticosti Island is about 125 miles long with a maximum width of about 35 miles. It is a south-dipping cuesta of carbonate strata. Wave-cut terraces up to 400 feet elevation occur on both north and south sides, being generally wider on the south side. The Mingan Islands and two areas on the north coast of the Strait of Belle Isle are composed of southerly dipping Paleozoic strata. In all probability a great part of the floor of the Gulf of St. Lawrence around Anticosti Island and extending to the Newfoundland Coastal Lowland constitutes a submerged part of the East St. Lawrence Lowland. *Newfoundland Coastal Lowland* is formed of a belt of soft, mainly unfolded Paleozoic strata that extends along the northwest coast of Newfoundland Island and includes Port au Port Peninsula in the southwest. Most of the belt is less than 400 feet in elevation although the summit on Port au Port Peninsula is at 1,160 feet elevation and east of St. John Bay is 2,075 feet.

Appalachian Region

The Appalachian Region in Canada is divided into a large number of physiographic units, which here are grouped together and somewhat generalized. The physiography is dominated by a well-developed peneplain, probably of Cretaceous age, which is generally highest in the northwest and slopes gently southeastward to the ocean. The various subdivisions are formed by the differential erosion of soft and hard rocks to form lowlands, highlands, and uplands. Seaward the Appalachian Region is bounded by the Atlantic Continental Shelf.

On the island of Newfoundland the Appalachian Region comprises three highland, two upland, and four lowland areas. The highlands are collectively called the *Newfoundland Highlands.* As a rule they are rugged with steep slopes, and are formed of crystalline Paleozoic and Precambrian rocks. Their elevations vary from 600 to 2,670 feet. The upper surface, formed by the peneplain, slopes gently southeastward. The western border of the highlands may be precipitous or pass without sharp break to the higher areas of Newfoundland Coastal Lowland. Two small lowland areas occur within the highlands area, the St. Georges Lowland lying along the southwest coast and the Grand Lake Lowland extending from White Bay to Grand Lake. Both lowlands are underlain by little metamorphosed rocks and seldom exceed 200 feet in elevation.

The *Atlantic Uplands of Newfoundland* lie between 600 and 1,000 feet elevation and represent parts of the peneplain occurring at lower levels because of the southeasterly slope rather than because the rocks are less resistant than those underlying the adjacent highlands. Where stream erosion has cut deeply, the uplands are rugged and rocky but elsewhere they present a rolling terrain of low relief. The uplands surround the small Bay d'Espoir Lowland which is underlain by soft Paleozoic sediments.

The *Newfoundland Central Lowland* is underlain mainly by Paleozoic strata but includes a few intrusive bodies that form prominent hills. It extends from sea level up to 500 feet elevation; inland it is surrounded by the uplands carved from igneous rocks. The surface is gently rolling and generally immediately underlain by glacial drift.

Nova Scotia is divisible into three highland areas, three uplands, and several small lowlands. As in Newfoundland the same peneplain is present on the highlands and is also apparent on the uplands. *Nova Scotia Highlands* include Cobequid Mountains in the west, the Antigonish Highlands in the centre, and the Cape Breton Highlands to the northeast. Cobequid Mountains have a flat to rolling surface, are about 10 miles wide, and reach elevations of about 900 feet. Antigonish Highlands are more dissected than Cobequid Mountains, but have the same type of surface and reach about the same elevations. The Cape Breton Highlands are deeply dissected around the margins, but the interior has a remarkably flat surface lying at elevations of 1,300 to 1,500 feet.

The uplands of Nova Scotia include North Mountain and the Atlantic Uplands of Nova Scotia. *North Mountain* is a narrow, flat-topped ridge, averaging about 550 feet high, formed of hard Triassic basalt. It extends along the southeast side of the Bay of Fundy for 120 miles and rises abruptly above the Annapolis Lowland. The *Atlantic Uplands of Nova Scotia* lie along the entire southeast coast for some 350 miles and contain small irregular lowland areas on Cape Breton Island. The old peneplain surface is a main feature and slopes across it from elevations of about 600 feet on the northwest side to the shore of the Atlantic on the southeast.

The principal lowlands of Nova Scotia are: the Cumberland Lowland, part of the larger Maritime Plain, but lying on the north and east sides of Cobequid Mountains; the Minas Lowland surrounding Minas Basin and Cobequid Bay; and the Annapolis Valley. These lowlands are areas of soft, late Paleozoic and Triassic sediments.

New Brunswick falls naturally into three large units, highland, upland, and lowland areas. *New Brunswick Highlands* are U-shaped, with the west arm covering much of the central part of the province, the curved part occupying the southern part, and the east arm bordering the Bay of Fundy. In as much as the surfaces of the accordant summits of the highlands represent the southeast-sloping peneplain, the northwest parts reach elevations greater than 2,000 feet and are higher than those bordering the Bay of Fundy, which are only about 1,000 feet. Along the Bay of Fundy the highlands present considerable diversity. They are composed of ridges of hard rock with minor lowlands of weak strata.

Chaleur Uplands lie astride the provincial boundary of Quebec and New Brunswick. They are 800 to 1,000 feet

high, developed on folded Paleozoic strata. The uplands present remarkably uniform, concordant summits whose regularity is broken only by a few hills and ridges rising slightly above the general level. Valleys such as those of Saint John and Restigouche Rivers are deeply entrenched into it. The boundary between the New Brunswick Highlands and Chaleur Uplands is not well defined but the latter are considerably lower than Notre Dame Mountains of Quebec to the north.

Maritime Plain stretches around the coast of New Brunswick and Nova Scotia from the south shore of Chaleur Bay nearly to Cape George. It extends inland as far as New Brunswick Highlands and includes Prince Edward Island and the Magdalen Islands. The plain is an area of low relief. Inland it reaches 600 feet elevation and slopes gently towards the coast. The underlying rocks are mainly flat or gently dipping strata through which, particularly in the Magdalen Islands, small intrusions stand as prominent hills. It probably extends beneath the Gulf of St. Lawrence to join with St. Georges Lowland of Newfoundland.

The Chaleur Uplands are bounded on the north by *Notre Dame Mountains,* which extend from near Thetford Mines to Gaspé Bay, a distance of some 400 miles. The eastern part is also known as the Shickshock Mountains, the highest elevations being in excess of 4,000 feet in the north. To the southwest the summits are lower and merge with Eastern Quebec Uplands. Mountain tops present the flat to undulating surface of the old peneplain. In the southwestern part of Appalachian Quebec are two other highland areas separated by part of the Eastern Quebec Uplands. The western of these, the *Sutton Mountains,* is a continuation of the Green Mountains of Vermont; they are 40 miles in extent and have a maximum of 3,175 feet. *Megantic Hills* lie astride the International Boundary and are part of the larger White Mountains of the New England States. Sutton Mountains and Megantic Hills are both developed on anticlinoria and are highest in the south. To the northeast they blend with *Eastern Quebec Uplands* which has a maximum elevation of 1,000 feet decreasing gradually to the northwest and merging with the East St. Lawrence Lowland without a distinct topographic break.

Atlantic Continental Shelf

Adjoining the Grenville and Appalachian Regions and forming the continental margin of North America, are the Gulf of St. Lawrence and the Atlantic Continental Shelf. The floor of the *Gulf of St. Lawrence* is relatively flat and featureless with gentle rises, one of which breaks surface as Magdalen Islands, and linear lows which represent old, glacially deepened river valleys. The main feature, the Laurentian Channel, is 60 miles wide, steep-sided, and flat floored. It shelves from depths of 200 metres near Quebec to 400 metres as it extends across the continental shelf to the ocean basin. A major tributary joins it from the Strait of Belle Isle.

The *Atlantic Continental Shelf* is generally shallow, less than 200 metres and slopes gently at a gradient of less than 1:1000, to where the gradient increases sharply to more than 1:40. The shelf is divided by Laurentian Channel; to the south is the Scotian Shelf and Georges Bank, the part to the north is the Grand Banks.

Scotian Shelf has an average width of about 125 miles. Along its length and lying some 10 to 80 miles offshore is a series of 200-metre depressions such as La Have Basin and Emerald Basin. A nearly continuous ridge or 'bank' 40 to 120 metres deep extends along the seaward edge of the shelf, divided by low saddles into Browns Bank, La Have Bank, Emerald Bank, Sable Island Bank, and Banquereau Bank, and culminating at Sable Island, a low, sandy island 25 miles long. The banks are composed of flat-lying beds and are like mesas produced by erosion. Many submarine canyons occur in the continental slope and the continental rise, the largest of which is The Gully heading in the shelf about 25 miles east of Sable Island. The *Grand Banks,* north of Laurentian Channel, are up to 300 miles wide and 450 miles long. Northeast of Newfoundland relief is 100 to 200 metres, exceptionally large for a continental shelf. The shelf break lies at 300 metres, which is abnormally deep. The southern part of Grand Bank includes low depressions and several smaller banks, Saint Pierre Bank, Green Bank, and Whale Bank. Virgin Rocks is a shoal 130 miles southeast of Avalon Peninsula rising to within 18 feet of the ocean surface. It is part of a dissected bank which rises 200 to 300 feet above the shelf floor.

Flemish Cap, 400 miles east of Newfoundland, is considered part of Atlantic Continental Shelf but is separated from Grand Banks by a channel 650 fathoms deep. A broad ridge, 60 to 100 miles wide, extends southeasterly from the Grand Banks down the continental slope towards the Mid-Atlantic Ridge. This feature, the Southeast Newfoundland Ridge, is part of the continental rise and marks the boundary between the Newfoundland Basin to the north and the North American Basin to the south.

The southern part of *Labrador Shelf* extends about 400 miles from the coast and includes an area, relatively shallow and undulating, the deeper parts of which lie near 250 metres and large shallower areas lying at less than 80 metres. Between Hamilton Inlet and Hudson Strait the shelf is about 75 miles wide dropping from a rim at 250 metres to the depths of Labrador Sea. This part has a line of broad banks near 80 metres depth which are separated from the coast by wide hollows at 200 metres or more depth.

Davis Shelf lies beneath Davis Strait and Baffin Bay between Canada and Greenland. Off Baffin Island, below Cumberland Peninsula and Cape Chidley, it is generally shallow being little more than 900 metres deep in many parts. The shelf is, however, deeper and wider than farther north and its surface slopes to a depth of about 600 metres before the steep edge is reached some 100 miles or more off Hall Peninsula. Opposite Hudson Strait

the sea floor also drops to oceanic depths of about 3,000 metres and an embayment with depths of 600 metres or more leads across the shelf. At the head of Baffin Bay east of Devon Island, it is also comparatively shallow, few parts reaching 600 metres. The soundings although scattered suggest a rolling surface sloping mainly southeasterly from 300 to 500 metres. Off Lancaster Sound a broad channel 800 metres or more deep extends towards the northern part of the basin in the central part of Baffin Bay where depths of 2,000 metres or more are reached. Along the northeastern coast of Baffin Island the slope into the basin steepens and a narrow shelf up to 30 miles wide with a somewhat sinuous rim at 200 metres deeper drops abruptly into the basin. Three channels extend across the shelf off Pond Inlet, Buchan Gulf, and Home Bay.

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III. The Canadian Mineral Industry

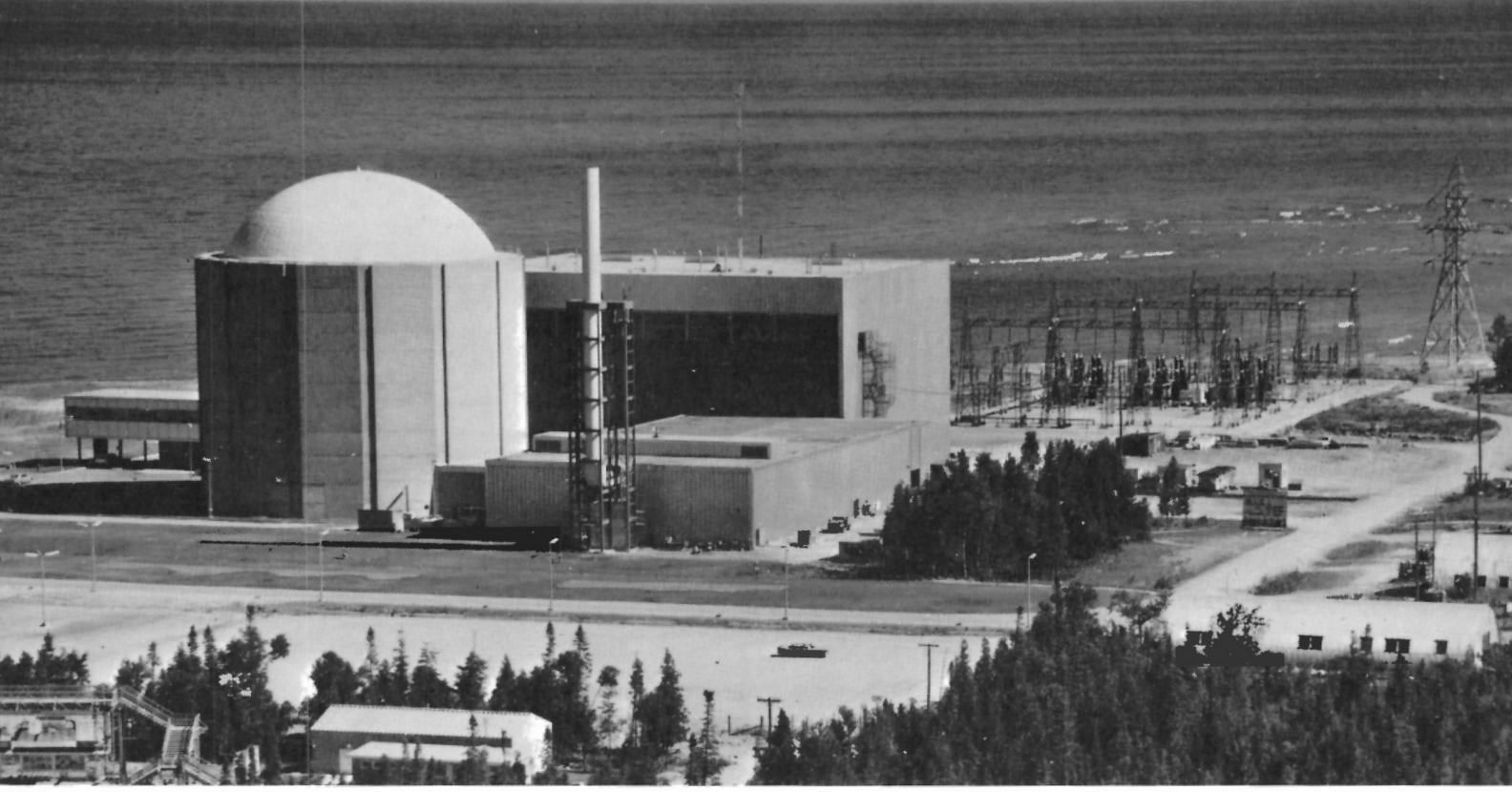
Economic Indicators of Growth 33

Growth of the Industry 36

Mineral Reserves 38

Mineral Supply and Demand 40

Future Developments 41



The Canadian mineral industry spans four centuries of development. In the first 300 years the record was one of intermittent activity, and even during the past 100 years progress in terms of production was comparatively slow until the end of World War II, when it began to increase rapidly (Table III-1). The industry had an output value of $64 million in 1900 and of $500 million in 1945, reaching $1 billion in 1950 and more than $4.7 billion in 1968. Relative growth since 1940 for the three sectors — metals, industrial minerals, and fuels — is illustrated in Figure III-1.

TABLE III-1 *Mineral production in Canada, selected years, 1890–1968* ($ millions)

Year	Metals (a)	Non-metals and Structural Materials	Fuels	Total
1890	3.6	6.4	6.8	16.8
1900	40.5	8.6	15.3	64.4
1910	49.4	24.7	32.7	106.8
1920	77.9	64.2	85.8	227.9
1930	142.6	69.1	68.2	279.9
1940	382.4	68.6	78.8	529.8
1950	617.2	227.0	201.2	1,045.4
1960	1,406.6	520.1	565.8	2,492.5
1965	1,907.6	761.4	1,076.5	3,745.5
1966	1,984.4	837.5	1,150.6	3,972.5
1967	2,285.5	854.5	1,258.9	4,398.9
1968	2,492.6	890.2	1,342.5	4,725.3

Source: Dominion Bureau of Statistics
(a) Does not include aluminum metal

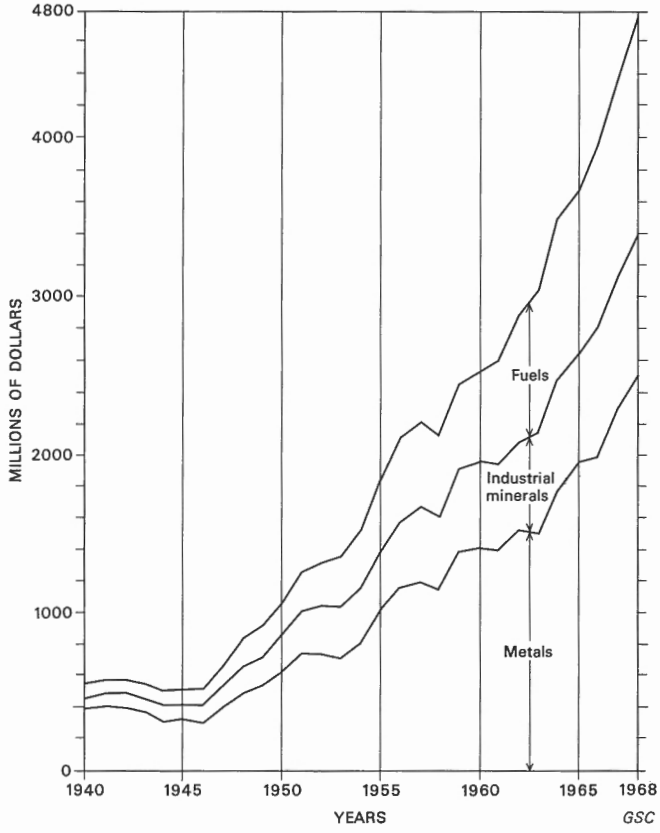


FIGURE III-1. Value of mineral production.



III

The Canadian Mineral Industry

R.B. Toombs[1]

Uranium-fueled electrical generating plant, Douglas Point, Lake Huron, Ontario.

ECONOMIC INDICATORS OF GROWTH

The importance of the mineral industry to the Canadian economy, and its recent growth compared with the general economic growth of the nation, can be assessed by means of economic indicators. They are particularly significant in judging the industry's importance and performance during its rapid postwar growth. One indicator is shown in Table III-2 by relating mineral production to gross national product (GNP). From 1945 to 1968 the value of mineral production increased 9.5 times while the GNP multiplier was 6.0. At the end of the war mineral value was less than 5 per cent of GNP, but it has since increased to almost 7 per cent. This comparison of mineral industry value at the primary stage is only a minimum assessment of the industry's importance to the economy.

Net value of production, a "value added" concept, affords a more direct indicator of the mineral industry as a producing sector in the Canadian economy. Net values

[1] Mineral Resources Branch

TABLE III-2 *Mineral production related to gross national product, selected years, 1930–1968* ($ millions)

Year	Mineral Production (a)	GNP	Mineral Production as % of GNP
1930	279.9	5,720	4.9
1940	529.8	6,713	7.9
1945	498.8	11,863	4.2
1950	1,045.5	17,955	5.8
1955	1,795.3	27,895	6.4
1960	2,492.5	37,775	6.6
1965	3,745.5	54,897	6.8
1966	3,972.5	61,421	6.5
1967	4,398.9	65,608	6.7
1968	4,725.3	71,454	6.6

Source: Dominion Bureau of Statistics
(a) Does not include aluminum metal

TABLE III-3 *Net values of production, 1950 and 1967, all industries* ($ millions)

	1950	1967
Mining	657.3	2,910
Mineral processing (a)	512.1	2,090
Total mineral industry	1,169.4	5,000
Agriculture	1,832.9	2,728
Forestry	487.1	687
Fisheries and trapping	97.4	175
Electric power	313.3	1,234
Total primary production	3,900.1	9,824
Construction	1,544.7	5,148
Manufacturing (ex. mineral)	5,430.0	14,915
Total secondary production	6,974.7	20,063
Canada total	10,874.8	29,887

Source: Dominion Bureau of Statistics
(a) Primary iron and steel, nonferrous smelting and refining, petroleum refining, non-metallic mineral processing.

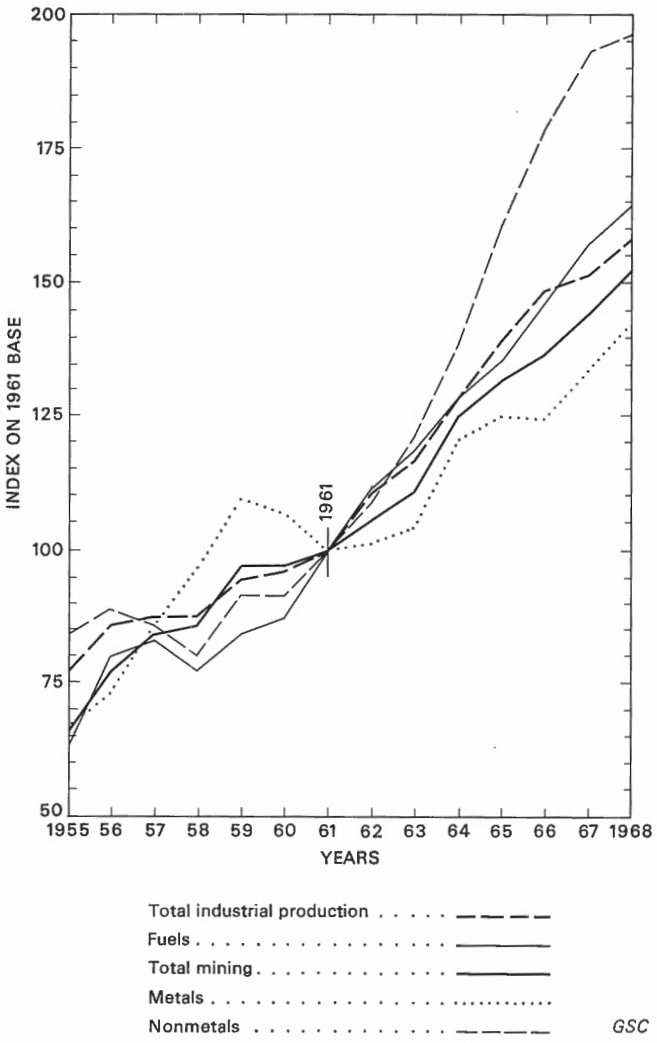


FIGURE III-2. Indexes of physical volume of mineral production and of total industrial production.

of production for mining[1] and mineral processing outpaced all other commodity-producing industries of the economy except electric power in the period 1950 to 1967 (Table III-3). The mineral sector increased from 30 to 51 per cent of primary production and from 11 to over 16 per cent of total primary and secondary industry production in the Canadian economy. During the 1960's, value added in mining itself, as a proportion of value added in all primary industries, remained well above one third and overtook agriculture.

Growth of the mining industry in terms of volume indexes is shown in Figure III-2. The mining index rose six times from 1946 to 1968 compared with slightly less than four times for the industrial economy as a whole, the greater increase in mining being due mainly to the rapid climb of the fuels index. Of the manufacturing indexes, those for petroleum refining, products of petroleum and coal, non-metallic mineral products, and primary iron and steel outpaced the index for total manufacturing.

A further indicator of growth is the gross domestic product at factor cost (GDP[2]). The mining sector GDP increased almost fourfold from 1950 to 1968 and in the late 1960's was nine tenths of the agriculture GDP, compared with one third in the early 1950's. In the same period mining moved ahead from less than double the value of the forest industry GDP to four times its value.

Although the mineral industry has provided direct employment in mining for less than 3 per cent of the total labour force in recent years, capital expenditures in the mining sector have been 6 to 7 per cent of the country's total annual expenditures. In the late 1960's, capital expenditures in mining and in mineral-based manufacturing industries, and in pipeline transportation and marketing related to petroleum and natural gas were about 12 per cent of total capital expenditures in the Canadian economy. This comparison considerably understates expenditure generated by the industry inasmuch as much of the railway expansion, many new townsites particularly in northern Canada, a number of electric power projects and communication systems, and some harbour facilities have been based wholly on mineral developments. The industry has a similar multiplying effect in the labour force, several times the number employed in mining being dependent directly or indirectly on mineral industry enterprises.

Export value has been usually about 60 per cent of the industry's production value, making the industry the

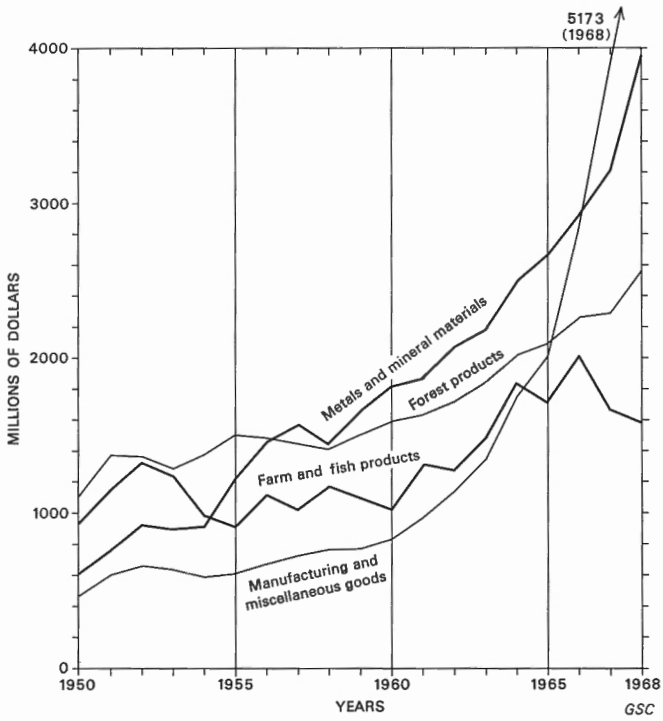


FIGURE III-3. Exports from Canada, 1950–68.

[1] Mining includes the primary production of metals, industrial minerals and fuels.

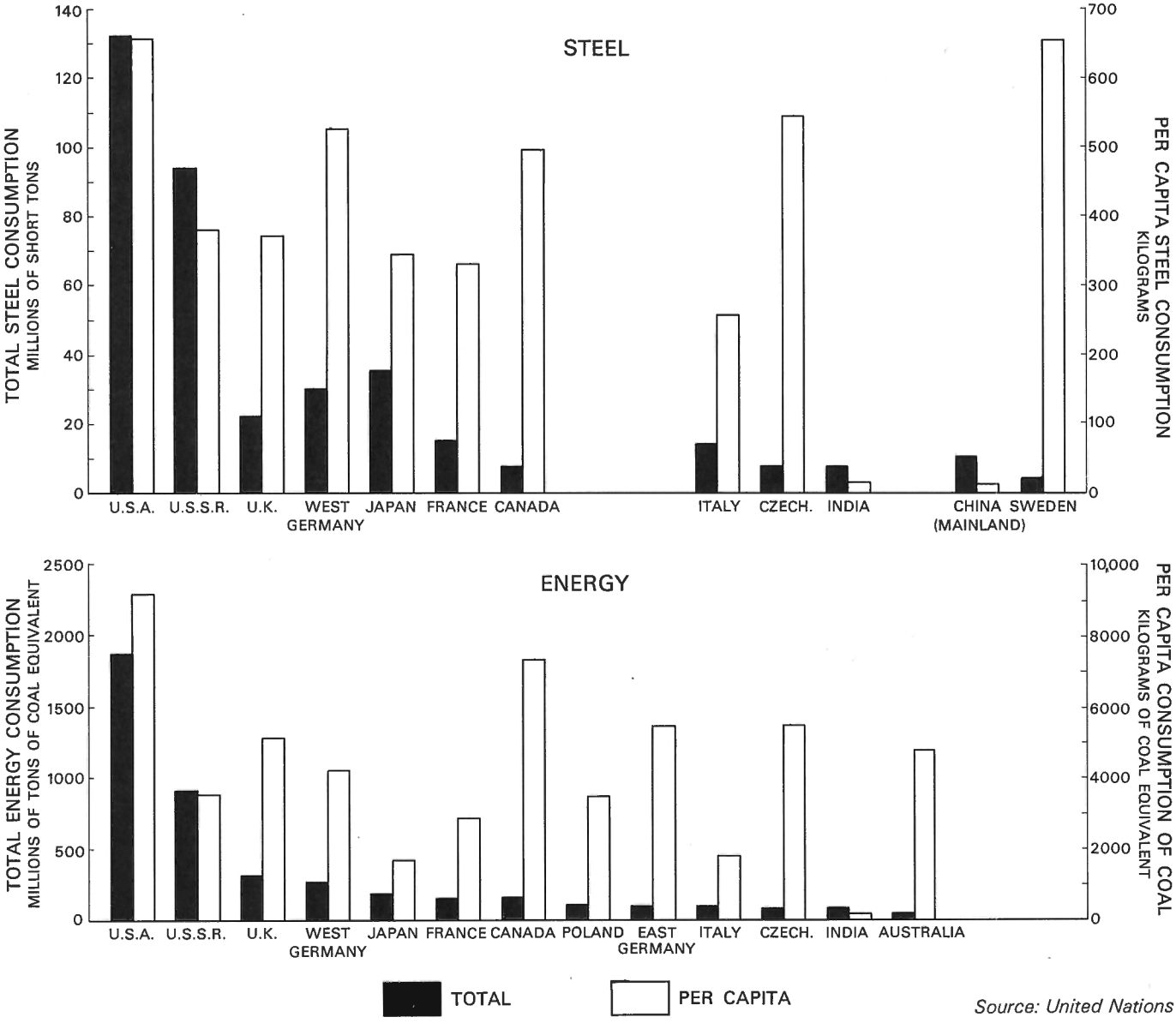
[2] GDP differs from the GNP by the exclusion of indirect taxes less subsidies and factor incomes paid to Canadian residents from production abroad, but including factor incomes paid to non-residents from production within Canada.

leading export sector in the economy. This position takes on added significance because 17 per cent of the GNP is accounted for by exports compared with 5 per cent in the United States. Figure III-3 illustrates mineral industry export trends relative to other exports.

The growth and magnitude of production, capital investment, and trade all illustrate the dynamic effect that expansion of the mineral industry has had on the economy since 1946. The multiplying effect of this expansion was possibly most apparent to 1956 during a time when the industry had to satisfy the accumulated demands of the post-World War II period and the strategic mineral requirements of the Korean emergency. Even in the late 1950's when the economy was less buoyant, there was little decline in the industry. By the mid-1960's the known resources were larger than those of the 1950's and the rate of industry growth was again increasing. In short, the mineral industry has been a dynamic force in times of industrial expansion and a strong support in less buoyant times.

Canada vis-à-vis Other Countries

In population Canada has about one half of one per cent of the world total and in area more than 7 per cent; the population density is one fifth that of the U.S.S.R., one tenth that of the United States, one hundredth that of Britain, and one twelfth of the world average. Canada holds third place among industrial nations in national income per capita and her high industrial activity provides an important domestic market for her minerals. However, because of a comparatively small population, Canada must export large quantities of mineral materials if her



Source: United Nations

FIGURE III-4. Total and per capita consumption of steel and energy in 1966.

mineral industry is to grow at a rate appropriate to the size of her resource base. Mineral exports do, in fact, help place the country in a prominent position in world trade. Canada is the world's fifth largest trading nation and is first in trade per capita among the world's leading traders; her growth and prosperity depend a great deal on international trade. The maturity of the country's economy vis-à-vis other nations is well indicated in energy and steel consumption comparisons. Per capita Canada ranks second in the world in the consumption of energy and fifth in the consumption of steel (Fig. III-4).

Canada's economic expansion in the 1950's was about equal to that of the United States. But some European countries, Japan, and U.S.S.R., starting from a very low level of economic activity in the late 1940's, made such major recoveries that their comparative rate of expansion outpaced that of both Canada and the U.S.A. in the late 1950's. However, Canadian mining and base-metals growth indexes kept pace with most of the world's highest; in the 1960's, the Canadian economy has been advancing rapidly due in considerable part to a buoyant mineral industry.

GROWTH OF THE INDUSTRY

The size, structure, and productive diversity of the Canadian mineral industry have been largely determined by resource developments, marketing progress, and government policies since the mid-1940's. Marked growth has taken place in recent years in all three sectors of the industry, the most outstanding being in the fuels sector. Value of total output increased more than ninefold in the period 1945 to 1968; the increase for fuels was more than fourteenfold, for industrial minerals (non-metals and structural materials) more than tenfold and metals eightfold. In volume, output increased almost sixfold, the increases for the fuels, industrial minerals and metals being ten, five and fourfold respectively. The all-time growth in value to 1945 was less than one eighth the increase that took place from 1945 to 1968.

Historical Background

The Canadian mineral industry has a long history dating from the days of the 16th- and 17th-century explorers Cartier, Frobisher, and Champlain. The coal and the iron-and-steel industries in the Maritime Provinces were well underway by the 1850's. In 1858 petroleum was discovered in southwestern Ontario and in the same decade placer gold discoveries were made in British Columbia. Asbestos was found in the Eastern Townships of Quebec in 1877 and nickel at Sudbury, Ontario, in 1883. A number of gold, silver, and base-metal discoveries were made during the 1880's and 1890's in southern British Columbia, including the Sullivan lead–zinc–silver deposits at Kimberley and the copper–gold deposits of the Greenwood and Rossland areas. Placer gold had been found on Yukon River in Yukon Territory in 1869; subsequent prospecting led to the famous Klondike gold-rush in 1897 and 1898 and to a peak value in Yukon gold production in 1900 of $22.3 million, a third of Canada's whole mineral production value for that year. The Cobalt strike of 1903 and the subsequent Kirkland Lake and Porcupine gold-camp developments opened up the northern Ontario mining region and led to the Noranda copper discovery in northwestern Quebec in 1921. Two years earlier, silver–lead deposits had been prospected in the Mayo district of the Yukon. The Turner Valley oil and gas field in Alberta was discovered in 1913. Each of these notable events in Canadian mining history was a forerunner of many others, all of which prepared the way for the greatly increased industry growth of recent years.

The Period 1946 to the Late 1960's

The 1946 to 1950 period was notable for important events in the development of the oil resources of western Canada, including the discovery of Devonian oil at Leduc; for a resurgence in gold and silver mining, and considerable expansion in asbestos and construction materials. There was comparatively little activity elsewhere in the mineral industry.

In the years 1951–55 there were a number of significant events that made an impact on Canadian mineral supply and demand that will continue for many years. Exploration, development, and new production in the Alberta and Saskatchewan oil lands, in the Quebec–Labrador iron-ore region, and in the uranium areas of Beaverlodge, Saskatchewan, and Elliot Lake, Ontario, made this a particularly important period. New copper deposits were opened up in the Gaspé and Chibougamau areas of Quebec, and major copper–zinc deposits at Manitouwadge, Ontario. Lead–zinc deposits at Pine Point, on Great Slave Lake, were explored. In the Bathurst–Newcastle district of New Brunswick important deposits of zinc–lead–copper were discovered. A nickel refinery was established at Fort Saskatchewan, Alberta; titanium-dioxide production facilities at Sorel, Quebec; and a ferrosilicon plant at Beauharnois, Quebec. Asbestos production commenced at McDame Creek in northern British Columbia, and there was good progress in developing other industrial mineral resources, particularly deposits of construction materials.

During the years 1956 to 1960, trends in development activities were mixed. Uranium production got underway on a large scale in the Beaverlodge, Elliot Lake, and Bancroft areas. Copper–zinc production began at Manitouwadge. Plans for producing nickel from the deposits of northern Manitoba were announced; mine and townsite development work at the Thompson nickel mine was started. Despite lead and zinc quotas imposed by the United States on Canadian imports, interest in the Matagami, Quebec, and Bathurst, New Brunswick, base-metal areas intensifed. In 1959, the United States Atomic Energy Commission announced that it would not take up its option for the purchase of Canadian uranium after 1962, and this had a dampening effect on uranium resource

development. The copper industry in Quebec, Ontario, and British Columbia was buoyant despite unsettled conditions of supply and demand in some other parts of the world. Iron-ore production rose significantly in 1959 in response to world demand and lengthy strikes in the United States. Canada prepared for a major role in world potash production following successful evaluation of the Saskatchewan deposits. During the late 1950's it became apparent that the U.S.S.R. was becoming as important an asbestos producer as Canada, but development and marketing of Canadian asbestos continued at high levels.

Large-scale expansion of natural gas production was made possible in 1960 by new export markets in the United States following completion of gas pipeline systems across Canada. Greatly expanded petroleum resources reached the point where production capacity began to exceed output by a large margin. Decline of the coal industry in the 1950's was arrested towards the end of the decade by increased federal government assistance and improved efficiency at some mines. During the late 1950's there were some indications of mineral industry retrenchment after the extensive expansion of the previous decade, but in the early 1960's this trend was reversed with increased activity in iron ore, petroleum, and the nonferrous mineral industries.

During the first half of the 1960's development of Canadian mineral resources again made rapid advances. Iron-ore development continued in Quebec and Labrador and new properties came into production in Ontario and British Columbia, establishing Canada as the world's fifth largest iron ore producer. Improvements in beneficiation processes to meet the requirements of changing steel technology permitted the opening up of deposits which even in the mid-1950's were considered too low-grade for economic operation. Canada's position as the world's main source of nickel was further strengthened by the opening of new mines in northern Manitoba and in Ontario and Quebec. Platinum metals production continued to grow, and Canada maintained its position as the world's second or third largest producer. In British Columbia and the Timmins area of northern Ontario, the development of copper deposits maintained Canada's position as fifth largest producer. Although the United States retained quotas on the imports of lead and zinc until late in 1965, important new mines were brought into production in the Bathurst area of New Brunswick, the Matagami area of Quebec, and at Pine Point, Northwest Territories. By the mid-1960's Canada was the world's leading producer of zinc and third largest of lead. Although gold and uranium production declined, some gold exploration continued, particularly in the Northwest Territories, and by the mid-1960's uranium exploration was again active in preparation for expected market growth in the 1970's. The search for other minerals also increased, and results were particularly successful for molybdenum in British Columbia and Quebec, putting Canada in second position in world molybdenum production by 1966.

In the industrial minerals sector of the industry, development of potash and sulphur raised Canada to a position of world prominence. The progress made in the Saskatchewan potash industry raised Canada's mine capacity to the equivalent of at least one third of the world's consumption by the late 1960's. In 1963 Canada became the world's second largest elemental sulphur producer as a result of expansion in natural gas production in western Canada, the sulphur being recovered as a byproduct. Year-by-year production increases continued for most non-metals and for the construction materials. An important deposit of asbestos was opened near Baie Verte on the north coast of the island of Newfoundland.

Increasing competition from other fuels led to a steep decline in coal production in the 1950's. Production was stabilized in the early 1960's as a result of large increases in government subsidies. The Cape Breton Development Corporation was formed in 1967 to bring about the transition of the economy of Cape Breton Island from coal mining to other forms of economic endeavour. In the mid-1960's, marketing prospects for western Canada coal began to improve. Natural gas resource development was accelerated following completion of continental pipelines in the late 1950's, and by 1964 annual output value reached sixth position in the mineral industry, a position it continued to maintain. Crude oil resource development more than kept pace with market growth; this has been Canada's leading mineral since 1953. New methods were being developed to open up the Athabasca oil sands, one of the world's greatest hydrocarbon reserves.

Structure and Capacity of the Industry

Although the ten leading mineral commodities — crude oil, copper, iron ore, nickel, zinc, natural gas, asbestos, cement, sand and gravel, and silver — account for almost four fifths of the annual value of mineral output, production of more than 50 other minerals is continually expanding. There is also increasing geographical diversification with pronounced effect on local economic development, particularly in areas outside the established mining camps. Company ownership is also quite diverse, although there are some exceptions for certain minerals. The large number of participants in the industry results in active resource development under a competitive free-enterprise economy. In the period 1955 to 1965 new mines brought into production numbered 125, most of which will operate for many years.

Extensive mineral processing facilities are continually being expanded, and as of 1968 they include aluminum plants at Arvida, Alma, Beauharnois, Shawinigan and Baie Comeau in Quebec and at Kitimat, British Columbia, with total annual capacity of 913,000 tons; copper or nickel – copper smelters at Sudbury in Ontario, Murdochville and Noranda in Quebec, and Flin Flon and Thompson in Manitoba, with capacity of 8.6 million tons annually. There are copper refineries at Montreal and Sudbury with total capacity of 452,000 tons annually, and nickel refineries at Port Colborne in Ontario, Thompson in

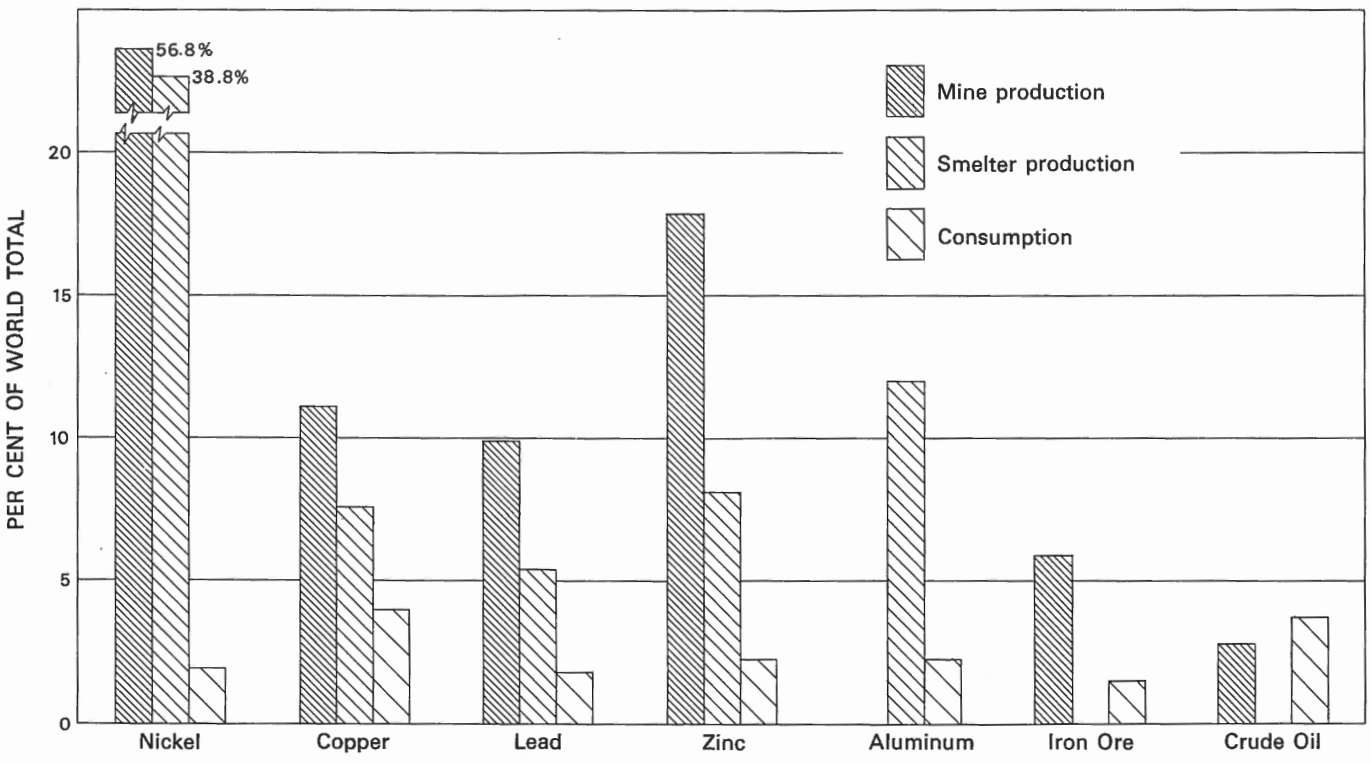
Manitoba, and Fort Saskatchewan, Alberta, with 166,000 tons capacity. The total productive capacity for all forms of nickel is 540 million pounds a year. There is a lead smelter and a refinery at Trail, British Columbia, with related refining facilities for the recovery of antimony, bismuth, cadmium, indium, and precious metals. Large zinc refineries operate at Trail, British Columbia, at Flin Flon, Manitoba, and Valleyfield, Quebec. More than four fifths of the steel ingot and casting capacity of 11 million tons is concentrated in four large plants at Hamilton and Sault Ste. Marie in Ontario and at Sydney in Nova Scotia; the remaining capacity is dispersed in 45 small plants throughout the country, many of which supply local needs. Capacity to supply the large-tonnage alloys, such as ferrosilicon, ferromanganese, and ferrochromium, has been more than adequate. There has also been sufficient capacity to meet domestic requirements of molybdic oxide, while requirements of ferromolybdenum, ferrocolumbium, and ferrovanadium may soon be met by domestic production. There have been adequate facilities for industrial mineral processing for many years. The country's petroleum refining capacity is usually capable of meeting domestic requirements, although imports are relied on to meet peak requirements of heating oils. Refineries at Montreal, Toronto, and Sarnia have accounted for about 60 per cent of total petroleum capacity in recent years. Natural gas processing capacity has expanded rapidly as new gas fields were opened up in western Canada; from 1955 to 1965 capacity increased thirteenfold to 6.1 billion cubic feet a year, and large annual increases have continued.

The expansion of production and facilities has been accompanied by an increase in foreign ownership of Canadian industry. In the period 1954 to 1963 nonresident ownership in mining and smelting rose from 53 to 61 per cent; in the petroleum and natural gas sector, it rose from 60 per cent to 64 per cent. This high degree of foreign ownership continued as the industry expanded further in the 1960's. Investment by U.S.-controlled companies accounts for well over four fifths of total nonresident ownership in petroleum and natural gas and in mining and smelting.

In the development of mineral resources during the period 1950 to 1968 annual capital expenditures on new plant, machinery and equipment, and repair expenditures, in the mining and oil industries to the production stage, rose from $163 to $1,100 million and aggregate expenditures in this period amounted to almost $12,000 million. About one half of these expenditures was related to oil and gas exploration, development, and production. Large amounts were also spent on mineral processing and fabricating facilities; in the mid-1960's these expenditures were of the same order of magnitude as expenditures on mineral resource development and production.

MINERAL RESERVES

Measured (proven) ore. Ore for which tonnage is computed from dimensions revealed in outcrops, trenches,



GSC

FIGURE III-5. Canadian mineral production and consumption as a per cent of world total—1965.

workings, or drillholes, and for which grade is computed from adequate sampling. The sites for inspection, sampling, and measurement are so closely spaced, on the basis of defined geologic character that the size, shape, and mineral content are well established.

Indicated (probable) ore. Ore for which tonnage and grade are computed partly from specific measurement, samples, or production data, and partly from projection for a reasonable distance on geologic evidence. The openings or exposures available for inspection, measurement, and sampling, are too widely or inappropriately spaced to outline the ore completely or to establish its grade throughout.

Inferred (possible) ore. Ore for which quantitative estimates are based largely on broader knowledge of the geologic character of the deposit and for which there are few, if any, samples or measurements. Estimates are based on assumed continuity or repetition for which there is geologic evidence; this evidence may include comparison with deposits of similar types. Bodies that are completely concealed but for which there is some geologic evidence may be included.

Potential ore. This category provides for the evaluation, as a resource, of mineral deposits the mining of which depends on improved prices, improvements in methods of mining or treatment, transportation facilities, etc.

Table III-4 gives measured and indicated reserves as established in 1956 and 1967 for some leading minerals. All reserves increased, notwithstanding the large aggregate production in the period covered by these two reserve surveys.

Reserves of nickel in 1967 in the measured and indicated categories totalled at least 7 million tons, about one half of world reserves, excluding the Communist bloc, of a grade of at least 1.25 per cent nickel. There are large additional resources of lower grade in the potential category throughout the world, particularly in the nickeliferous iron laterites of Cuba and Indonesia. The Canadian reserves consist of sulphides with an appreciable content of platinum and other metals.

Iron ore reserves are very large. The greatest of potential ores are in the Labrador Geosyncline, particularly in the southern part, from Wabush Lake in Labrador to Mount Wright, Mount Reed, and Lac Jeannine in Quebec, a distance of 100 miles, where the many deposits total over 10 billion tons of ore. There are many other deposits throughout Canada, particularly in central Quebec, northern Ontario, Baffin Island, and Yukon Territory, which give promise of large tonnages. In 1967, measured and indicated iron ore reserves in Canada were about 33 billion tons. This could be increased considerably when inferred and potential reserves are developed.

Reserves of uranium were rapidly increased during the active exploration period 1950–56 when more than 10,000 occurrences were discovered. One of the Elliot Lake, Ontario, deposits proved to be among the largest in the world. Only a relatively few deposits in Canada reached production, but a number retain economic interest. In 1967 measured and indicated reserves of uranium were 490,000 tons of U_3O_8 contained in ore mineable at a price range of less than $10 a pound of U_3O_8. In additon, there was an equally large volume of lower grade material which could be mined at higher prices. Canada, United States, and South Africa have almost 86 per cent of the known uranium resources in the non-Communist world recoverable at less than $10 a pound of U_3O_8.

TABLE III-4 *Measured and indicated reserves of some Canadian minerals, 1956 and 1967*

Mineral	Unit	Measured and Indicated Reserves 1956	1967
Copper	Millions of tons of contained copper	14.50	19.52
Iron	Millions of tons of ore	(1)	33,106
Lead	Millions of tons of contained lead	9.96	12.15
Nickel	Millions of tons of contained nickel	6.03	7.13
Uranium	Tons of U_3O_8	237,000	490,000
Zinc	Millions of tons of contained zinc	23.23	31.12
Asbestos	Millions of tons of rock	958	1,192
Molybdenum	Tons of molybdenum	1,920	613,810
Potash	Billions of tons of K_2O	5.0	6.4
Sulphur	Thousands of tons in natural gas	20,000 (2)	134,759
Crude oil (3)	Thousands of barrels	2,849,370	8,168,924
Natural gas (3)	Millions of cubic feet	19,621,637	45,682,051
Natural gas liquids (3)	Thousands of barrels	279,934	1,378,868

(1) Not available on same basis
(2) Reported in 1958 by Government of Alberta
(3) Measured reserves only
Sources: 1956 — Royal Commission on Canada's Economic Prospects
1967 — Mineral Resources Branch, Dept. of Energy, Mines and Resources.
Note: These reserve estimates do not include inferred and potential reserves and are therefore not a measure of the country's ultimate mineral reserves.

In 1967, measured and indicated reserves of copper contained in orebodies at producing and near-producing mines totalled 19.5 million tons. Canada's copper reserves were about 10 per cent of the world total. With 31 million tons, Canada led in zinc reserves, having about one quarter of the world total in the measured and indicated categories. In addition, there are large resources in the inferred and potential categories. Measured and indicated reserves of lead were 12 million tons, well over 15 per cent of the world total.

Of the other metallic minerals, Canada has large reserves of ilmenite, molybdenum, cobalt, and columbium, of which molybdenum reserves were particularly actively developed in the 1960's. The occurrences of manganese and chromium are large but too low grade to be mined economically at present. One small but high-grade deposit of tungsten has been brought into production. Metals such as silver, antimony, bismuth, cadmium, and selenium occur in significant quantities in the major base-metal deposits and their recovery proceeds at a rate controlled by the development of the major metals. Canada has always had an important gold industry, third largest in the world, and there are many undeveloped gold occurrences but, under the long-term condition of a fixed selling price and rising costs, there has been little incentive to establish new reserves.

Canada has important reserves of the non-metals, particularly asbestos, potash, and sulphur. It is first in the world in asbestos and potash and second in elemental sulphur.

Proven reserves of crude oil are about 2 per cent of the world total. There is a large potential in unexplored areas, and the reserves of the Athabasca oil sands of Alberta place Canada among the leading countries of the world as a potential source of hydrocarbons. Natural gas measured reserves are second only to those of the United States. Known coal deposits in western Canada are very large but are generally considered in the potential category. With the technological and marketing developments of the late 1960's, large quantities could be reclassified into the measured and indicated categories.

MINERAL SUPPLY AND DEMAND

In 1967 Canada was the world's leading producer of nickel, zinc, and asbestos; the second largest producer of uranium, molybdenum, gypsum, and selenium; the third largest producer of aluminum metal, cobalt, lead, gold, platinum metals, silver, sulphur (all forms), tellurium, and titanium concentrates (ilmenite); fourth largest of cadmium, iron ore, magnesium, and potash; fifth largest of copper and barite; and in the over-all record it ranked high among producing countries for a number of other minerals. The country's world position in mineral supply continues to exceed greatly its standing as a consumer of mineral materials.

Canada stands in third place in the value of production of metals and non-metals and seventh place in the production of mineral fuels. It is the world's third largest diversified mineral producer after the United States and the Soviet Union. On a world scale, the Canadian mineral industry is a major source of supply, and its large resource potential indicates that this position in the world economy will be maintained for many years.

Seven metals, asbestos, petroleum, and natural gas account for four fifths of the value of Canada's exports of minerals in raw and fabricated forms. The changing relationship of the leading exports since 1950 and the over-all growth to 1965 are shown in Table III-5. These data do not include gold exports but do include aluminum metal exports. Figure III-6 shows the principal destinations of Canadian mineral exports in raw and semi-processed forms, and the prominent position of United States as a market for Canadian minerals. Although the U.S. will no doubt continue to be Canada's most important mineral market, the outlook is for a more diversified world market.

TABLE III-5 — *Mineral exports from Canada, by principal commodity* ($ millions)

Commodity	1950	1955	1960	1965(b)
Iron ore	53.8	167.6	228.6	360.8
Aluminum	103.2	211.0	268.1	372.5
Copper	77.2	155.0	206.7	272.6
Lead	38.1	37.2	26.0	64.2
Nickel	105.3	215.2	258.3	397.2
Zinc	58.7	70.6	63.5	141.4
Uranium	—	26.5	263.5	53.7
Asbestos	62.6	94.8	120.1	160.7
Fuels (a)	3.2	41.1	119.3	419.9
All others	67.5	132.8	129.1	539.3
	569.6	1,151.8	1,683.2	2,782.3

Data source: Dominion Bureau of Statistics
(a) Only coal in 1950, with a slight increase in this commodity through the period. Crude petroleum accounted for 70% in 1965 and natural gas most of the remainder of the fuels group.
(b) Not exactly comparable with export values up to 1960 because of a change in D.B.S. trade classification procedures in 1961.

TABLE III-6 — *Exports of principal minerals in relation to production, 1967*

	Unit of Measure	Production	Exports	Exports as % of Production
Copper	s.t.	592,299	404,895	68.3
Iron ore	l.t.	37,787,859	31,406,530	83.1
Lead	s.t.	321,563	258,514	80.4
Nickel	s.t.	250,180	246,525	98.5
Silver	oz.	36,426,079	24,143,093	66.3
Zinc	s.t.	1,086,557	1,033,357	95.1
Asbestos	s.t.	1,400,708	1,342,044	95.8
Gypsum	s.t.	5,119,955	3,896,134	76.1
Potash	s.t.	2,204,231	2,004,504	90.9
Elemen'l sulphur	s.t.	2,322,223	1,773,671	76.4
Coal	s.t.	11,395,754	1,338,353	11.8
Crude oil	bbl.	351,287,792	150,344,567	42.8
Natural gas	Mcf	1,471,597,800	505,164,622	34.3

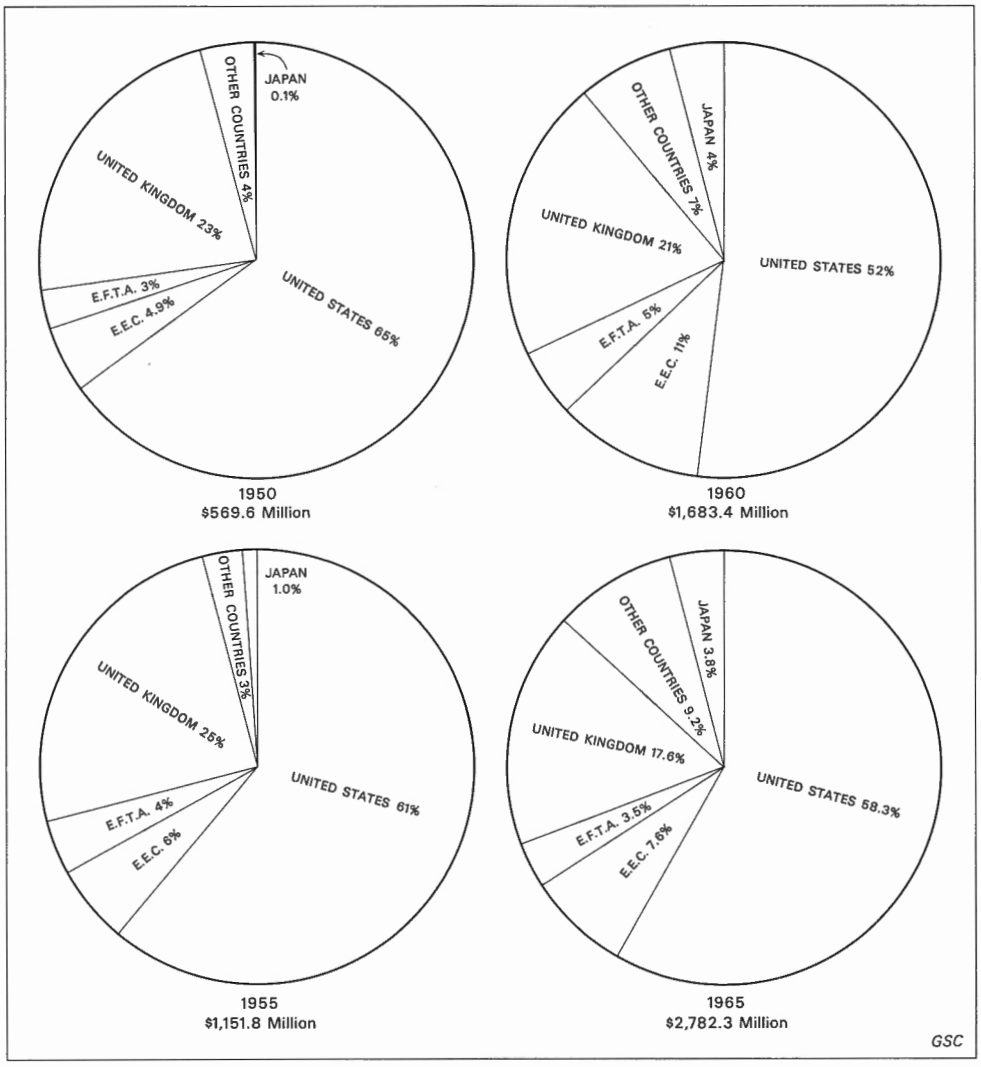


FIGURE III-6
Export market of Canadian minerals in raw and semi-processed form.

Although Canada imports some minerals, in recent years its exports in crude and fabricated forms have exceeded the value of its imports in these forms by as much as $1,500 million annually. Crude oil imports into eastern Canada to supply refineries beyond the economic reach of western Canadian oil have been accounting for two thirds of the value of mineral imports. Table III-6 illustrates the extent of export orientation for leading minerals by relating exports to production. Inasmuch as imports for all minerals, except crude oil and coal, are very small the percentages shown are a measure of the net surpluses available for export after domestic mineral requirements have been met. Canada is deficient in a relatively small number of minerals, the most important being manganese, chromium, tin, bauxite, and phosphate rock.

FUTURE DEVELOPMENTS

Beyond 1970, the opening up of the Athabasca oil sands on a large scale and of potential oil areas in the north and offshore, a resurgence in the demand for uranium, a breakthrough in the markets for iron ore and nonferrous metals in western Europe, markets for coal in Japan and a greater role for Canadian minerals in North American supply could lead to an accelerated growth rate in Canada's mineral output. In any event, mineral production volume in 1980 is likely to be at least double the output of 1965.

Inasmuch as Canada's mineral industry is strongly export-oriented, future developments will be determined in considerable part by the factors that affect international trade. These include matters relative to mineral surpluses throughout the world, trade restrictions, stockpile policies of leading consumers, and national policies governing development and marketing of minerals, particularly in the newly developing countries that are important sources of minerals. Such factors may add greatly to those problems of export market growth that are purely competitive in nature.

Resource development and production problems that arise in Canada stem from widespread distribution of resources in a large country, difficult conditions of terrain

TABLE III-7 *Value of mineral production in Canada by main geological regions, 1967*

	Metals		Industrial Minerals		Fuels		Total All Minerals	
	$ Millions	% of Total	$ Millions	% of Total	$ Millions	% of Total	$ Millions	% of Total
Canadian Shield	1,780.2	78.9	97.9	11.1	—	—	1,878.1	42.5
Appalachian Orogen	147.4	6.5	205.0	23.2	59.2	4.6	411.6	9.3
St. Lawrence Platform	2.6	0.1	283.9	32.1	9.3	0.7	295.8	6.7
Interior Platform	97.9	4.3	215.3	24.4	1,160.1	91.1	1,473.3	33.4
Cordilleran Orogen	228.9	10.2	81.7	9.2	45.0	3.6	355.6	8.1
Total	2,257.0	100.0	883.8	100.0	1,273.6	100.0	4,414.4(p)	100.0

(p) Preliminary

and climate, and a comparatively small population. In turn, these conditions give rise to high costs of operation in remote locations and high transportation costs. However, unlike problems arising from international trade barriers, the problems of the domestic industry can be lessened by increasing the efficiency of operation through the full use of newly devised scientific and technological methods. The extent of the economies that are so achieved will determine the rate of industry growth in so far as this, in turn, is affected by competitive marketing. The mineral production record of recent years is evidence of Canada's ability to develop an efficient producing industry.

The mineral industry in Canada receives strong support and encouragement from government. The various incentives designed to encourage mineral industry growth include attractive mineral rights legislation, tax incentives particularly related to exploration and new mine development, a government role in resource surveys and research, certain limited subsidies and bounties, trade promotion, and government appraisals of industry progress by means of economic studies. Canadian mining legislation provides for security of land tenure, the equitable disposition of mineral rights, and well defined regulations concerning mineral property development. Under the Canadian system of government, all mineral rights — with the exception of those in the federally administered Yukon and Northwest Territories, offshore areas, Indian Reserves, National Parks and some small areas within the provinces — are vested in the ten provinces as provided for in the British North America Act of 1867. The provinces enact legislation to administer and regulate the mineral industry within their boundaries. The stability of the legislative and tax system, and its many incentives, have attracted to the industry the large amounts of capital needed to ensure future growth.

Progress in exploration and production in the past 20 years and the favourable legislative environment in Canada give promise of even greater progress in resource development in the next 20 years. From a production value level of $4.7 billion in 1968, the mineral industry can look forward to an ever increasing role in the Canadian economy and to a major responsibility in helping to meet the rapidly growing mineral requirements of the world.

TABLE III-8 *Value of mineral production in Canada by provinces, 1950–1968* ($ millions)

Province	1950	1955	1960	1965	1968
Ontario	367	584	983	992	1,355
Alberta	136	326	396	794	1,092
Quebec	220	357	446	716	729
Saskatchewan	36	85	212	328	357
British Columbia	139	189	186	280	389
Newfoundland	26	68	87	208	310
Manitoba	32	62	59	183	210
New Brunswick	13	16	17	82	88
Northwest Territories	8	26	27	77	116
Nova Scotia	59	67	66	71	57
Yukon Territory	9	15	13	13	21
Prince Edward Island	—	—	1	1	1
Total	1,045	1,795	2,493	3,745	4,725

IV. Geology of the Canadian Shield

Introduction 44

Superior Province 54

Slave Province 71

Bear Province 77

Churchill Province 85

Southern Province 108

Nain Province 119

Grenville Province 121

Diabase Dyke Swarms 131

Shock Metamorphic Structures 135

Hudson Platform 137

Selected References 145



INTRODUCTION

The Canadian Shield is a great region of Precambrian rocks that occupy nearly half of Canada and form the bedrock for a land area of about 1,864,000 square miles, mainly in Canada but extending south into the Lake Superior and Adirondack regions of the United States of America. Except on the northeastern side facing the Greenland Shield, the Canadian Shield is surrounded by Phanerozoic sedimentary rocks forming a platform cover that is but little disturbed, as it is protected by its virtually stable Precambrian basement. Several inliers of Precambrian rocks protrude through the platform cover and, in the bordering geosynclines, Precambrian rocks are involved in folding in the Appalachian, Cordilleran, and Innuitian Mountain Belts. The rocks of the platform cover may have at one time extended over the whole of the Shield but now remain only as the Hudson Platform, the Foxe Basin, and several scattered small erosional remnants and down-faulted blocks. Although the Shield has been virtually stable since the end of Precambrian time, it has been disturbed by Phanerozoic faults especially along the southern boundary and along the borders of the Boothia Uplift in the north. Phanerozoic faults are almost entirely absent along the more stable western boundary.

An estimate of the chemical composition of the Shield approximates that of granodiorite and it contains more silicon and potassium and less magnesium, calcium, and some other elements than are estimated to be present in the entire crust of the earth (Shaw, *et al.*, 1967). The greater part of the Shield has now been geologically mapped and much of it has been covered by gravity and aeromagnetic surveys. Geological studies have solved many of the local features of stratigraphic succession and orogenic relationships, and have permitted the subdivision of the Shield into structural provinces. Many of the local features have now been dated and correlated by isotopic age methods and an overall scheme of orogenic and time-stratigraphic classification has become possible. The several topics are summarized in the following sections and the summaries are followed by more detailed accounts of the geology of the structural provinces.

Structural Provinces

The subdivision of the Shield into a number of structural provinces and subprovinces is based mainly on overall differences in internal structural trends and style of folding. Boundaries between provinces are drawn where one trend is truncated by another, either along major unconformities or, in their absence, along orogenic fronts. The structural divisions, although based primarily

[1] The Introduction was prepared by C. H. Stockwell. The accounts of Superior, Slave, Bear, Southern, and part of Churchill Provinces were prepared by J. C. McGlynn. The sections on Labrador, Cape Smith, and Belcher Subprovinces are by J. A. Donaldson. R. F. Emslie is responsible for Nain and Grenville Provinces, and B. V. Sanford and A. W. Norris for Hudson Platform. W. F. Fahrig prepared the account of the diabase intrusions and K. L. Currie the shock metamorphic structures. When acknowledging, reference should be made to individual author and page number.



IV

Geology of the Canadian Shield

C.H. Stockwell, J.C. McGlynn, R.F. Emslie, B.V. Sanford, A.W. Norris, J.A. Donaldson, W.F. Fahrig, and K.L. Currie[1]

McDonald fault scarp bordering East Arm Fold Belt of Churchill Province, Northwest Territories.

on geological evidence, have been greatly strengthened by the isotopic dating of orogenies, discussed below. The provinces and subprovinces differ from orogens in that they have broader scope, containing not only orogenically deformed rocks but also virtually undisturbed cover material and a variety of post-orogenic intrusions.

Based mainly on structural criteria, the Shield has been divided into seven structural provinces and several subprovinces (Fig. IV-1). The divisions and their nomenclature are, in the main, a compromise between those recognized by M. E. Wilson (1939), J. E. Gill (1948, 1949), J. T. Wilson (1949), and Farquhar and Russell (1957). Structural trend lines are shown on the Tectonic Map of Canada (Map 1251A) and in more generalized form on Figure IV-2. These show the trend of gneissic and schistose structure, bedding, formational boundaries, and the crest lines of positive magnetic anomalies. Linear trends, for the most part, represent the surface trace of long tight folds whereas curved ones may represent more open folds or, more commonly, tight folds bent by batholithic intrusions; they may also represent complexly refolded folds and include gneissic domes and basins.

Superior Province

Superior Structural Province is characterized by a predominance of easterly trending structures, although northerly trends prevail in the large lobe east of Hudson Bay. In the western part of the province the easterly trends are most pronounced in the Cross Lake, English River, and Quetico Belts. Curved structures are more characteristic of the intervening areas in which batholithic intrusions predominate, such as in the Cat Lake, Wabigoon, and Wawa Belts. The Kapuskasing Belt trends northeasterly across the easterly structures and is flanked by northeast-striking faults. The folded rocks are overlain unconformably by unfolded or gently folded rocks of Mistassini Homocline and Cobalt and Nipigon Plates. Post-orogenic intrusions including bodies of syenite and carbonatite and dykes of diabase, occur within the Superior Province, as well as a few down-faulted blocks of Paleozoic sediments and a circular structure with Permian–Triassic volcanic material.

Slave Province

In Slave Structural Province trends are commonly curved, but on the whole are mainly northward. Post-orogenic features include an intrusion of syenite, gently folded rocks of the Bathurst Plate, and diabase dykes.

Bear Province

The boundary between Bear and Slave Structural Provinces is marked mainly by an unconformity. Rocks of the Wopmay Belt above the unconformity are in part, homoclinal at the base, but elsewhere are folded right up to the contact. The homoclinal and folded rocks are unconformably overlain by rocks of the Coppermine Homocline. Post-orogenic features include dykes of diabase. Closely related to Bear Province are two inliers in the Phanerozoic cover, the one comprising flat-lying material of the Brock Inlier in Interior Platform, and the other, gently folded rocks of the Minto Inlier in Arctic Platform.

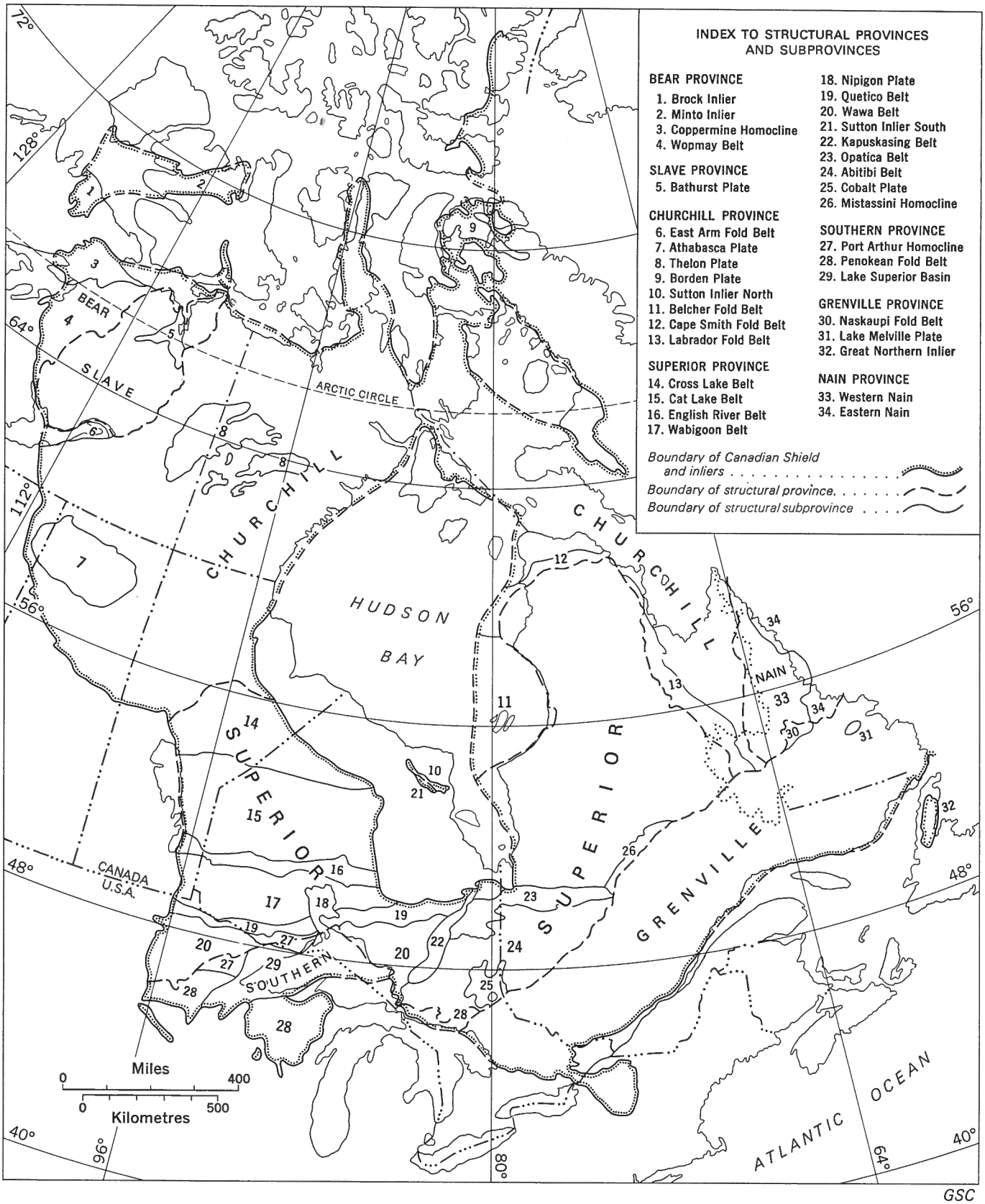


FIGURE IV-1. Structural provinces and subprovinces of the Canadian Shield (by C. H. Stockwell).

Churchill Province

Boundaries between Churchill Structural Province and Superior and Slave Structural Provinces are marked in places by profound unconformities, or, where these are absent, by orogenic fronts. The unconformities are well established at the base of the rocks forming the Labrador, Cape Smith, Belcher, and East Arm Fold Belts. Each is asymmetrical with, characteristically, a narrow homocline at the base, which passes outward into folded rocks with axial planes dipping away from the basement contact and which, still farther out, either grades into more highly metamorphosed rocks of the main body of Churchill Province or is separated from them by a fault. The Sutton Inlier of Hudson Platform comprises a homocline, obviously a continuation of that of the Belcher Fold Belt, lying unconformably on a basement. Boundaries along the east side of Slave Province and along the north part of Superior Province in Manitoba are defined, respectively, by the Thelon Front and the Nelson Front. In each of these localities the orogenic trends on the Churchill side tend to parallel the boundary and to truncate older trends on the other side of the front.

Within the main body of Churchill Province, curved structures predominate, but linear structures are well represented and, in a general way form a large arc, concave to the south, around Hudson Bay so that trends are mainly northeast in the western part, easterly close to Hudson Bay, and southeasterly in the extreme eastern part of the province. A smaller arc, concave to the east, lies in the country between Athabasca Lake and Great Slave Lake. In the interior of the province, the folded rocks are unconformably overlain by flat-lying or only gently folded rocks of the Athabasca, Thelon, and Borden Plates. Near the borders of the province, small bodies of nearly flat lying rocks unconformably overlie folded rocks of the Labrador and East Arm Fold Belts. A circular, possibly cryptovolcanic, structure is found in Athabasca Plate. Postorogenic diabase dykes and a few outliers of Paleozoic and one of Cretaceous sediments also occur within the province.

Southern Province

Southern Structural Province lies partly in Canada and extends into the Lake Superior district of the United States. The boundary between Southern and Superior Structural Provinces is defined, almost everywhere, by an unconformity. The Penokean Fold Belt is characterized by easterly trending folds and the Port Arthur Homocline by gentle dips to the south. These are overlain unconformably by gently folded cover rocks of the Lake Superior Basin.

Nain Province

Nain Structural Province and its division into eastern and western parts are not well known. Its western boundary appears to be an orogenic front for northerly trends in the Nain seem to truncate northwest trends in the Churchill, especially toward the southern part of the boundary. The division into two parts is tentatively made on the basis of differences in isotopic dates and on the presence of anorthosite in the western part. Small areas of nearly flat lying rocks unconformably overlie the folded rocks of both east and west parts.

Grenville Province

Grenville Structural Province lies mainly in Canada but extends into the Adirondack region of the United States. The fault-flanked, Great Northern Inlier in St. Lawrence Platform is closely related to rocks of Grenville Province. In one small area the boundary is an unconformity where homoclinal and folded rocks of the Naskaupi Fold Belt lie on rock of western Nain Province. Elsewhere the boundary is an orogenic front known as the Grenville Front, which extends northeast in a nearly straight line for 1,200 miles truncating along its course the easterly trends of Southern and Superior Provinces, cutting off the Cobalt Plate and the Mistassini Homocline, and continuing northeast to truncate the southeast trend of the Labrador Fold Belt and the northerly trends of Nain Province. For parts of its length the boundary is a fault; in other parts it is a sharp metamorphic front across which the rocks of the several provinces may be traced into more highly metamorphosed and more intricately folded equivalents in Grenville Province. Within the body of the province folding is exceedingly complex and on a small scale; irregular curved structures occur throughout, and numerous gneissic domes and basins have been formed. No large linear structures are apparent but there is an overall northerly preference to the curved trends, especially in the southern and central parts of the province, where they lie at an angle to the long axis of the province as a whole. Post-orogenic features include stocks of syenite along the southeast border, diabase dykes, flat-lying rocks of the Lake Melville Plate, bodies of carbonatite, several outliers of Paleozoic cover, and a circular structure with Permian–Triassic volcanic material.

Precambrian Orogenies

An orogeny is defined as a period of mountain building accompanied by folding. It is an abstract unit of time, and the folding may or may not be accompanied by approximately contemporaneous metamorphism and granitic intrusion. Relative ages between orogens are given by angular unconformities and by the truncation of older structures by younger ones. Isotopic dates confirm the relative ages as determined by geological studies and are invaluable in supplying a scale in years, in outlining the geographical extent of the several orogens, and in providing long range correlation.

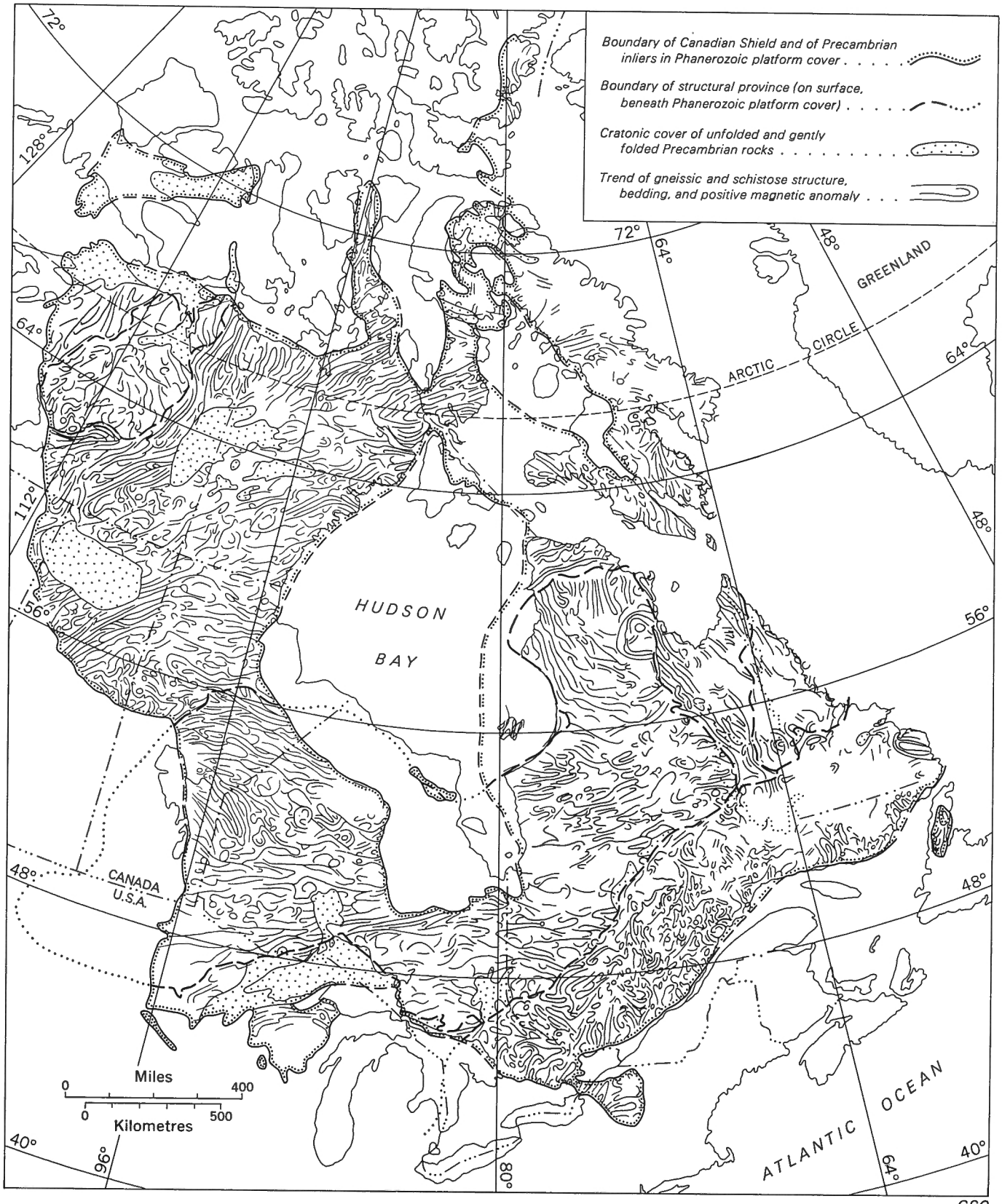


FIGURE IV-2. Structural trends in the Canadian Shield (by C. H. Stockwell).

Because of uncertainties in the interpretation of isotopic dates and because of their inherently large analytical error, it is not possible to define orogenies precisely in absolute time. Accordingly, following the principles recommended by the American Commission on Stratigraphic Nomenclature, the orogenies are defined by concrete rock in type regions rather than by abstract time. Type regions serve as standards for reference and any improvements in their dating or any mistakes made in correlation do not invalidate the units. Four orogenies have been recognized, named, and defined in the Shield. The Kenoran Orogeny is defined as the last important period of widespread folding, metamorphism, and intrusion in Superior Province, which is chosen as the type region. Similarly defined are the Hudsonian Orogeny of Churchill Province as the type region, the Elsonian Orogeny of western Nain Province as the type region, and the Grenvillian Orogeny of Grenville Province as the type region.

Numerous isotopic ages have been determined and are well distributed throughout the Shield. By far the largest number are K–Ar ages, mainly on micas, but a few have been determined by the Rb–Sr whole rock isochron method and by other methods. The K–Ar and Rb–Sr dates have been plotted on a histogram where they are arranged, objectively, according to structural provinces (Fig. IV-3). The K–Ar dates include only those for orogenic minerals whereas the Rb–Sr whole rock dates are from slightly metamorphosed sedimentary and volcanic rocks as well as from gneisses and orogenic intrusions.

The K–Ar mica dates show a considerable spread in each province but, more significantly, the histogram demonstrates that the majority fall in distinct groups indicating the dominant orogeny. In each group the number of determinations falls off rapidly on either side of a peak or mean value, forming a Gaussian probability curve that could result in considerable part from analytical error, which amounts to about ±80 m.y. for the older dates and ±40 m.y. for the younger. Consequently, the time interval from beginning to end of an orogeny is difficult to determine, but, because the analytical errors balance out, the mean values are more significant than the range. The mean K–Ar ages from orogenic micas for the several orogenies in their type regions are as follows: the Kenoran at 2,480 m.y., the Hudsonian at 1,735 m.y., the Elsonian less well defined at 1,370 m.y., and the Grenvillian at 955 m.y. Orogenies in structural provinces, other than those chosen as type regions, are correlated with greater or lesser degree of certainty with the type regions. Thus, the Kenoran Orogeny is well established in Slave Province where most K–Ar orogenic mica dates group around a mean of 2,460 m.y., and it also occurs in Nain Province. Similarly, the Hudsonian Orogeny is reasonably well established in Bear Province where the mean is about 1,785 m.y., and in Southern Province where most K–Ar dates group around a mean of 1,685 m.y. An orogeny in the Great Northern Inlier is correlated with the Grenvillian. These correlations seem valid because the dates used were determined by the same method on the same minerals formed under comparable orogenic conditions. As orogenies are time units, the names used in the type regions

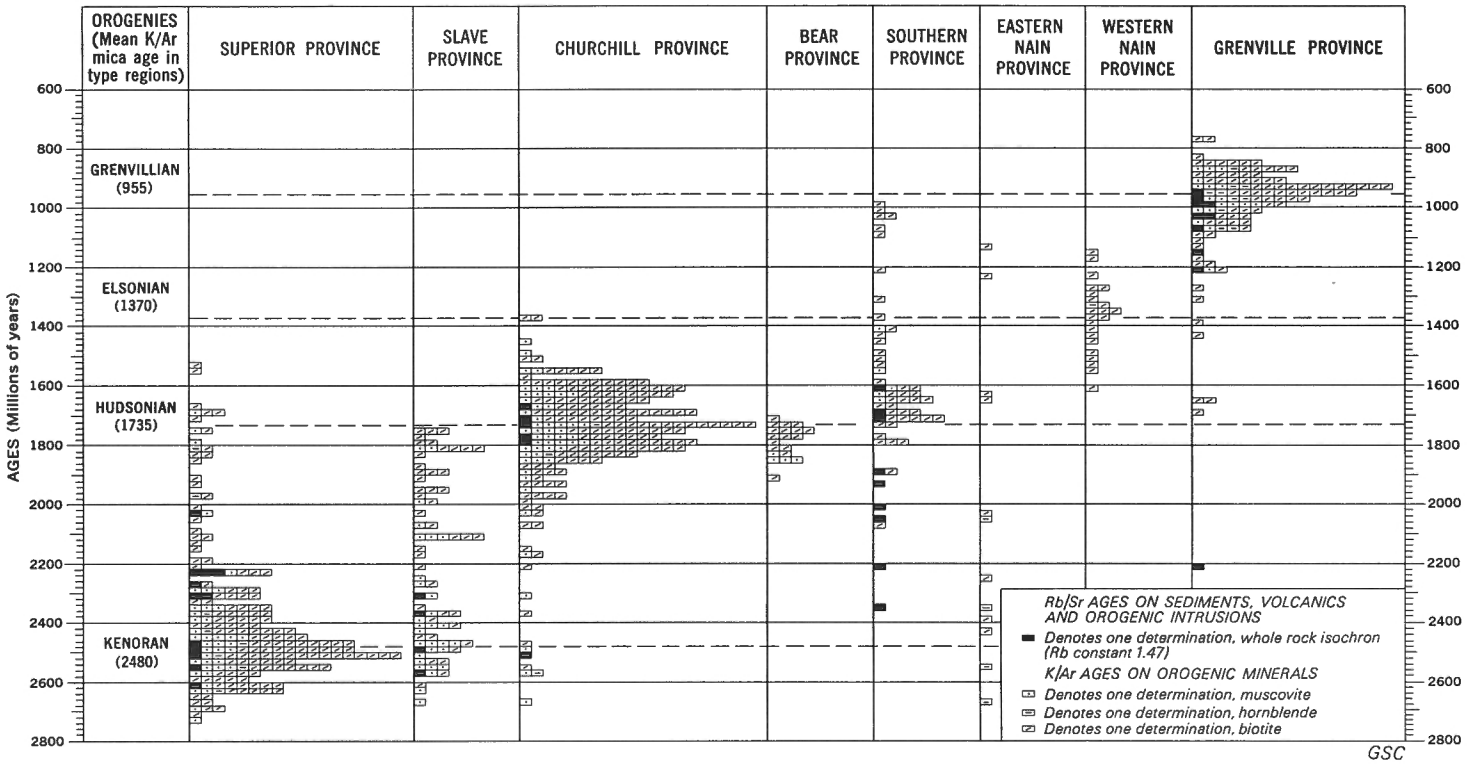


FIGURE IV-3. Histogram of K-Ar and Rb-Sr dates and derived mean K-Ar ages of orogenies of the Canadian Shield (by C. H. Stockwell).

are applied also to contemporaneous orogenies elsewhere, even though widely separated geographically. The orogenies form the basis for the map units on the Tectonic Map of Canada (Map 1251A).

Several of the provinces have been involved, in part, by orogenies older than the dominating one. For example, Superior Province contains profound pre-Kenoran unconformities where volcanics have been folded and intruded by granite, before the unconformable deposition of overlying sediments that were deformed during the Kenoran; but the effects of the Kenoran have been so strong that no distinctly older K–Ar dates have survived. Likewise, Churchill Province contains pre-Hudsonian unconformities, but, as shown by the histogram, a number of dates as old as the Kenoran have survived. Southern Province, dominantly affected by the Hudsonian, yields a few older K–Ar dates around 2,000 m.y. that possibly indicate a pre-Hudsonian orogeny. Grenville Province is known or inferred to contain rocks previously involved in the Kenoran, Hudsonian, and Elsonian Orogenies, all redeformed to some extent during the Grenvillian, but some older dates have survived. Because of the extensive re-working, the Grenvillian Orogeny, although the youngest in the Shield, is the most complex. In each province it is the last dominating orogeny that gives the province its characteristic structure. Several of the provinces have also been modified by local effects of orogenies younger than the dominant one. For example, the Kenoran of Superior Province has been modified locally as indicated by dates as recent as Hudsonian. A similar effect is more pronounced in Slave Province where, in a number of determinations on muscovite–biotite pairs from the same rock sample, the muscovite retains the Kenoran age, but the biotite gives much younger dates, as it is more susceptible to moderate effects of the Hudsonian.

The mean K–Ar ages so far considered are those of the two micas, biotite and muscovite, considered together. When these two minerals are considered separately, the mean muscovite age is found, as a rule, to be 5 to 20 m.y. older than the mean biotite age, which suggests a greater loss of argon from biotite on cooling. Ages obtained from other materials and those determined by other methods agree with the K–Ar orogenic mica ages in some samples but disagree in others. Thus, in Grenville Province a few Pb–Ur and Pb–Th dates for uraninite and thorianite from pegmatite average 960 m.y. as contrasted with coeval biotite dated by the K–Ar method at some 30 m.y. younger; again loss of argon from biotite on cooling is suggested. The 960 m.y. average is in essential agreement with a mean K–Ar muscovite age of 975 m.y. and with a mean hornblende date of 970 m.y. for the province as a whole, suggesting that 960–975 m.y. may be close to the true mean age of the Grenvillian Orogeny. In other provinces the few hornblende dates available likewise average slightly older than mica dates.

In all the provinces, the Rb–Sr whole rock isochron dates, using a Rb decay constant of 1.47, show a considerable spread, some agreeing with the mean mica dates for the orogenies, others being older or younger. Those that are distinctly older are generally interpreted as survival values from older orogens, and thus confirm the geological inferences and some old K–Ar dates, as in Churchill, Southern, and Grenville Provinces. In a small area of the southern part of Grenville Province, however, Rb–Sr dates range from 955 to 1,208 m.y., the older ones only slightly older than the inferred true mean age of the Grenvillian. The older dates could represent early stages of the Grenvillian or, more probably, Elsonian rocks variously up-dated due to effects of the Grenvillian which determined the time at which the Rb–Sr systems last became closed. Ur–Pb dates on zircons from this area range from 1,050 to 1,310 m.y., the oldest approaching the mean K–Ar age for the Elsonian Orogeny. The Rb–Sr dates that are younger than the K–Ar age of the dominant orogeny, as in Superior Province, are difficult to interpret but may have been up-dated by younger events. In Superior and Slave Provinces some Rb–Sr whole rock isochrons are from only slightly metamorphosed volcanic rocks but nevertheless give Kenoran dates. These can be interpreted either as the age of volcanism or, more probably, as the time of orogeny. In the Abitibi Belt of Superior Province modal Pb ages around 3,000 m.y. have been obtained from sulphide deposits and have been interpreted as the age of volcanism (Roscoe, 1965).

Precambrian Time-Stratigraphic Classification

The several Precambrian orogenies, each followed by a profound erosional interval, provide the framework for a natural time-stratigraphic classification of the sedimentary, volcanic, and intrusive rocks. Direct dating of time of deposition of stratified sequences, however desirable, has been largely unsuccessful in the Canadian Shield, probably because most such rocks have been metamorphosed. However, an indirect method of bracketing their time of deposition between maximum and minimum limits has been remarkably successful. The maximum age of a sequence that lies unconformably on a basement is obtained by dating the orogeny that affected the basement; the minimum is given either by dating the orogeny that affected the sequence itself or by obtaining ages from cross-cutting features such as dykes. The maximum may also be determined by dating detrital minerals or boulders in sediments. Intrusive rocks, where retaining their normal ratio of decay products, can be dated directly. The four dated orogenies provide the framework for five time-stratigraphic divisions, boundaries between which are drawn at the close of each orogeny. The close of an orogeny is difficult to define precisely in terms of isotopic age, mainly because of the large analytical error. This difficulty causes no problems in the mapping and classification of unconformity-based sequences in type regions, but, unavoidably at present,

TABLE IV-1 *Precambrian time-stratigraphic classification in relation to orogenies of the Canadian Shield (by C. H. Stockwell)*

Eon	Era	Sub-Era	Orogeny (mean K–Ar mica age, m.y.)
Proterozoic	Hadrynian		
	Helikian	Neohelikian	Grenvillian (955)
		Paleohelikian	Elsonian (1370)
	Aphebian		Hudsonian (1735)
Archean			Kenoran (2480)

leads to some uncertainty in correlation especially with non-orogenic rocks that formed under conditions of cooling very different from those of orogenic minerals. For example, dates obtained from post-orogenic dykes or flows in undeformed cover rocks might, in border-line cases, lead to uncertainties in correlation with the orogenic classification, especially as dates on anorogenic materials also have large analytical errors. Difficulties are encountered also when the basement has been rejuvenated by a younger orogeny and only the younger age can be determined.

The time-stratigraphic classification of Precambrian rocks in relation to the limiting orogenies is depicted in Table IV-1. The distribution of rocks of the several units is shown on the Geological Map of Canada (Map 1250A). For convenience in description, mapping, and discussion each of the time units has been named (Stockwell, 1964). The large first-order units are called eons and comprise the Archean and Proterozoic. These terms have been used in Canada for a long time and are still used in virtually the same sense as originally defined many years ago (Alcock, 1934). This major twofold division of all Precambrian time is useful for comparing or contrasting geological features of each and it also permits a gross time-classification of those rocks for which information is too meagre for more detailed assignment. The Archean Eon includes rocks involved in or emplaced during the Kenoran Orogeny and all older rocks. Profound unconformities are known within the Archean, but, because of strong overprinting during the Kenoran, it has not yet been possible to date the underlying rocks and, consequently, it has not been feasible to correlate them or to subdivide the Archean on a time-stratigraphic basis. The Proterozoic spans the time from the close of the Kenoran Orogeny to the beginning of the Cambrian.

The Proterozoic Eon is subdivided into second-order units called eras, as follows: the Aphebian, spanning the time from the close of the Kenoran to the close of the Hudsonian; the Helikian, from the close of the Hudsonian to the close of the Grenvillian; and the Hadrynian, from the close of the Grenvillian to the beginning of the Cambrian. The Helikian is subdivided into two suberas and the boundary between them is drawn at the close of the Elsonian Orogeny, which results in an older unit called the Paleohelikian and a younger one called the Neohelikian. If any more formal subdivisions are made, such as has been done for the Helikian, it would be highly desirable to choose a stratigraphic or structural boundary that is recognizable in the field and datable by isotopic methods. Informal subdivisions of the several time-stratigraphic units may be found useful. For example, a body of granite emplaced during the Hudsonian Orogeny may be classed informally as late Aphebian. Each of the terms may be used either for concrete time-stratigraphic units or for abstract geologic-time units. The units as defined above have the advantage that no part of the geological column or any gaps in it remain unnamed. The terms system, period, series, and epoch have not been used in the Precambrian classification because it is not yet possible to measure the time from beginning to end of deposition of a sequence of stratified rocks. Instead, the rock-stratigraphic terms, formation, group, and supergroup, have been used. Correlations can be made by lithological similarity, comparison of sequences, and relation to unconformities and intrusions.

Geological Summary

Archean

The Archean was a time of great volcanic activity and of deposition of immense thicknesses of sediments. Almost everywhere the volcanic and sedimentary rocks are, apparently, conformable with one another and many are interbedded, but angular unconformities have been recognized at several localities. Rocks beneath the unconformities include sedimentary, volcanic, and granitic rock, the last only rarely recognized with certainty, and conglomerate above contains pebbles and boulders of these materials. All these rocks were then folded, metamorphosed to some degree, and intruded by granitic rocks during the late Archean Kenoran Orogeny.

The Archean rocks are best preserved in Superior and Slave Provinces but have also been recognized in parts

of Churchill, Grenville, and eastern Nain Provinces. In Superior and Slave Provinces the volcanics are predominantly of basic composition, but more acid varieties are also found. The chemical composition as determined in several representative areas gives proportions of 65 per cent basalt, 25 per cent andesite, and 10 per cent rhyolitic material, and places them in the calc-alkaline suite typical of orogenic environments. In places, the basic volcanics have been metamorphosed to amphibolite and gneiss but more commonly are only slightly altered and primary structures such as pillows, flow lines, and amygdules are well preserved. Ash beds, tuff, and agglomerate are interlayered with the flows, and iron-formation is closely associated with them. Sills of basic rock are common and seem to be genetically related to the flows. The Archean sedimentary rocks are, as a rule, poorly sorted and consist mostly of greywacke, argillaceous material, arkose, and conglomerate but better sorted quartzite and limy beds are found locally. Primary features such as bedding, graded bedding, and crossbedding are well preserved in many places, but, in extensive areas, much of the sedimentary materials have been converted to gneiss and migmatite. During the Kenoran Orogeny, the folding and metamorphism of these materials was accompanied by widespread emplacement of bodies of granitic material leaving between them only isolated remnants and belts of the older rocks. Granulites are found in Superior Province near the Nelson Front, within the Kapuskasing Belt, and are widespread east of Hudson Bay.

In Churchill Province, Archean volcanic, sedimentary, and granitic rocks are less well known as they have been deformed and altered during the Hudsonian Orogeny. Their presence, however, is confirmed locally by isotopic dates of Kenoran age. In places they lie unconformably beneath folded Aphebian sediments and, where effects of the Hudsonian have been moderate, bear a close lithological resemblance to the Archean of Superior and Slave Provinces. Where converted to gneiss, which is widespread, the Archean is difficult to distinguish from the Aphebian. Archean rocks have been traced from Superior Province across the Grenville Front into Grenville Province where they have generally been re-worked to amphibolite and gneiss during the Grenvillian Orogeny. Their distribution beyond the border region is uncertain due to their highly altered condition and inadequate mapping and study. Kenoran dates on gneisses and granitic rocks of eastern Nain Province indicate the presence of Archean rocks in that area.

Proterozoic

Sedimentary rocks of the Proterozoic show a marked change in depositional environment for they are, on the whole, much better sorted and contain much more limestone than those of the Archean. The better sorting is evidenced mainly by the greater abundance of quartzite. As a result, the rocks are less monotonous and rock-stratigraphic units of formational rank have been recognized and mapped to a much greater extent than in the Archean. Along with volcanic materials they have been deposited as cratonic cover in some places as well as in geosynclines. They vary from almost undisturbed and unmetamorphosed types to highly folded and greatly metamorphosed varieties.

Aphebian

Rocks of Aphebian age occur both as cratonic cover and as geosynclinal deposits. The cover rocks lie unconformably on Archean rocks of Superior and Slave Provinces and, protected by their almost stable basement, are only gently folded and are, as a rule, unmetamorphosed. They are represented by the Cobalt Plate of conglomerate, arkose, greywacke, and shale, and by the Bathurst Plate and Mistassini Homocline of conglomerate, sandstone, greywacke, argillite, shale, and dolomite. Geosynclinal deposits of Aphebian age were deposited in regions bordering Superior and Slave Provinces, where typically they form the long narrow geosynclines of Labrador, Cape Smith, Belcher, and East Arm Fold Belts of Churchill Province, and the broader geosynclines of Bear and Southern Provinces. The deposits include conglomerate, arkose, greywacke, sandstone, siltstone, quartzite, limestone, dolomite, argillite, shale, greywacke, chert, iron-formation, and their metamorphic equivalents. Red beds are common and many of the limestones are stromatolitic. Volcanic rocks, mainly basaltic, are common in some parts of the succession and basic sills are found in association with both the volcanics and the sediments. In places, these rocks form virtually undisturbed homoclines adjacent to their Archean basement, but farther out have been folded, intruded by granite rocks, and converted to gneisses during the Hudsonian Orogeny. The Sudbury norite–micropegmatite of Southern Province was also emplaced during the late Aphebian.

Folded Aphebian rocks are also found in the interior part of Churchill Province where they have been converted to gneisses much more commonly than in the border regions, but some areas of less metamorphosed Aphebian rocks, such as conglomerate quartzite, limestone, dolomite, local iron-formation, and volcanics are well preserved, forming irregular areas rather than well-defined belts. Many of the gneisses are difficult to distinguish from rejuvenated Archean gneisses and are classed as Aphebian or Archean. All were involved in the Hudsonian Orogeny accompanied by widespread granitic intrusions. Aphebian rocks are also found in Grenville Province where they comprise marble, quartzite, iron-formation, volcanics, amphibolite, aluminous paragneisses, and migmatite. Structures are complex as a result of reworking during the Grenvillian Orogeny. These rocks have been traced into the province for some miles southwest of their less metamorphosed equivalents of the Labrador Fold Belt and eastward from those of Southern Province. In the intervening central part of Grenville Province, isolated remnants of lithologically similar material are probably also

of Aphebian age. Post-orogenic features of Aphebian age include a body of alkaline syenite cutting Archean rocks in Slave Province, carbonatite complexes in or near the Kapuskasing Belt (Gittins, *et al.*, 1967), and numerous diabase dykes in Slave and Superior Provinces.

Helikian

Rocks of undivided Helikian age consist predominantly of little deformed cratonic cover on Hudsonian Orogens. In Churchill Province they include the Athabasca Plate of conglomerate, sandstone, and shale, and the Borden Plate of conglomerate, sandstone, quartzite, siltstone, shale, limestone, dolomite, basalt, and andesite. Small areas of Helikian conglomerate, arkose, and sandstone also overlie the Hudsonian Orogen of East Arm Fold Belt. In the lower part of the succession of Coppermine Homocline and Lake Superior Basin, they include conglomerate sandstone, shale, and dolomite overlain by thick flood basalt interbedded with sandstone and shale; these rocks overlap onto the Kenoran Orogen of Superior Province to form Nipigon Plate. Post-orogenic diabase dykes of Helikian age are found in all provinces except Grenville.

Paleohelikian. In western Nain Province intrusive rocks accompanying the late Paleohelikian Elsonian Orogeny include bodies of anorthosite and related basic and granitic intrusions. Similar intrusions, probably also of that age, are scattered throughout large parts of Grenville Province where they have been moderately altered during the Grenvillian Orogeny. In Grenville Province the Hastings Group of conglomerate, greywacke, argillite, dolomite, limestone, and schist may also be Paleohelikian because it probably overlies the Grenville Group and granitic rocks unconformably (M. E. Wilson, 1965), and is cut by basic intrusions that were probably emplaced during the Elsonian Orogeny. Elsewhere in the Shield, and unaffected by the Elsonian or Grenvillian Orogeny, rocks of Paleohelikian age form little deformed cover on Hudsonian Orogens. In Churchill Province they include formations of the Thelon Plate comprising conglomerate, sandstone, siltstone, limestone, and alkalic volcanic rocks. The lower formation of this succession is cut by a post-orogenic body of Paleohelikian syenite. Also of this age are small areas of cover rocks on Hudsonian Orogens elsewhere in the main part of Churchill Province and in the Labrador Fold Belt.

Neohelikian. In Naskaupi Fold Belt, conglomerate, quartzite, slate, argillite, and volcanic rocks are of Neohelikian age as they lie unconformably on the Elsonian Orogen and were folded during the Grenvillian. The Wakeham Group, which also lies within Grenville Province, may be Neohelikian as it is probably younger than anorthosite, apparently lies unconformably on gneisses, and was folded during the Grenvillian. Some granites of Grenville Province were emplaced during the late Neohelikian Grenvillian Orogeny. Neohelikian rocks also occur outside the province. For example, a cover of Neohelikian sedimentary and volcanic rocks lies on the Archean of eastern Nain Province. Of Neohelikian age also, but apparently somewhat older than the Grenvillian Orogeny, are post-orogenic intrusions of the Muskox Complex of basic rocks in the Hudsonian Orogen of Bear Province, several carbonatite complexes, and the Caldwell syenite in the Kenoran Orogen of Superior Province, and the Logan sills of Southern Province.

Hadrynian

The Grenvillian Orogen is overlain unconformably by virtually unfolded conglomerate and sandstone of Lake Melville Plate which could be either Hadrynian or younger. Post-orogenic intrusions of Hadrynian age include diabase dykes in Grenville, Bear, and Churchill Provinces and syenitic stocks in Grenville Province. Also of Hadrynian age are sandstone, shale, and dolomite of at least the uppermost part of the succession of Coppermine Homocline, Brock Inlier, and Lake Superior Basin, and similar rocks together with basalt of Minto Inlier. Beyond the limits of the Canadian Shield the Hadrynian is well represented in the Appalachian and Cordilleran fold belts.

Geophysical Surveys

Gravity

The greater part of the Shield has been covered by reconnaissance gravity surveys (Map 1257A). Throughout much of the area covered the gravity pattern as shown by contours consists of irregular non-linear areas ranging from 0 to —70 milligals. This general rather featureless pattern contrasts with some definite linear trends. These are closely related to major geological features, especially to boundaries between structural provinces. Thus, a conspicuous low anomaly from —70 to —110 milligals closely parallels the Grenville Front for some 500 miles, lying for most of its length on the Grenville Province side. Also, the Nelson Front is closely paralleled, on the Kenoran side, by a high anomaly from —30 to 0 milligals that crosses the easterly trends of underlying rocks and continues beneath adjoining platform cover for a total length of 800 miles. This high anomaly, for part of its length, is paralleled on either side by complementary lows. Of less prominence, is a low anomaly along the Kenoran side of part of Thelon Front with a complementary high on the Hudsonian side. The contact between the Kenoran Orogen of Superior Province and the overlying deposits of Labrador and Cape Smith Fold Belts is closely followed by a low of —30 to —60 milligals and, within the Cape Smith Fold Belt, this trend is paralleled by other high and low anomalies which follow the geological structures.

In the interior of Superior Province high anomalies from —30 to 0 milligals follow the Kapuskasing Belt, and a less prominent one trends in part along English River

Belt. In Bear Province another linear gravity high from −30 to +20 milligals passes through and beyond the Muskox Complex.

Magnetism

Much of the southern and western parts of the Shield have been covered by total intensity, low altitude aeromagnetic surveys and the results are published as detailed contoured maps. For larger areas the regional magnetic gradient, due to deep-seated processes, has been removed leaving a pattern resulting from the relatively shallow magnetic part of the crust (Map 1255A). On this map the pattern of alternating areas and belts of low and high intensity reflect, for the most part and in a gross way, the pattern of geological structures. For example the dominantly northern trends in Slave Province and the easterly trends in the southern part of Superior Province are well shown. The northeast to east trends in the southwest part of Churchill Province are particularly well emphasized, and also the dominantly small-scale, irregular, contorted pattern reflecting the extremely complex structures in Grenville Province. The magnetic pattern also brings out opposing structures along orogenic fronts. A good example is a broad, featureless, magnetic low within Grenville Province, close to the Grenville Front and trending parallel with it for 250 miles. This low magnetic anomaly coincides with part of the long low gravity anomaly described previously. A series of magnetic highs, closer to the front and parallel with it, truncates the easterly magnetic trends of the adjoining Superior and Southern Provinces. Similarly, along the Nelson Front, the easterly magnetic trends of Superior Province are sharply cut off by a magnetic low and a complementary high that parallel the front on the Churchill side. Within Churchill Province, the McDonald fault, which marks the south boundary of East Arm Fold Belt, lies along the northern edge of a pronounced magnetic high. Within the Kapuskasing Belt, northerly trending magnetic highs cross the easterly trends of adjoining parts of Superior Province, and correspond in part with the gravity highs already mentioned. Likewise, the English River Belt, for much of its length, is marked by a magnetic high which coincides partly with a gravity high. The magnetic pattern over sedimentary cover depicts the structural trends of the basement. Dykes of diabase show very distinctly on the detailed maps as positive, and locally as negative magnetic anomalies, and numerous carbonatite and other basic intrusions stand out from their surroundings as high anomalies, as do also magnetic iron-formation and some basic lava flows.

Seismology

Crustal thickness beneath the exposed Shield and platform areas is not well known, but local seismic studies with chemical explosions indicate that the crust has a general thickness of the order of 35 to 45 kilometres, that it thickens to 50 kilometres beneath Lake Superior and Williston Basins, and that it thins to 25 kilometres beneath Hudson Platform.

SUPERIOR PROVINCE

Tectonic Summary

The oldest rocks in Superior Province are volcanic and sedimentary rocks of Archean age. These rocks were formed in a number of easterly trending eugeosynclines that were possibly of different ages and tectonically independent. These rocks are involved in the Kenoran Orogeny during which they were folded, faulted, metamorphosed, and intruded by granitic rocks. Many of the major shear zones in the volcanic belts may have been initiated during late phases of the orogeny. Archean sequences may also have been affected by older orogenies, the effects of which were masked by the last event. The mean K–Ar age of Kenoran granitic rocks is about 2,480 m.y.

After the Kenoran Orogeny the rocks were deeply eroded and Aphebian and Helikian sediments were deposited in a number of basins. At different times the Archean rocks were intruded by diabase dyke swarms, cut by major faults, and intruded by alkaline intrusions. At least nine swarms of diabase dykes occur in Superior Province. The oldest is about 2,200 m.y. and younger swarms are about 2,000, 1,900, 1,400, 1,200, and 1,000 m.y. Many sets of steep-dipping faults marked by zones of breccia, shearing, or mylonite cut Archean rocks. The orientation of these faults vary and their relative ages are usually unknown, but several periods of faulting occurred that may be related to the diabase intrusions and to deposition in the Aphebian sedimentary basins. These younger deformations may have occurred during the Hudsonian and Grenvillian Orogenies and formation of the Lake Superior Basin.

Volcanism and Sedimentation

The Archean sedimentary and volcanic rocks in the southern and western parts of the province occur in a series of long, easterly trending belts, separated by areas underlain by granitic rocks. The volcanic and sedimentary rocks have various stratigraphic relationships throughout the province and even in single belts. In most areas where

structural data are good, the volcanic rocks are the older and are conformably overlain by the sediments. However, in a few areas, the sediments occur at the base of the exposed sequence. In some belts the normal succession of volcanic and sedimentary rocks is followed by more volcanics and sometimes by a second volcanic-sedimentary sequence. Frequently the volcanic rocks, particularly the acidic types, appear to pass along strike through a transition zone of interfingering sediments and volcanics into a sedimentary sequence. There seems to be a gradual transition from volcanism to sedimentation without any significant time break, and at times the sedimentation and volcanism were proceeding simultaneously in different parts of the geosynclines (Fig. IV-4). Local unconformities are known but they may not be of regional extent. An exception is the angular unconformity between the Timiskaming rocks in the Abitibi Belt and underlying volcanic-sedimentary assemblage.

Thicknesses of Archean complexes vary between belts and within single belts (Fig. IV-5). Maximum thicknesses of 25,000 feet are common and as much as 50,000 feet has been reported. Volcanic rocks form the greater part of thick successions and the sediments are thickest where the volcanic part is thin (Moorhouse, 1965). Goodwin (1965) estimates that the Archean comprises about 80 per cent volcanic and 10 per cent sedimentary rocks. As these rocks are granitized no figure represents the true thickness, and the thicknesses of the sediments in particular may be underestimated by a large factor.

The Archean volcanic sequences have much in common. They comprise varying proportions of basalt, intermediate to acidic lavas, tuffs, and breccias with minor amounts of sediment. Basic lavas predominate in the lower two-thirds of most sequences while acidic flows, if present, are found in the upper parts. The intermediate to acidic phase is more local in distribution, occurring as thick sequences in scattered areas along individual volcanic belts that probably represent strato-volcanoes built on basic lava. In some areas this volcanic cycle is repeated or partly repeated producing variations in the stratigraphy of the belts. Sedimentation often appears to be contemporaneous with the buildup of the acidic volcanic rocks, erosion of which may then provide part of the detritus. A resurgence of volcanism after sedimentation has started usually results in narrow bands of basic or, more commonly, acid volcanic rocks near the base of the sediments. The lavas of Superior Province belong to the calc-alkaline suite; however, the basalts appear to be derived from a tholeiitic magma type (Wilson, *et al.*, 1965; Goodwin, 1965). Sills, dykes, and irregularly shaped intrusions of gabbro and diorite are present within the lavas, usually in the basic parts. Acidic, often porphyritic, fine-grained stocks occur in or near the acidic volcanic rocks. Both types of hypabyssal intrusion are probably contemporaneous with nearby volcanic activity. Ultrabasic rocks occur as parts of differentiated gabbroic masses or as small discrete masses that may lie in fault zones and whose ages relative to the surrounding granitic rocks are often not known. The constituent minerals are normally altered to serpentine. Little is known about these rocks but it may be that all of the ultramafic masses have been intruded along major fault zones, usually into volcanic rocks during an early stage of the tectonic cycle.

Typical Archean sediments in the various belts are interbedded greywacke and shale, whose composition varies considerably so that feldspathic, lithic, and quartz greywackes and subgreywackes are common. The shale or argillite is normally interlayered with the greywacke as the upper part of graded beds or as thick homogeneous units. Common primary structures, in addition to graded bedding, are flame structures, groove casts, small scale crossbedding, and parallel and convolute laminations.

Conglomerates occur within the greywacke sequences commonly near their base just above volcanic rocks. These

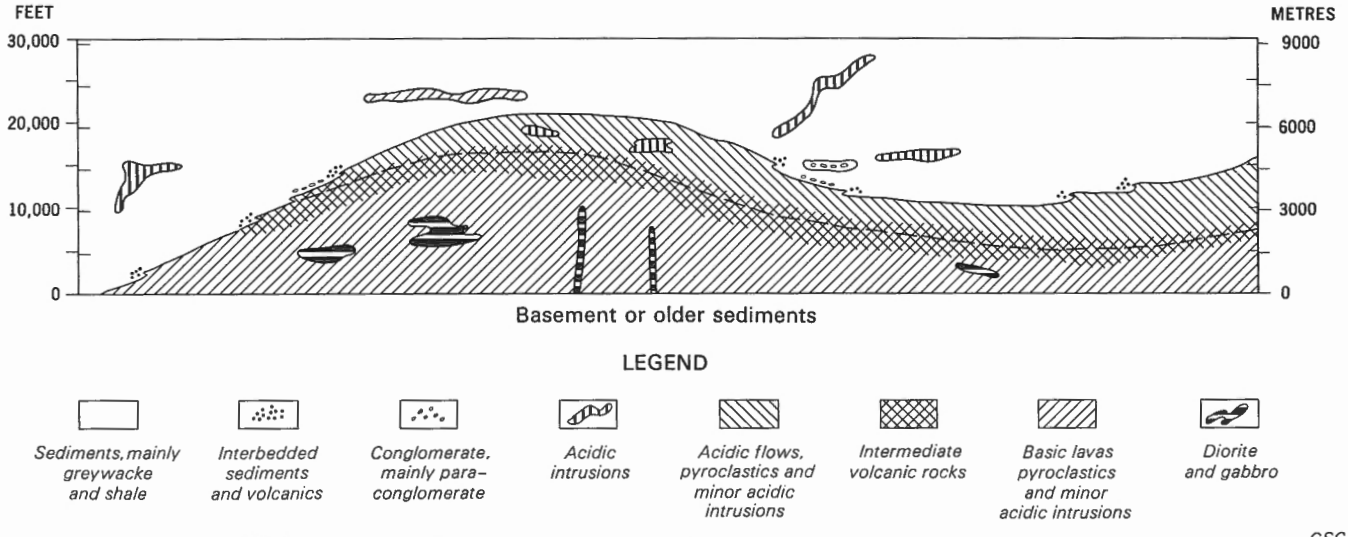


FIGURE IV-4. Diagrammatic section of an Archean volcanic-sedimentary complex (by J. C. McGlynn).

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units may be as much as several thousand feet thick, but often wedge out rapidly along strike. Such conglomerates are usually massive and poorly sorted and comprise phenoclasts that vary from granule- to boulder-size in a greywacke matrix. The phenoclasts are subangular to rounded, consisting of volcanic rock and quartz with some porphyritic rocks and minor amounts of granitic rocks, greywacke, and argillite. In the past, workers have termed as "granite" clasts of acidic porphyritic intrusions related in time to the volcanic rocks, and thereby have given the impression that granitic boulders form a large proportion of the clasts in Archean conglomerates. Donaldson and Jackson (1965) show that boulders of granitic rocks and some gneisses generally occur only in small amounts, less than 10 per cent.

Other sedimentary lithologies include iron-formation, relatively clean quartzite, and limy rocks. The iron-formations occur in beds that range in thickness from a few inches to several hundred feet and extend along strike as much as tens of miles. Typically they consist of thin chert layers interbedded with magnetite- and hematite-rich bands (Chapter V, *Frontispiece*). Some beds contain iron-rich silicates, sulphides, and iron carbonates, which commonly occur with the acidic tuffaceous phases of volcanic sequences, below layers of basic flows, or within other sediments just above the volcanic sequences. Because of the close spatial relationship with volcanic rocks, a common inference is that iron and silica in these sediments were produced during the volcanic activity. Iron-formations are often associated with sulphide-rich black shale. Limestones in the Archean are relatively rare and typically are thin beds associated with iron-formations or with black shales and thin quartzite beds usually in volcanic rocks, or are beds intercalated with tuffs. Donaldson and Jackson (1965) describe Archean sandstone with a carbonate cement that grades to quartzose carbonate. Archean sediments of Superior Province have most of the characteristics of flysch facies (Pettijohn, 1943), as the primary structures are compatible with deposition by turbidity currents in an eugeosynclinal environment.

There has been much speculation about the nature of the source of Archean sediments. It has been suggested

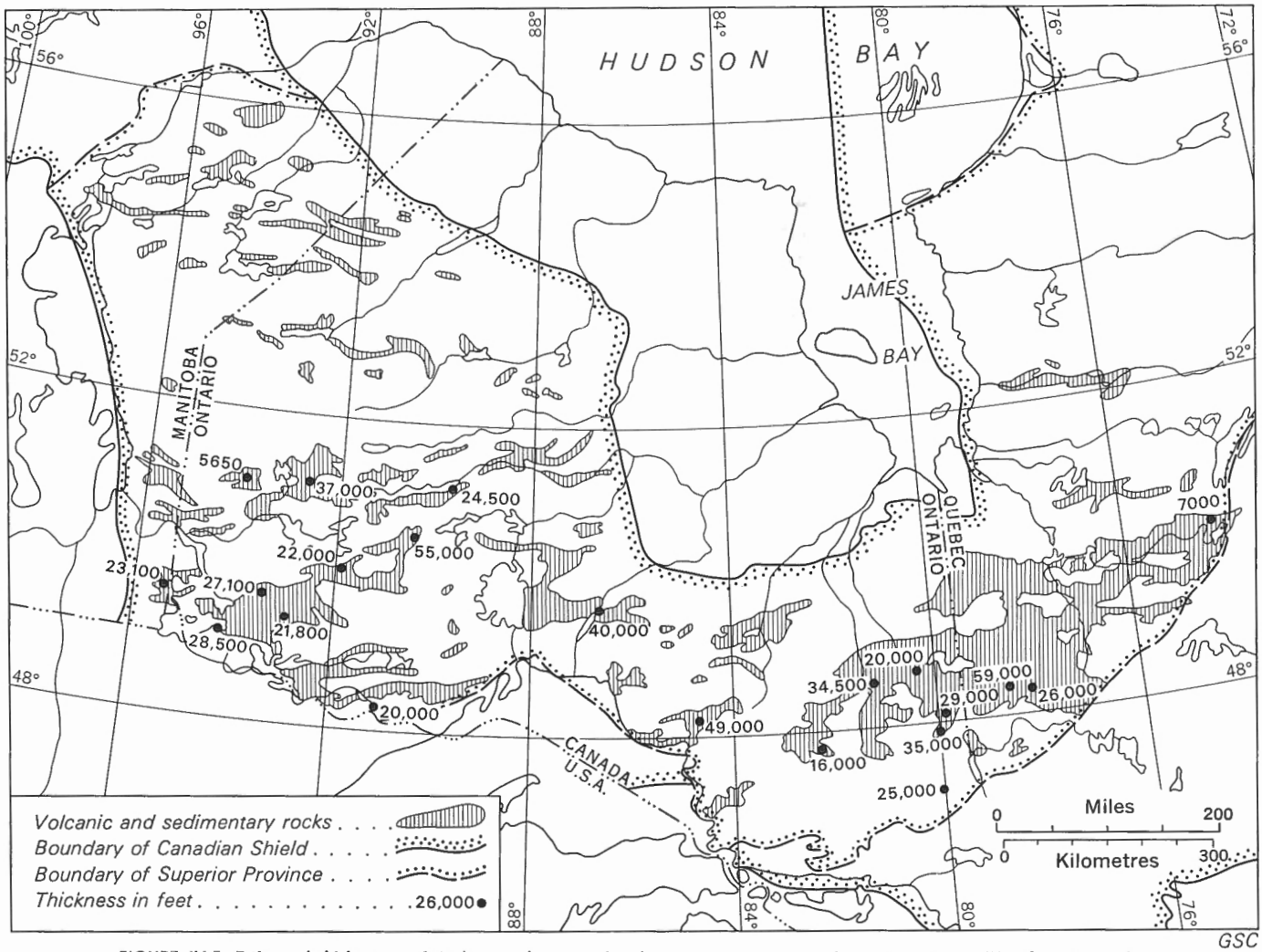

FIGURE IV-5. Estimated thicknesses of Archean volcanic and sedimentary sequences in Superior Province (Moorhouse, 1965).

that the bulk were derived from volcanic rocks. Certainly more than half the clasts in most conglomerates are volcanic material that can be related to nearby volcanic rocks. It is usually suggested that rapid erosion of the acidic piles, which was in part subaerial, supplied much of the quartz and feldspar of the greywacke. Donaldson and Jackson (1965) express doubt that volcanic rocks can be an adequate source for the quartz in the quartz greywackes of the North Spirit Lake area. Although quartz is a minor constituent of the basic lavas, in most acidic volcanic rocks it occurs in smaller grains than those found in the sediments. Furthermore, granitic and granite gneiss pebbles in the conglomerates must have had a source other than the volcanic sequences. In any single volcanic belt the acidic lavas and pyroclastics are scattered and form a relatively small part of the total volume of lava. They are a source, but only for neighbouring parts of the basin. The volcanic rocks flanking the sedimentary basins probably provided many of the coarse clasts in the conglomerates and part of the quartz and feldspar in the sands, and granitic and quartz boulders and some of the sand detritus were probably supplied by areas underlain by plutonic rocks. As the composition of greywacke type and the clasts in the conglomerates vary throughout the region the relative importance of the two sources also varies.

Whether the various belts represent the remnants of separate geosynclines of different age or of one large complex geosyncline, since distorted by orogeny, is unknown. It seems more likely, however, that the volcanism was localized along a number of easterly trending zones of weakness and that volcanic rocks never covered the whole region as a continuous sheet. Because of their thickness and resistance to granitization the rocks have survived subsequent orogeny and may have influenced the tectonic trends of that orogeny. As sedimentation also took place within and about these belts, source areas must have existed between some of the volcanic belts. These areas of ancient crust have since been modified or destroyed during subsequent orogenies as there is little direct geological evidence of the existence of basement. The granitic rocks along the margins of belts of Archean rocks invariably are intrusive into them. Archean rocks along the gneiss belts are either separated from the gneisses by faults or grade into them through zones of increasing metamorphism. One possible exception is reported from the Cross Lake area in Manitoba (Rousell, 1965) where sediments of the Cross Lake Group, said to conformably overlie basic lavas, rest unconformably on granodiorite gneiss that is, therefore, interpreted as basement to both volcanic and sedimentary rocks. Timiskaming rocks in the Kirkland Lake area are separated from older Archean volcanics and sediments by an angular unconformity. These sediments and associated potash-rich trachytic lavas which contribute debris to the sediments differ from normal Archean successions. Timiskaming rocks may be younger than normal Archean sediments, having formed in a special local environment not found elsewhere in the province.

Kenoran Orogeny

All the Archean rocks in Superior Province have been affected by the Kenoran Orogeny which involves a sequence of events, possibly complex but to a large extent undetermined, that include folding, shearing, metamorphism, and intrusion of granitic rocks. Archean rocks in most parts of the province are folded about easterly trending axes. The folds plunge at moderate angles either east or west. Some folds are modified by late cross-folds whose axial planes dip steeply and strike northwest, north, or northeast. The younger folds have fairly steep plunges. Thick bands of volcanic rocks commonly are broadly folded or occur as homoclinal successions; thin volcanic bands and sediments are more complexly folded. Chlorite–carbonate shear zones in thick volcanic sequences near or along volcanic-sedimentary contacts in many places are nearly parallel with the regional trend of fold axes and locally occur along the axial zones of folds. They may form during the folding and cut granitic rocks, but in some areas they appear to be older than late tectonic intrusions.

Granitic rocks of various compositions intrude Archean rocks and large volumes of the sediments, and some volcanic rocks have been converted to granitic gneisses by processes of granitization. Highly metamorphosed and granitized rocks are extensively developed in several long easterly trending belts, the Quetico and English River Subprovinces of western Ontario. Other unnamed belts are located in New Quebec and along the southern margin of the Noranda–Kirkland Lake volcanic belt. The zones of gneisses are locally separated from older rocks by major fault zones. In most areas, however, recognized sedimentary or volcanic rocks can be traced through increasing zones of metamorphism and degrees of granitization into these gneisses. The gneisses are cut by batholiths of more massive granitic rocks. The structural trends, probably foliation, in these belts tend to be regular and linear, striking in an easterly direction. These rocks appear to be Archean rocks transformed into gneisses in deeply depressed parts of the geosyncline. Relatively massive granitic masses are higher level intrusions. The areas between the Archean rocks and gneiss zones are underlain in large part by granitic rocks with curvilinear structural trends (Stockwell, 1964). The granitic rocks in these areas appear to be syntectonic batholiths of slightly gneissic granitic rocks separated by zones of older rocks or granitic gneisses and migmatites that are cut by late tectonic, massive, and slightly more acidic granitic instrusions. Similar rocks are intruded into the volcanic-sedimentary belts. Age determinations from samples of all these granitic rock types give ages that range from about 2,300 to 2,600 m.y. and average 2,490 m.y. Compositions of the various granitic masses vary but granodiorite, quartz diorite, and quartz monzonite appear to be most abundant with lesser amounts of granite and syenite. The potash-rich rocks are usually massive, younger than the other granites, and late tectonic.

Pegmatites and pegmatitic granites are common components of the gneiss belts, but pegmatite dyke swarms, concentrated around related granites, are found in several widely scattered areas in the province. The related granites are usually massive, late tectonic, albite granites or potash-rich, muscovite granites or granodiorites. Several of these pegmatite areas occur in gneiss zones. Pegmatites may or may not be internally zoned and are commonly zoned away from the related granite. Some contain spodumene, beryl, or tantalite–columbite.

The Archean rocks are always metamorphosed to some degree, varying from greenschist to amphibolite facies. Common mineral assemblages in greywacke and shales in the higher grade facies include plagioclase–biotite–garnet or andalusite, cordierite, staurolite, biotite and plagioclase and locally sillimanite. In lower grade facies, chlorite, sericite, biotite, and plagioclase are the essential minerals. In basic volcanic rocks blue-green hornblende, plagioclase, and locally garnet occur in the amphibolite facies, and in the less metamorphosed rocks chlorite or actinolite, plagioclase, and commonly epidote. In the gneiss zones metamorphism reaches amphibolite and locally granulite facies and in northern Quebec large areas are underlain by rocks of the granulite facies. As metamorphism increases primary textures and most structure may be progressively destroyed.

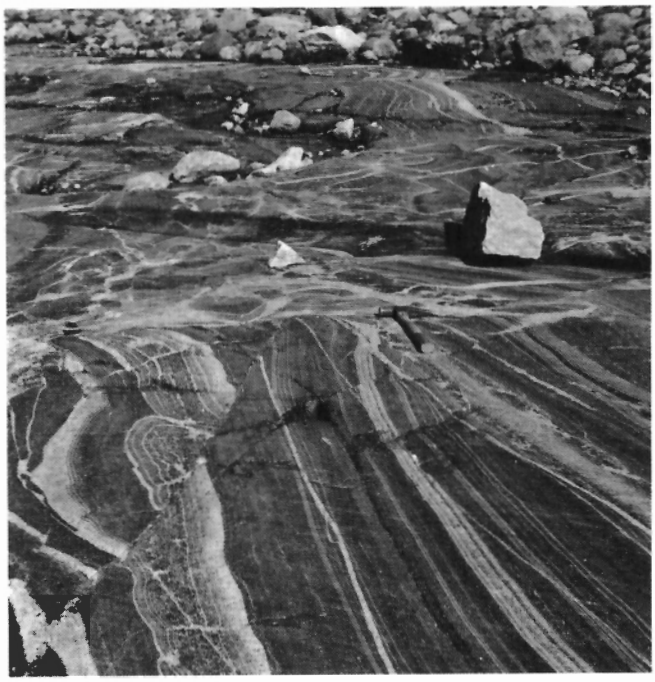

PLATE IV-1. Partly migmatized Archean greywacke, Superior Province, Povungnituk, Quebec.

Northeastern Superior Province

The geology of northeastern Superior Province differs from that of the rest of the province (Eade, 1966a; Stevenson, 1963, 1965). Recognizable Archean pre-granitic rocks are much less abundant; large areas are underlain by granitic gneisses and by metamorphic rocks of the granulite facies. Structural trends in the southern half are similar to those in the remaining part of Superior Province, but in the northern part strike almost at right angles. Lack of recognizable sedimentary volcanic rocks, high grade of metamorphism, and intense granitization all suggest that a deeper level of the orogen or crust is exposed in the northern part, which may also account for the change of structural style.

The oldest known rocks are basic lavas, intermediate and acidic lavas and pyroclastics, small ultramafic intrusions, some greywacke, quartzite conglomerate, and iron-formation. These rocks are best developed east of James Bay, occurring in two easterly trending belts separated by a wide zone of highly metamorphosed sediments, migmatites, granitized rocks, and granitic gneisses. Narrow bands of highly metamorphosed sedimentary and volcanic rocks that are possibly equivalents are scattered throughout the granitic rocks in the northernmost part of the province. The zone of gneisses between the two southern belts of Archean rocks apparently originally consisted largely of sedimentary rocks that have been metamorphosed, granitized, migmatized, and then intruded by granodiorites, presumably during the Kenoran Orogeny. Archean rocks also grade into granitic gneisses through zones of migmatites and mixed gneisses. These gneisses are also intruded by massive biotite and/or hornblende granites, granodiorites, and quartz diorites that are probably late tectonic.

In the southeastern and northern parts of the region hornblende-granulite facies rocks underlie large areas. They are coarse-grained, rusty or white weathering, faintly banded gneisses grading to strongly banded, rusty weathering paragneiss characterized by fine-grained, biotite-rich metasedimentary material. Greenish plagioclase with a greasy lustre, blue or dark grey quartz, hypersthene or augite, hornblende, and biotite are common constituents. Garnet is locally present. Because of the banding of these gneisses and the presence of biotite-rich bands, Eade (1966a) suggests that they are derived from sedimentary rocks, possibly of the greywacke type. Rocks of the amphibolite facies grade into the rocks above and contain similarly oriented minor structures. Quartz, plagioclase, biotite, hornblende, epidote, and usually microcline are characteristic minerals in the granitized varieties of this facies. The structure of the rocks in the area is not known in detail. South of about latitude 57 degrees, Archean rocks and their metamorphosed and granitized equivalents occur in easterly to east-northeasterly trending folds. These structures probably are modified by cross-folds along northwest and northeast axes. North of latitude 57 degrees the regional trend is northwest to north-northwest. This area is essentially underlain by granitic gneisses with few rem-

nants of recognizable volcanic or sedimentary rocks and may represent a deeper crustal level.

The gross chemistry of this part of the Canadian Shield is shown in Table IV-2, based on analyses by a combination of X-ray fluorescence and chemical methods, of 78 composite samples (Eade, *et al.*, 1966). No significant regional variations in chemical composition were found and therefore variation of primary rock type in the area is unlikely. A comparison of the average compositions of rocks of the amphibolite and granulite facies shows minor but probably significant differences. Values for Al_2O_3, total FeP_2O_5, TiO_2 and possibly MgO and CaO, are higher in the granulite facies than in the amphibolite, and the amounts of K_2O, H_2O, and possibly SiO_2 are lower. These differences are attributed to loss by upward movement of more mobile constituents during regional metamorphism.

Isolated patches of clastic sedimentary rocks are scattered along two belts in the southeastern part of the region. These rocks have been named the Sakami Formation by Eade (1966a). The lower part of the formation is a red bed sequence composed of boulder conglomerate, arkose, mudstone, and siltstone, and the upper part is quartzite. Exposed thicknesses vary from 1,600 to 1,800 feet. The rocks are unmetamorphosed and flat lying except near the bordering north to northeasterly striking faults and form tectonically preserved basins. The lower red beds may have been deposited in fault-controlled basins but the upper quartzites were probably much more extensive. The Sakami rocks are similar to other Proterozoic sequences such as the Dubawnt Group and the Martin and Athabasca Formations in Churchill Province.

The youngest rocks are diabase dykes that strike northerly, northwesterly, and east-northeast. The northwesterly striking dykes are abundant in the northern part of the area. They have been dated at about 2,150 m.y. (Fahrig and Wanless, 1963). North-northwesterly dykes in the southern part of the area yield ages of about 1,900 m.y.

TABLE IV-2 *Average chemical composition of rocks from northeastern Superior Province (Eade, et al., 1966)*

	1	2	3	4	5
SiO_2	64.5	64.7	63.4	65.26	60.3
Al_2O_3	16.1	16.0	16.7	15.67	15.6
Fe_2O_3	1.5	1.5	1.5	1.44	—
FeO	2.9	2.9	3.4	2.82	7.2*
MgO	2.3	2.3	2.2	2.55	3.9
CaO	3.3	3.3	3.6	3.51	5.8
Na_2O	4.0	4.0	4.0	3.83	3.2
K_2O	2.78	2.81	2.58	2.93	2.5
H_2O (total)	0.8	0.9	0.7	0.82	—
TiO_2	0.48	0.47	0.53	0.46	1.0
P_2O_5	0.16	0.16	0.19	0.02	—
MnO	0.08	0.08	0.07	0.07	—
CO_2	0.2	0.2	0.2	0.16	—
Total	99.10	99.32	99.07	99.54	99.5

1. Average composition of 200,000 square miles of northern Quebec, calculated from rapid rock analysis of 78 composite samples using X.R.F. and other methods.
2. Average composition of all amphibolite facies rocks of the area.
3. Average composition of all granulite facies rocks of the area.
4. Average composition of selected rocks determined by classical chemical analysis (J. A. Maxwell, analyst).
5. Average composition of the continental crust, one part basic and one part felsic rock (Taylor, 1964).

* Total iron.

Mistassini Subprovince

Three groups of sediments occur in the Lake Mistassini region that are probably Aphebian in age as they are present as a little-deformed cratonic cover on the Archean. The older rocks constituting the Chibougamau Group are between Lake Chibougamau and Lake Mistassini. They consist of paraconglomerate, arkose, and shale lithologically similar to those of the Gowganda Formation of the Huronian Supergroup of Southern Province and are probably correlatives.

The Otish Mountains Group, northeast of Lake Mistassini, is about 3,000 feet thick and consists of medium- to coarse-grained quartzite and subarkose, which near the base contain numerous layers of grit and lenses of quartz-pebble conglomerate (Bergeron, 1957b). These rocks are overlain by red subarkose, shale, and siltstone. They are not metamorphosed and are very gently folded. Deformation increases to the southeast near the Grenville Front. According to S. M. Roscoe, the Otish Mountains strata are similar to the Lorrain Formation of the Huronian Supergroup and they may be correlative.

The Mistassini Group occurs around Lake Mistassini. It has been divided into five formations (Bergeron, 1957a). The lowest formation, the Papaskwasati, comprises well-bedded quartzite and feldspathic quartzite with quartz-pebble conglomerate lenses near the base. The succeeding Cheno River Formation is composed of conglomerate, subarkose, and greywacke interbedded with dolomite. Grey dolomites with shale and ferruginous dolomite interbeds of the Lower Albanel Formation are followed by a thick sequence of sandy dolomites with stromatolitic zones, dolomite, and quartzite of the Upper Albanel Formation. The youngest formation, the Temiscamie, is composed, in order of succession, of quartzite with conglomerate lenses, slate, iron-formation, and slate. The thickness of the Mistassini Group is at least 6,500 feet and disconformities occur at the bases of the Upper Albanel and Temiscamie Formations. The Mistassini strata have a north–northeasterly to northeasterly strike and a regional dip to the southeast. They are gently folded about northeasterly striking axes and cut by faults that trend parallel with the fold axes. Along the southeastern margin of the basin, a

thrust fault that dips 44 to 55 degrees to the southeast separates Mistassini rocks from gneisses of Grenville Province. The sediments are sheared and steeply dipping near the fault and the gneissic rocks on the south side are crushed and sheared. On the basis of stratigraphic and lithologic similarities, particularly the presence of iron-formation, the Mistassini Group is tentatively correlated with Aphebian rocks of the Kaniapiskau Supergroup of the Labrador Geosyncline. The iron-formations, although much deformed in Grenville Province, are nearly contiguous. The relative ages of the Otish Mountains and Chibougamau sequences with Mistassini rocks are not established.

Strata of the Otish Mountain Group are intruded by gabbro sills that are dated at about 1,900 m.y. North–northwesterly diabase dykes intrude Mistassini, Otish, and older rocks.

Timmins–Noranda–Malartic Area

Volcanic and sedimentary rocks of the Timmins–Noranda–Malartic area are part of the largest and probably thickest Archean complex in Superior Province. These rocks underlie much of Abitibi Subprovince. Despite much geological work details of the Archean stratigraphy and of lateral facies variation within the belt remain largely unknown. The stratigraphy in local areas has been established but it is impossible to correlate between these areas with confidence.

In the Cadillac–Malartic area, Gunning and Ambrose (1939) have divided the Archean rocks into four groups (Fig. IV-6). The Malartic Group at the base consists of basic lavas and pyroclastics with acidic lavas and tuffs near the top. Numerous gabbroic and ultramafic masses occur in the basic lavas and felsic intrusions are included with the acidic volcanics. A minimum thickness of 10,400 feet has been estimated at one locale where the structure is known, but it seems likely that in other places at least 20,000 feet of flows may be exposed. The Kewagama Group overlies these rocks conformably. In several places interbedding of the sedimentary and volcanic rocks indicates a gradual transition from volcanism to sedimentation. The Kewagama Group comprises greywackes and argillites, which commonly occur as graded couplets, some quartzites, and scattered lenses of conglomerate that usually are present near the base of the group. Boulders in the conglomerates consist of volcanic debris with some rare cobbles of quartz and granitic rocks. North of Cadillac fault, the Kewagama sediments are 1,000 to 3,000 feet thick, and locally as much as 10,000 feet. The Blake River Group of volcanic rocks conformably overlies the Kewagama and the contact is characterized by interbedding and interfingering of sediments and volcanics. The Blake River Group comprises basic to intermediate flows and pyroclastics, and acidic flows and tuffs with related intrusive gabbros, diorites, and felsitic rocks. The acidic flows are most abundant in the upper part of the sequence, particularly in the thick sections. These volcanics thicken from west to east along strike from a few hundred to thousands of feet, mainly due to a greater amount of acidic material. Gunning and Ambrose estimate a minimum thickness in the Malartic area of about 16,000 feet and in the Noranda area to the east a section of 40,000 feet has been measured (Baragar, 1966). These rocks are overlain by the Cadillac Group which consists of greywacke, some conglomerate, and minor amounts of tuff, iron-formation, and quartzite. A minimum thickness of 15,000 feet of strata is exposed. Gunning and Ambrose (1939) consider that these rocks overlie the Blake River volcanics conformably. No structural discordance was noted and interbedding and interfingering along strike of the two units were found in several areas. However, to the east, south of Noranda, M. E. Wilson (1962) presents evidence for an unconformity at the base of the Cadillac Group. The stratigraphic relationship determined by Gunning and Ambrose is on the north limb of a syncline, the axis of which trends east to southeast and lies north of the Cadillac fault. The fold axis is within Cadillac sediments and these rocks are on opposite limbs of the fold. The stratigraphy of the older rocks on the south limb of the fold is certain. A wide zone of shearing along the Cadillac fault obscures relationships and the hinge of the fold is breached by the fault. A thin band of volcanic rocks occurs beneath the Cadillac sediments on the south limb of the syncline at the east end of the belt of sediments. The sediments and volcanics are structurally conformable and both face north toward the fold axis. Gunning and Ambrose (1939) on the basis of lithologic similarity, stratigraphic position, and orientation of minor structures include these volcanics with the Blake River Group. Metamorphosed greywacke and shale of the Pontiac Group lay beneath the volcanic rocks with apparent conformity and face north near the contact with the volcanics. The Pontiac Group is more highly metamorphosed and complexly folded than rocks north of the Cadillac fault and grades southward into granitic gneisses. Thin conformable bands of basic volcanic rock are present within the sediments which are correlated by Gunning and Ambrose with the Kewagama Group. The extensive distribution of the Pontiac Group and the lack of underlying Malartic rocks have been attributed to southward thickening of the sediments, thinning of the Malartic volcanics, or tectonic thickening of the less competent sediments. Latulippe (1966) postulates that the Pontiac sediments are the equivalents of both Malartic volcanics and Kewagama sediments (Table IV-3).

In the Noranda area great thicknesses of volcanic rocks are present that appear to be equivalent to the Blake River Group. They are overlain by sediments of

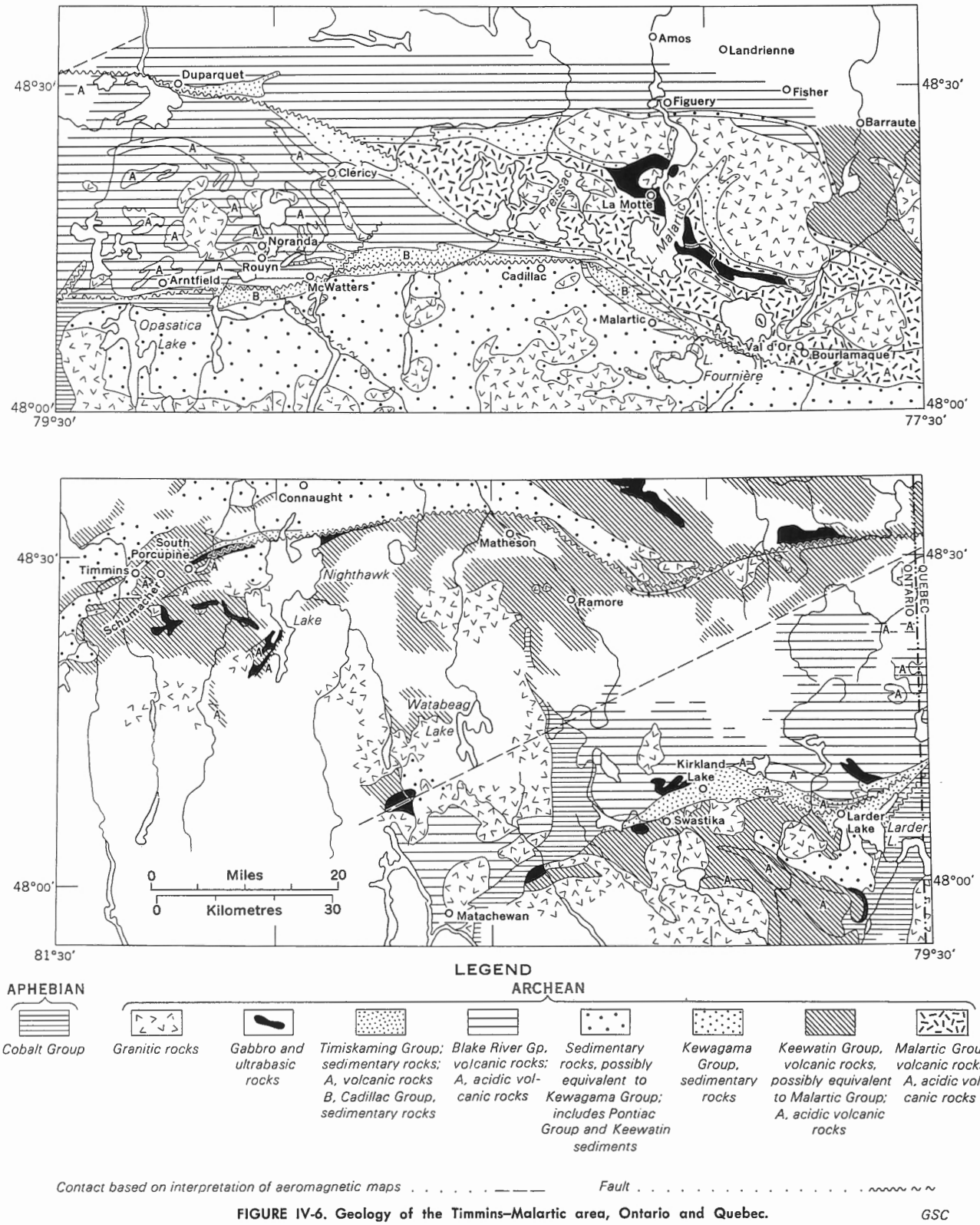

FIGURE IV-6. Geology of the Timmins–Malartic area, Ontario and Quebec. *GSC*

the Cadillac Group, separated at least locally by an unconformity. The Cadillac sediments also unconformably overlie the Pontiac Group which, although not in contact with volcanic rocks of the Blake River Group locally, contains bands of flow rocks. Lead isotope data from sulphide deposits in the volcanics can be interpreted to

yield ages of 2,900 to 3,000 m.y. (Roscoe, 1965). If these deposits are related to the volcanism then this is possibly the age of the Blake River rocks.

In the Kirkland Lake and Larder Lake areas of Ontario, similar rocks occur along the strike of those described above. The oldest rocks are the Keewatin volcanics which are similar to the Blake River Group and are no doubt part of the same volcanic sequence. Greywacke and shale occur within and overlie the volcanic rocks. The Timiskaming Group unconformably overlies the Keewatin near Kirkland Lake, although to the east near Larder Lake the rocks are structurally conformable and the contact is mapped as a disconformity (Hewitt, 1963). Typical Timiskaming rocks occur in the north limb of an easterly striking syncline, the south limb of which is cut off by the easterly striking Larder Lake fault. As much as 16,000 feet of rocks are exposed in this structure and as the top of the sequence is not present, this represents a minimum thickness for the Timiskaming Group. The succession comprises greywacke, shale, conglomerate, and trachytic flows and tuffs. The conglomerate, 4,000 feet thick, occurs as lenses at a number of places within the other sediments. Trachyte and acidic porphyry constitute the bulk of the pebbles; granitic pebbles are very rare. The volcanism and sedimentation were contemporaneous, as the volcanic rocks were eroded to provide the boulders in the conglomerates and probably also the tuffaceous debris in the fine-grained greywacke. Roscoe (1965) points out that the Timiskaming rocks are of a chemical suite unique to Superior Province. They are alkalic rocks rich in potash, soda, barium, and strontium. Lead–zinc veins within the sequence yield lead isotope ratios that have modal ages of 2,350 m.y. which suggests that the mineral deposits and possibly also their host rocks were formed during the Kenoran Orogeny. They would be, therefore, much younger than the Keewatin volcanics and sediments. Greywacke, shales, and conglomerates south of Larder Lake and the Larder Lake fault normally are correlated with the Timiskaming rocks described above, but do not contain trachytic volcanics. Also there are differences in the composition of the conglomerates in the two belts (Hewitt, 1963). Trachytic rocks are a major component of the pebbles in the northern conglomerates but are rare or absent in the south. Pebbles of chert, jasper, and iron-formation are more abundant in the northern conglomerates. It is possible that the two sequences are correlative, but the sediments south of Larder Lake may be equivalent to the Pontiac and Kewagama Groups, in which case the underlying lavas would be equivalent to the Malartic Group;

TABLE IV-3 *Alternative correlations of Archean rocks, Abitibi Subprovince*

Porcupine	Kirkland-Larder Lakes	Noranda	Malartic
Timiskaming sediments 1,500′ minimum	Timiskaming greywacke shale trachyte tuff (may be younger than Cadillac) 16,000′	Cadillac conglomerate greywacke shale	Cadillac conglomerate greywacke shale 5,000′
unconformity	unconformity or disconformity	unconformity	
	Keewatin basic volcanics acid volcanics tuff	Blake River basic volcanics acid volcanics tuff 40,000′	Blake River basic volcanics acid volcanics tuff 6,000′ minimum
Keewatin sediments greywacke shale conglomerate 3,000′	(may include sediments equivalent to Pontiac or Cadillac and volcanics equivalent to Malartic)	Pontiac greywacke shale metamorphic equivalents	Kewagama greywacke shale 10,000′ minimum
Keewatin basic volcanics acid volcanics tuff 6,000′ minimum		(sediments may be equivalent to Malartic)	Malartic basic volcanics acid volcanics tuff 10,000′ minimum

or they may correlate with the Cadillac Group which is likewise devoid of trachytic lava.

In the vicinity of Timmins, northwest of Kirkland Lake, a similar succession has been established (Ferguson, 1966). The oldest rocks are basic lava flows, about 4,000 feet thick, massive or pillowed, and with minor related phases such as variolitic flows, breccias, and basic tuffs. Gabbro, diorite, and related differentiates intrude the lavas, and thin layers of argillite and of cherty rocks are interbedded. The lavas are conformably overlain by as much as 2,000 feet of acidic pyroclastics with minor lavas and related intrusions. Roscoe (1965) points out that these volcanic rocks have a different expression on regional aeromagnetic maps than do the lavas in the Kirkland Lake area (Fig. IV-7). On such maps a northeasterly trending contact can be drawn between two magnetic regional patterns. A few top determinations indicate that flows on both sides of this line face south and therefore the lavas in the Porcupine area are oldest; they may be equivalent to Malartic flows. The lavas are conformably overlain by as much as 3,000 feet of greywacke and argillite commonly referred to as Keewatin or Keewatin-type sediments. They are overlain unconformably by a minimum of 1,500 feet of thinly bedded alternations of greywacke and shale of various types with a few beds of quartzites and conglomerate of variable thickness at or near the base. Basic and acidic volcanic rocks comprise the bulk of the pebbles in the conglomerate. The sediments are generally considered to be equivalent to the Timiskaming Group. Ultrabasic rocks appear to be intruded along faults such as the Destor–Porcupine and are later than the youngest sediments.

A fourfold division of the Archean rocks has been established at the eastern end of the Timmins–Noranda–Malartic belt. In other parts of the belt where the stratigraphy is known such a division may be valid but regional correlations are uncertain. All the Archean rocks are folded along easterly trending axes. These folds appear to be modified in places by cross-folds of various orientations, and are cut by several easterly trending faults with wide chlorite–carbonate shear zones. Major faults of this type, such as the Larder Lake, Cadillac, and Destor–Porcupine, occur at or near the axes of synclines. Lead isotope data from gold deposits within the fault zones or related structures suggest that they formed during the Kenoran Orogeny, and accordingly the shearing probably

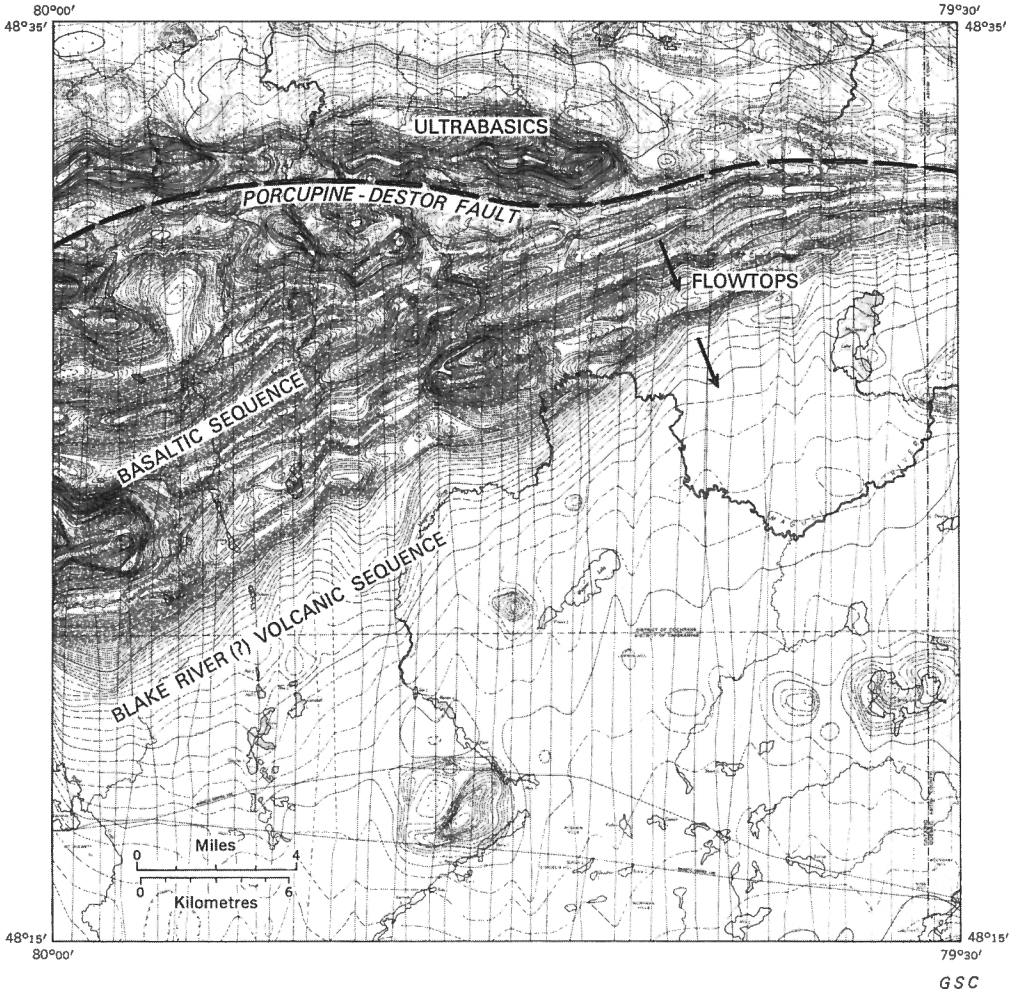

FIGURE IV-7
Aeromagnetic expression of the volcanic sequence in the Timmins-Kirkland Lake area, Ontario (Roscoe, 1965).

took place during the folding. The Archean rocks are cut by granites of various composition. The younger, higher level stocks tend to be richer in potash than the larger batholiths and granitic gneisses. Metamorphism, except near the contacts with granitic rocks, is generally in the greenschist facies. These rocks are cut by faults that trend northwest, north-northwest, and northeast and that are much younger. Some cut rocks of Aphebian age, displace diabase dykes that are as young as 1,200 m.y., or flank a graben of Ordovician and Silurian sediments.

Kapuskasing Subprovince

In Kapuskasing Subprovince the intensity of the gravity and magnetic fields is anomalously high (Fig. IV-8). Much of the geology is typical of Superior Province (Bennett, *et al.*, 1966). The oldest rocks are basic to acidic volcanic flows and pyroclastics with greywacke and shale. They occur in easterly trending belts, folded about easterly to southwesterly axes. Metamorphism is at the amphibolite grade, and along the margins of belts the sedimentary and volcanic rocks grade to biotite–hornblende–quartz–feldspar gneiss and migmatite. Banding is parallel with structures in the volcanic and sedimentary rocks and the gneisses are considered to be their highly metamorphosed and granitized equivalents. Basic rocks intrude the older rocks and possibly also the gneisses. All rock types are cut by synorogenic granitic intrusive rocks that include granodiorite, quartz monzonites, granite, and pegmatites. These masses occur as large batholiths or smaller stocks and are commonly bounded by migmatite. A south-southwesterly striking belt of amphibole–pyroxene–garnet–feldspar gneiss of the granulite facies of metamorphism is partly bounded by faults. Foliation is parallel with the trend of the belt and therefore transects the regional easterly structural trends of Superior Province. The belt is about 120 miles long and 6 to 12 miles wide. The pyroxene-bearing gneiss appears to occur in a zone of complex faulting in which individual faults strike north to northeast and are marked by zones of intense shearing and bands of mylonite. The gneiss is in blocks separated from each other by faults

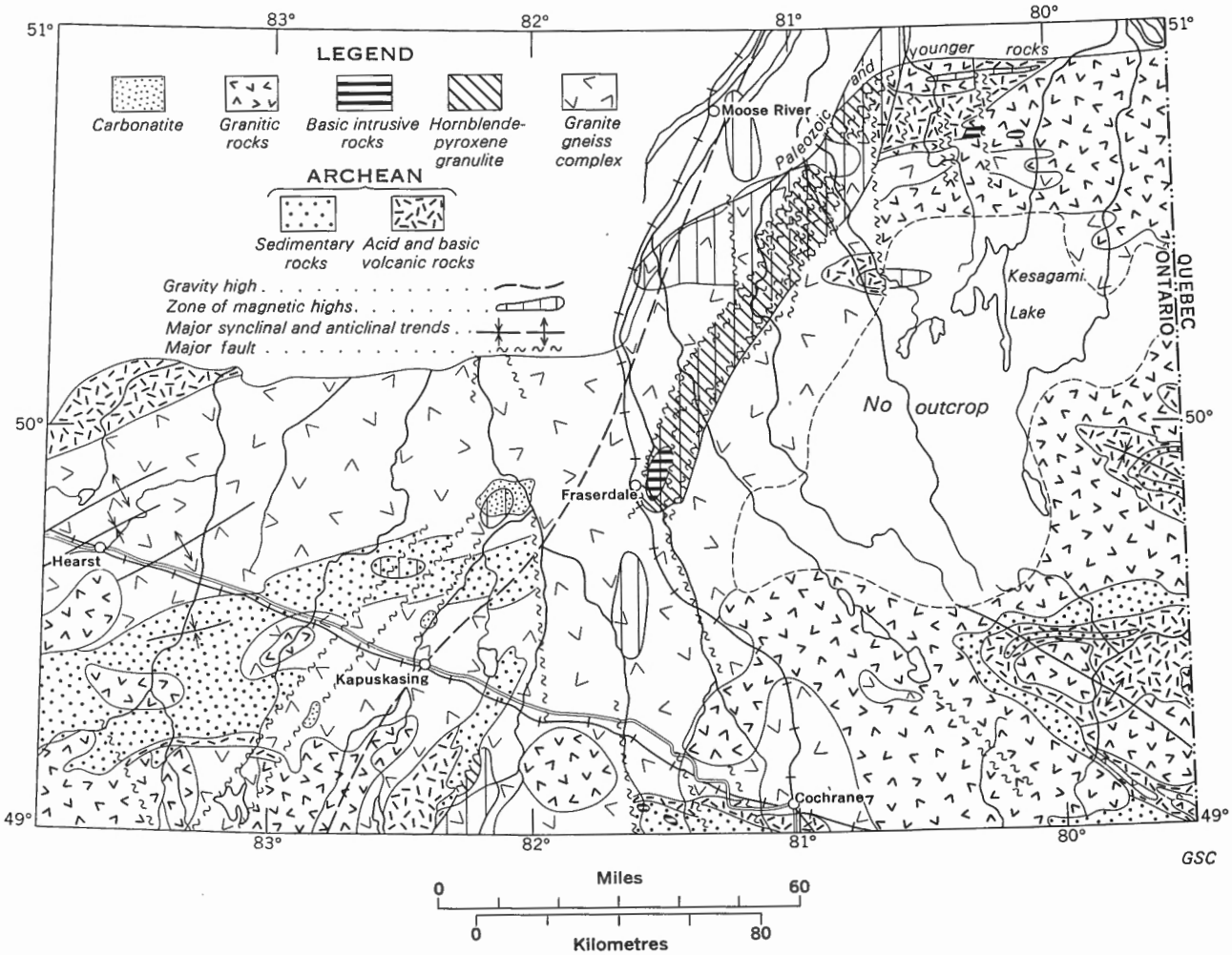

FIGURE IV-8. Geology of a part of Kapuskasing Subprovince, Superior Province, Ontario.

and appears to owe its present position to movement along the faults.

Of the fifteen known alkaline complexes in Superior Province ten are in the subprovince (Fig. IV-9). According to Parsons (1961) they are small, elliptical-shaped masses that cut granitic gneiss, basic volcanic rocks, or pyroxene-bearing gneiss. Individual masses generally have a central core of carbonatite bounded by nepheline syenite of variable composition and zones of pyroxenite or pyroxene and carbonate-rich rock. The essential minerals of the syenite are alkaline feldspar, nepheline, soda pyroxene, and ferric biotite. Parsons suggests that the sequence of events included first the drilling of a central diatreme by explosive gases and the ejection of alkaline volcanic rocks, since eroded; this was accompanied by alkaline metasomatism of the country rocks and followed by the carbonatization of the rocks of the volcanic neck and the emplacement of carbonatite and syenite. Many of the carbonatite complexes lie in or near major northeast-striking faults (Bennett, *et al.*, 1966) that are probably related to the faults that bound the zones of pyroxene-bearing gneiss. Eleven of the alkaline complexes have K–Ar dates that with one exception fall into two well-defined groups averaging about 1,050 m.y. and 1,700 m.y. (Gittins, *et al.*, 1967). One complex yielded a date of 1,560 m.y. These dates suggest that the faulting took place either almost continuously from about 1,700 to 1,050 m.y. ago or mainly during the two separate intervals that correspond approximately to the Hudsonian and Grenvillian Orogenies. The younger period of intrusion also roughly correlates in time with the extrusion of Keweenawan lavas in the Lake Superior region. It is possible, then, that this narrow fault complex was imposed on the stable Superior craton during the Hudsonian Orogeny and reactivated during the Grenvillian Orogeny.

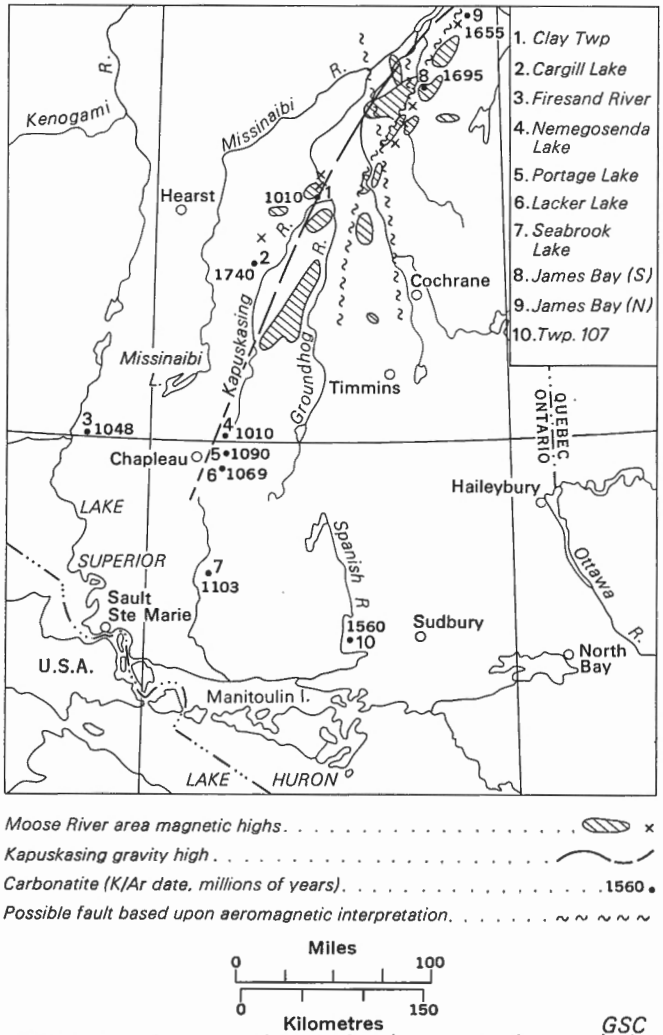

FIGURE IV-9. Location and ages of carbonatites in the Kapuskasing Subprovince, Superior Province, Ontario.

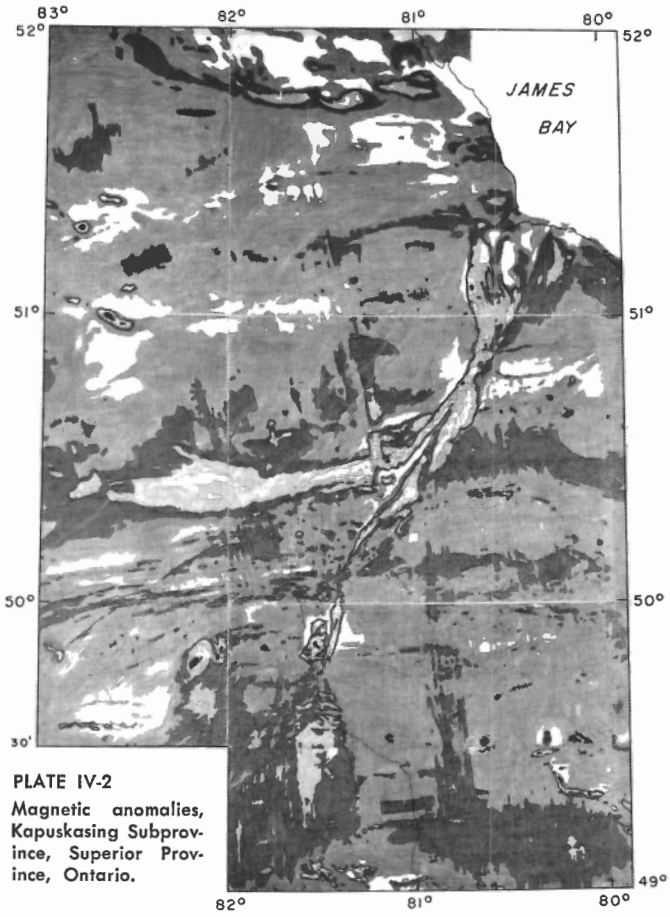

PLATE IV-2
Magnetic anomalies, Kapuskasing Subprovince, Superior Province, Ontario.

Two separate, but probably related, geophysically anomalous zones are present in the area. The zone of anomalously high magnetic intensity is in two parts (Pl. IV-2). In detail these regional highs are composed of a number of individual magnetic peaks separated by areas of lower magnetic intensity. The northern part corresponds approximately to the belt of pyroxene-bearing gneiss and also to a large gabbro intrusion. The southern part is underlain by foliated granitic rocks that include some zones of pyroxene-bearing gneiss. Other small

magnetic anomalies are related to the alkaline complexes or gabbro and ultramafic intrusions. Linear anomalies occur over diabase dykes and some faults. Two narrow linear belts of low intensity correspond with the complex of faults that bound the pyroxene-bearing gneiss. Gaucher (1966) considers that the aeromagnetic anomalies can be directly correlated with the magnetic susceptibility of the rocks presently exposed.

The Kapuskasing gravity high (Innes, 1960) is a broad regional feature, the axis of which trends southwesterly from Moosonee on James Bay to Chapleau and lies about 20 miles west of the magnetic high. Some local gravity anomalies appear to be related to the dense pyroxene-bearing gneiss or the basic intrusions. The regional gravity high, however, may reflect not only the exposed pyroxene gneiss, but also unexposed zones of these rocks or deeper seated crustal displacements involving the mantle.

Quetico Subprovince

Quetico Subprovince is characterized by linear structures that trend easterly contrasting with the curvilinear structures in the granitic terrains beyond the borders of the belt. The belt is as much as 60 miles wide and extends from Rainy Lake eastward for about 500 miles to Kapuskasing. The rocks are highly metamorphosed sediments, migmatites, granitized gneisses, and gneissic or massive granitic rocks of approximate granodiorite composition. Towards the margins of the belt the degree of metamorphism and granitization tends to decrease so that the nature of the original sediments can be studied. The belt of gneisses is bounded along a large part of the strike length by bands of Archean volcanic and sedimentary rocks.

Most of the sediments and their metamorphosed equivalents within the western part of Quetico Belt have at one time or another been referred to as Couchiching, a name applied by A. C. Lawson in 1888 to metamorphosed sediments that occur below the Keewatin lavas in the Rainy Lake area. The term has since been used for sediments and their metamorphosed equivalents in various parts of Quetico Belt, and it has been argued that such rocks occur stratigraphically below, within, and above Keewatin-type lavas or are separated from the lavas by a fault (Pettijohn, 1937; Goldich, *et al.*, 1961). The sediments are for the most part metamorphosed to about the amphibolite facies and consist essentially of quartz, plagioclase, and biotite. Muscovite, iron oxide, garnet, microcline, staurolite, hornblende, and cummingtonite occur in various combinations. The structure is gneissic, and the micas well oriented and concentrated in bands. Near the boundaries of the belt where the metamorphism is less intense, relict clastic textures and graded bedding can be found, which indicates that the rocks were originally a type of greywacke. Their chemical composition in the Lac des Mille Lacs area northwest of Port Arthur corresponds to that of greywacke. The thickness is unknown, but in a number of areas several thousand feet is exposed. In the interior of the belt, the sediments are granitized and intruded with granitic rocks. Migmatite zones varying in width occur along the margin of granitic rocks. The latter rocks vary in composition from granodiorite to granite and may be richer in potash than the granitic rocks outside the belt.

The Couchiching rocks are folded into easterly to southwesterly trending folds. Foliation is similarly oriented as a rule. Structures in other adjoining Archean rocks also have the same orientation so that Couchiching-type sediments are structurally conformable with Keewatin-type lavas and sediments. Faults are marked by rather wide shear zones containing chlorite and carbonate in volcanic rocks and by mylonite in the more siliceous rocks. They occur along the contact between the Couchiching sediments and Keewatin-type basic volcanics; one extends for 60 miles, from Lac des Mille Lacs to Steeprock Lake (Hawley, 1930); a second major fault occurs along the southern boundary of the belt in the Kashabowie Lake area. Top determinations, particularly in the sediments on both sides of these faults, indicate that the Couchiching strata underlie the volcanic rocks.

Recent work (Goldich, *et al.*, 1961) in the type area confirms Lawson's work and places the Couchiching metasediments conformably below Keewatin lavas. In Bennett and Tanner townships east of Atikokan Couchiching-type rocks are stratigraphically below and conformable with basic and acidic lavas and appear to be the lowest unit of the Keewatin Group. In Gorham township north of Port Arthur along the south boundary of the belt, many top determinations near the contact place Couchiching sediments below basic lavas. Interbedding of volcanics and sediments takes place near the contact, which suggests a transition to volcanism without an intervening break. East of Lac des Mille Lacs, the Couchiching-type sediments occur below lavas. All these areas are west of Lake Nipigon where most of the Couchiching-type sediment is found. East of Lake Nipigon in the Long Lake district, similar metasediments many thousands of feet thick occur along the margin of Keewatin-type lavas and sediments. On the basis of scattered top determinations and structural considerations, Macdonald (1941) places them stratigraphically below the lavas and considers them to be Couchiching type.

Along the margin of Quetico Subprovince in long, narrow, east-trending bands are conglomerate, greywacke, and shale with some iron-formation that are often referred to as Timiskaming or Timiskaming-type sediments. In the east they have also been given local names. They occur stratigraphically above Keewatin-type lavas, are conformable with them, and often are interbedded or interfingered with the lavas. At Steeprock Lake iron mine near Atikokan on the north border of the Quetico

Belt, Jolliffe (1966) includes in the Steeprock Group a basal conglomerate as much as 600 feet thick, a dolomite as much as 1,500 feet thick, iron-formation, and a tuffaceous unit. The conglomerate is lenticular and discontinuous and contains beds of greywacke. Pebbles include granite, diorite, basic rocks, and quartz. The dolomite includes chert beds and breccias and contains considerable amounts of iron and manganese. The strata containing the ore are separated from the dolomite by an erosional unconformity. They comprise three members of variable thickness, the lowest of which is a manganiferous, ferruginous chert. The middle member comprises goethite and hematite with minor quartz and bands of chert. The uppermost or pyrite member occurs as conformable lenses along the upper contact of the ore zone and consists of pyrite-rich, aluminous or cherty sediments, in part carbonaceous. The uppermost strata of the Steeprock Group are pyroclastics with a few intercalated basic pillowed flows and sills. Jolliffe suggests that the Steeprock sediments are Archean in age and that they rest unconformably on granite. He does not attempt to correlate the rocks with other sediments in the region. Shklanka (1965) reports that the conglomerate at the base of the Steeprock Group is probably separated from the dolomite by volcanic rocks and that the granite, reported to be beneath the unconformity at the base of the sediments, is separated in places from Steeprock strata by faults or intrudes them in places. He also reports Steeprock sediments to be conformably interbedded with Keewatin-type volcanic rocks, and it would appear therefore that the Steeprock sediments are part of the sequence of Keewatin-type volcanics and sediments. Similar sediments that include limestone and iron-formations occur within or overlie the Keewatin-type lavas in the Lumby Lake area about 25 miles northeast of Steeprock Lake and at Finlayson Lake 10 miles north of Steeprock Lake. The sediments appear to be conformable with the lavas and are cut by granitic rocks.

All K–Ar age determinations from micas from the various granitic and metamorphic rocks in the Quetico Belt give Kenoran ages ranging from about 2,300 to 2,600 m.y. Rb–Sr dates from granitic rocks at Bad Vermilion Lake in the Rainy Lake area give ages in the range 2,500 to 2,600 m.y. (half-life 50 x 10^9) (Hart, *et al.*, 1963). Rb–Sr ages from samples of Couchiching and Keewatin rocks in the Rainy Lake area are about 2,700 m.y. (half-life 50 x 10^9) suggesting that these rocks are nearly equivalent in age. Zircons from the sediments and volcanics have an age of 2,750 m.y. which confirms the Rb–Sr data. However, dates from the zircons from two granites, one of which is the Bad Vermilion Lake granite, are also 2,750 m.y. The lack of agreement of the two methods on material from these granites suggests that either the zircons were inherited from older rocks or the granites were altered and, therefore, the system did not remain closed for strontium. There is no good geological evidence, however, to support either hypothesis.

English River Subprovince

English River Subprovince is characterized mainly by linear structures that strike easterly in contrast to curvilinear structures in granitic terrains beyond the borders of the belt. The belt extends eastward from Lake Winnipeg to the Phanerozoic cover of the Hudson Platform and has a maximum width of 60 miles.

The rocks in the interior of the belt are highly metamorphosed sediments, their granitized equivalents, migmatite, and granitic rocks that range in composition from quartz diorite to granite. The latter occur as elongate masses parallel with the trends of the metasedimentary rocks. The grade of metamorphism of the sediments is in the amphibolite facies and at least locally in the granulite facies. The rocks grade through zones of migmatite and granitic gneiss to the quartz diorite, granodiorite, and granite. Metamorphism has generally destroyed the primary sedimentary structures and textures but the composition of the metasediments suggests that originally they were greywacke and shale. The borders of English River Subprovince are defined by belts of basic lava, intermediate to acidic lava, pyroclastic rocks, greywacke, and shale. The succession is conformable and usually the basic lavas are basal. However, near Birch and Uchi Lakes, sediments at the base of the succession appear to be conformable with the overlying volcanic rocks (Goodwin, 1964). At Lake St. Joseph greywacke-type sediments are also the oldest part of the succession (Dyer, 1934). As these rocks are followed into the English River Belt, they become highly metamorphosed and resemble the sedimentary gneisses within the belt. Dyer named these older rocks the Miniss Group and suggested correlation with Couchiching sediments, for which there is little evidence other than that they underlie the lavas. Sedimentary gneisses near Savant Lake are divided into three groups: basic lavas, sediments, and acidic lavas and pyroclastics. The sediments include conglomerate, iron-formation, and greywacke. South of Rice Lake, Manitoba, the Rice Lake Group consists of basal basic lavas overlain by acidic lavas, and thence greywacke and shale with bands of basic lava (Stockwell, 1945). The San Antonio Formation comprises little metamorphosed conglomerate and feldspathic quartzite that unconformably overlie the Rice Lake Group and quartz diorite. The strata have been thrust over the older rocks and thrown into folds with easterly trending axes. These rocks are not known to be cut by granite and may be Aphebian.

The metasediments within the English River Belt, and along its margins, are folded about easterly trending axes that commonly plunge east. Locally these structures are distorted by cross-folding about northwest-trending axes. Easterly striking faults occur within the belt, and in the Werner Lake area small ultrabasic bodies are intruded

along these faults. North-to-northeast-striking faults also occur. A belt of high gravity extends east of the Manitoba border for about 180 miles, lying entirely within English River Subprovince. The cause of this feature is unknown, but faults containing ultramafic masses and an alkaline intrusion west of Lac Seul may be related. Micas from the English River Belt have given K–Ar dates of about 2,400 m.y. Some yield dates of 1,700 to 1,800 m.y. The latter coincide with the age of the Hudsonian Orogeny but there is little other evidence of its effects. In the Rice Lake area, deformation of the possibly Aphebian San Antonio Formation may have occurred during the Hudsonian. It is possible also that some faulting and intrusion of diabase dykes took place at about this time.

Cross Lake Subprovince

Cross Lake Subprovince is characterized by volcanic and sedimentary complexes that occur in long, slightly sinuous, easterly trending belts (Fig. IV-10). These rocks are intruded and granitized by granodiorite and tonalite. In the northwest, the subprovince is bordered by a band of gneisses mainly in the granulite facies of metamorphism and by granitic gneisses. These metamorphosed rocks were originally sediments and possibly volcanic rocks.

The oldest exposed rocks are the Hayes River Group of massive, pillowed, or amygdaloidal basic flows. Dacite and tuffs occur in the upper part of the sequences but locally rhyolite, agglomerate, and tuff with minor inter-

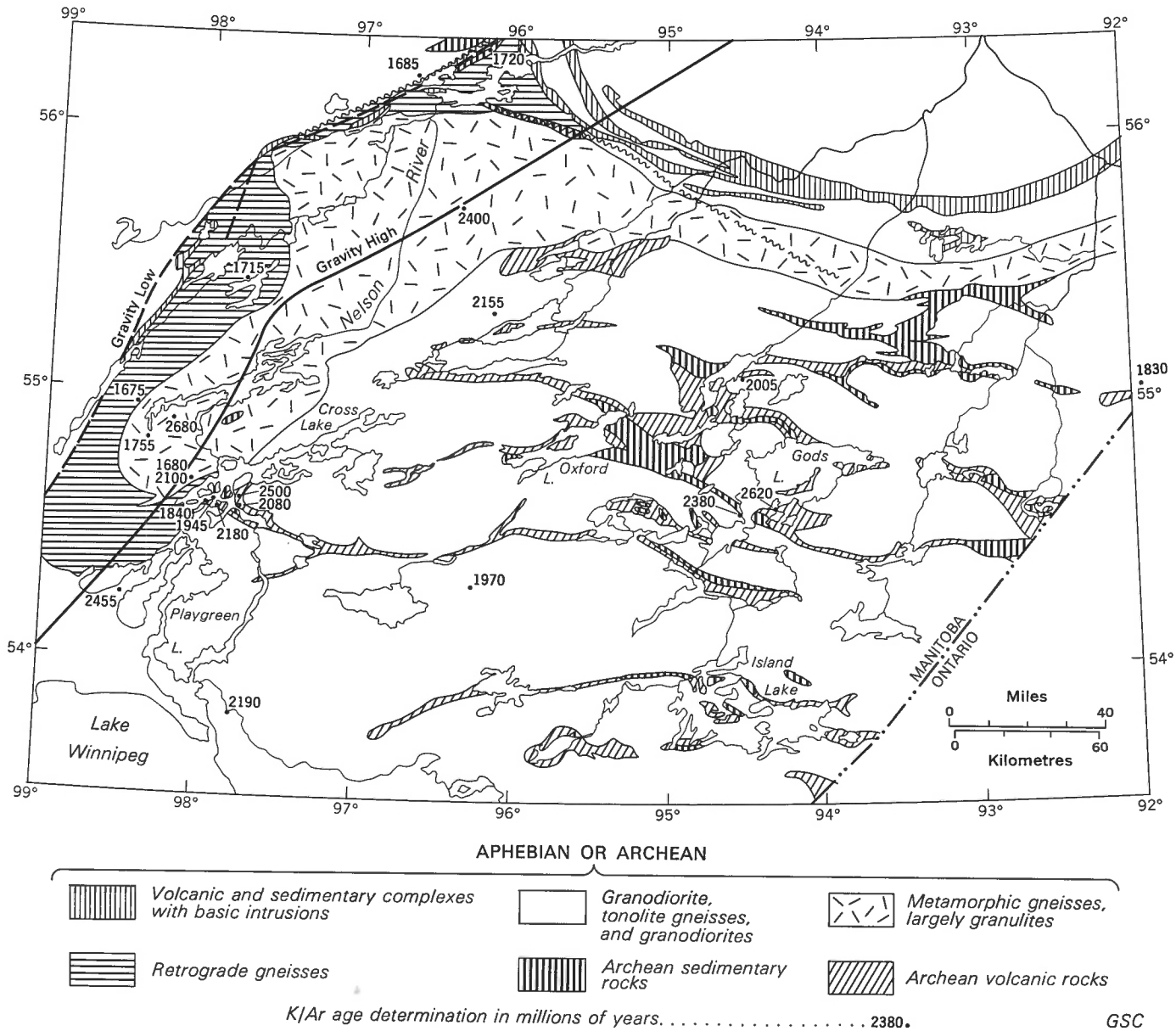

FIGURE IV-10. Geology of part of Cross Lake Subprovince, Superior Province, Ontario.

mediate or basic flows are the youngest beds. These lavas are generally named the Hayes River Group although there is no direct evidence for correlation from belt to belt. Basic and acid intrusives occur that are probably the intrusive equivalents of the volcanic rocks. Some of the quartz diorites contain quartz "eyes" that also occur as boulders in the overlying conglomerates, which suggests that an orogeny has intervened between the volcanism and sedimentation. However, the diorites may be only the intrusive equivalents of the volcanic rocks. Sediments overlie the lavas in most of the belts. Unconformities occur at Oxford, Knee, and God's Lakes but generally the rocks appear to be structurally conformable and interbedding of lava and sediment is common. It seems, therefore, that the sequence of lavas and sediments is essentially a conformable one with some local unconformities. The bulk of the sediments that comprise the Oxford, Cross Lake (Pl. IV-3), and Island Lake Groups consists of greywacke, subgreywacke, arkose, shale, and, in the upper part of the sequences, feldspathic quartzite. Graded bedding is common and the cleaner sandstones are crossbedded. The conglomerates occur as wedge-shaped masses generally near the base of the sequence. The pebbles are basic and acidic volcanic rocks, greywacke and other sediments, quartz, and granitic rocks set in a greywacke matrix. Composition of the pebbles varies with the different conglomerates. Granitic rocks appear to be more abundant in these conglomerates than in other parts of Superior Province. A large percentage are tonalites or granodiorites similar to rocks outcropping along the margins of the volcanic and sedimentary belts. The tonalite has been considered to be both older and younger than the Hayes River Group. The boulders in the conglomerates indicate, however, that such rocks existed in the source areas. Only in the Cross Lake area is the presence of basement established with certainty. Rousell (1965) reports on unconformity between sediments of the Cross Lake Group and a granodiorite and also suggests that, because the sediments are conformable with the underlying lavas, this granodiorite is the basement of the whole volcanic sedimentary complex.

The hornblende-bearing tonalites and granodiorites that are later than the Hayes River and Cross Lake Groups are usually gneissic and, particularly where their contact with older rocks is crosscutting, commonly grade into the older rocks through a zone of migmatite and gneiss. Locally, the tonalite–granodiorite complex is cut by stocks or small batholiths of massive pink granite or granodiorite with minor mafic minerals. They are probably late tectonic intrusions. The volcanic and sedimentary rocks are tightly folded along eastward and locally northeastward trending fold axes and locally, northward trending cross-folds are developed. They are cut by persistent easterly striking faults along which there probably has been considerable movement. Later faults striking northeast are common locally and appear to have had only small displacements.

In the Fox Lake Belt in the northern part of the subprovince are pillowed lavas, massive basalt or andesite, breccias, agglomerates, and tuffs. Slates, argillites, and schists are present within the volcanics. Quartz–feldspar–mica schist and gneiss with cordierite, garnet, or staurolite are along the northern part of the belt, apparently overlying the volcanic rocks (Davies, *et al.*, 1962). Basic and ultrabasic sills intrude the volcanic sequence. One sill, about 2,500 feet thick, is layered partly serpentinized pyroxenite, peridotite, dunite, and gabbro. The Fox Lake rocks may be Archean or Aphebian in age and may be



PLATE IV-3

Garnetiferous greywacke of Archean Cross Lake Group, Cross Lake, Manitoba.

the eastern extension of the sediments and volcanic rocks of the Thompson Belt.

The belt of gneisses in granulite facies of metamorphism along the northwest boundary of the province is considered to be the equivalent of Archean rocks of Cross Lake Subprovince (Bell, 1966; Rousell, 1965; Patterson, 1963). Chemical analyses of the paragneisses within this belt correspond to those of greywackes. The gneisses are commonly pyroxene-bearing and consist of mineral assemblages typical of various phases of the granulite facies. They are converted by granitization to granodiorites. Within the adjacent Churchill Province a zone of gneisses of the amphibolite facies is interpreted by Bell as representing the granulitic rocks retrogressively metamorphosed, possibly during the Hudsonian Orogeny. The granulites coincide approximately with a belt of high gravity (Fig. IV-11) and with one of high magnetic intensity (Pl. IV-4). In the north the granulite band trends easterly and the grade of metamorphism decreases to the amphibolite facies. The coincident gravity high becomes diffuse. The Fox River Belt of volcanic and ultrabasic rocks is marked by a gravity high that merges with the one over the granulites. Wilson and Brisbin (1961) have concluded that the gravity anomalies are produced by density contrasts resulting from major warps deep in the crust and are not correlative with the near-surface material. Faulting in the boundary zone, they suggest, may be related to the warping of the crust and also may have provided zones of weakness along which ultrabasic rocks of the Thompson Belt of Churchill Province were intruded. The zone of faulting coincides approximately with a belt of low gravity that lies parallel with the gravity high over the granulite belt. However, no evidence exists that the boundary between Churchill and Superior Provinces is a single fault or even a continuous zone of faulting. It is rather a metamorphic front within the Archean rocks. In mixtures of rocks such as granulites and granitic gneisses, there are significant density contrasts that affect gravity, and, in more recent work, Hall and Brisbin (1965) have tentatively concluded that the gravity anomalies are related in part to variation of density in the upper granitic layer of the

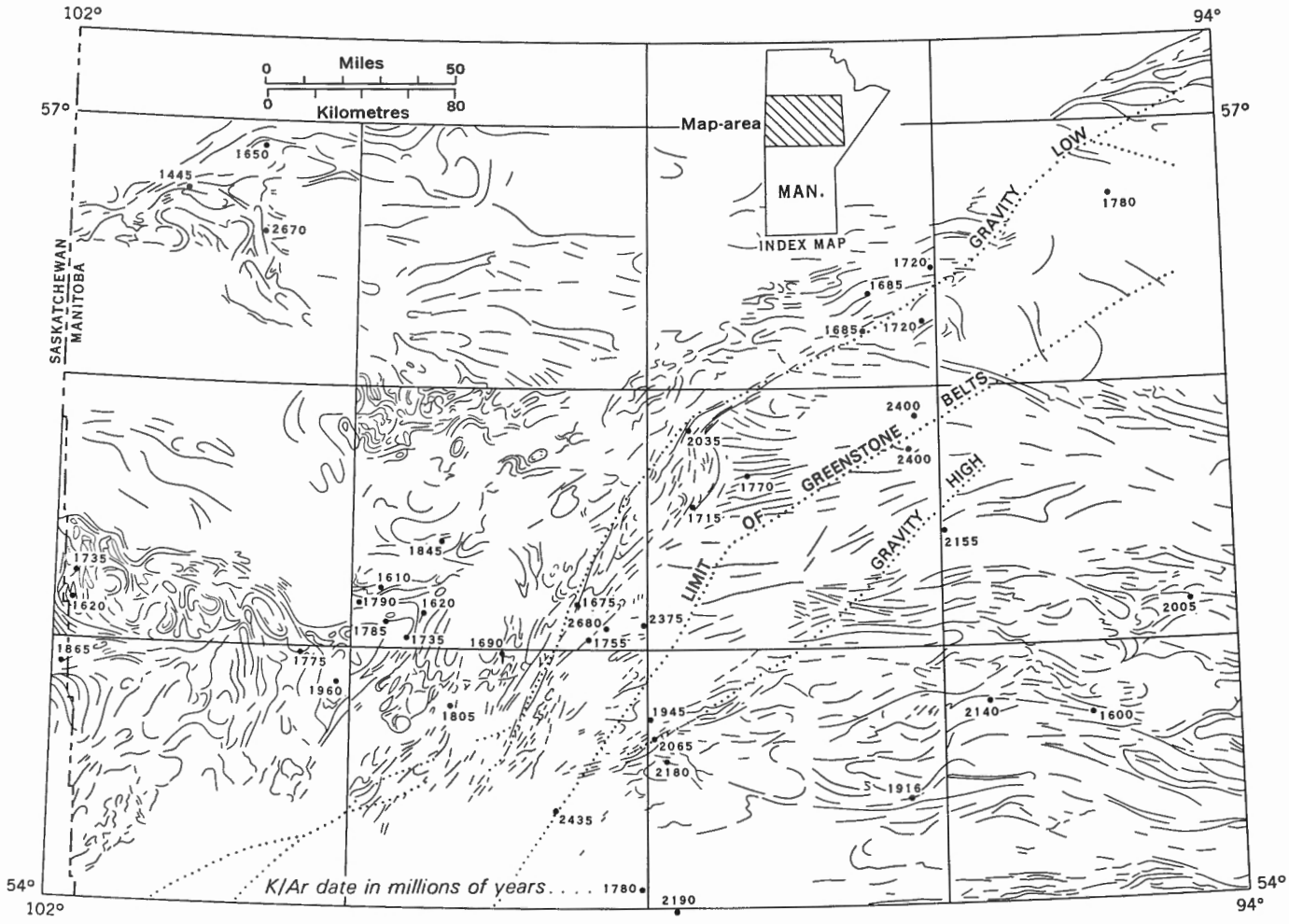

FIGURE IV-11. Magnetic and structural trends in the vicinity of the boundary between Superior and Churchill Provinces, Manitoba (Wilson and Brisbin, 1961).

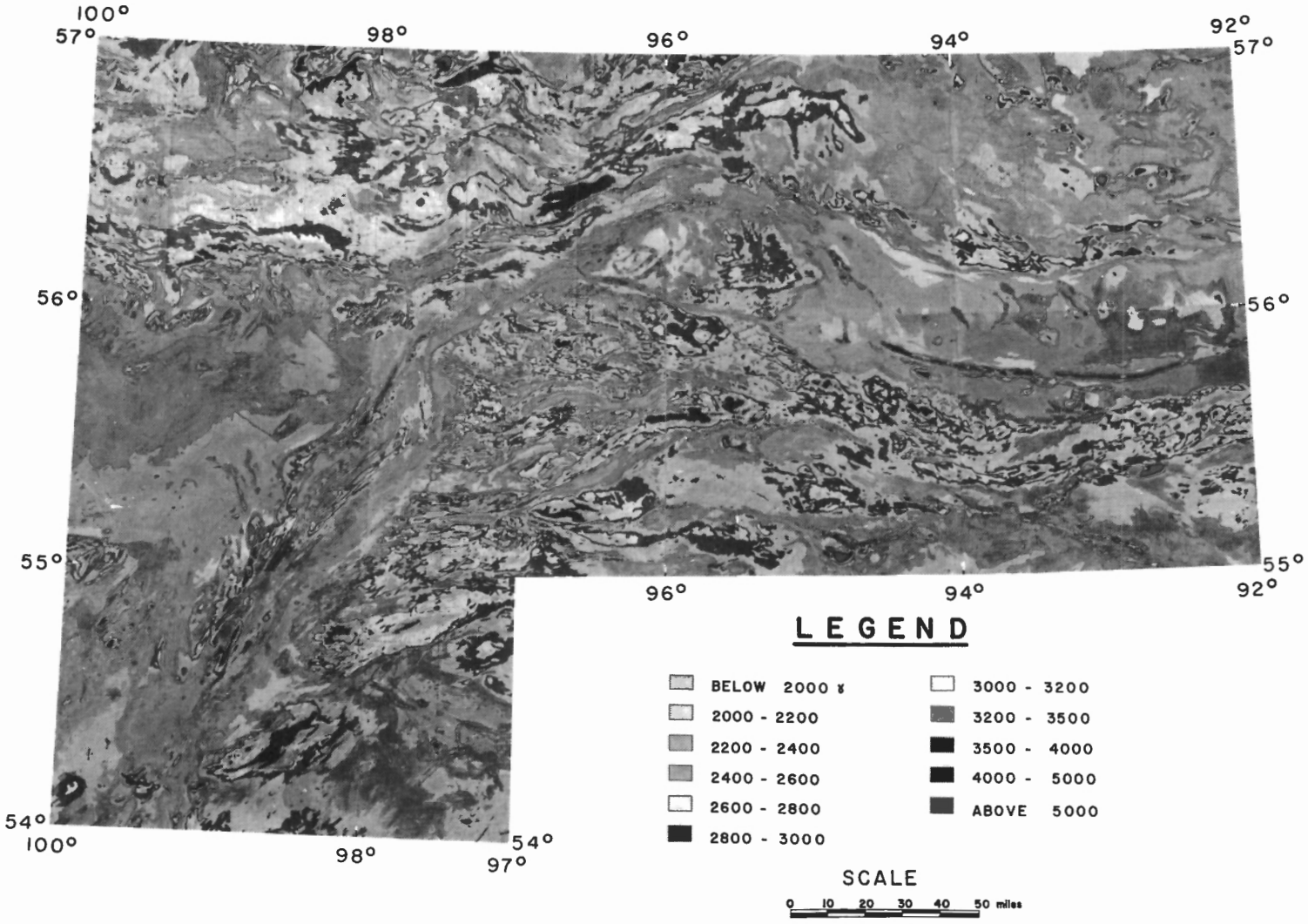

PLATE IV-4. Magnetic anomalies, vicinity of boundary between Churchill and Superior Provinces, Manitoba.

crust. The magnetic anomalies also correlate with the surface geology and appear, therefore, to be produced by near-surface material (Kornik and MacLaren, 1966).

K–Ar dates in Cross Lake Subprovince are unlike those for most other parts of Superior Province. Most biotites and hornblende from the granulitic rocks yield Kenoran ages that range from 2,100 to 2,700 m.y. A sample of biotite from the granodiorite beneath the Cross Lake sediments gives an age of about 2,500 m.y. (Rousell, 1965). Biotites from the granodiorite, tonalite, and metamorphosed equivalents of the volcanic-sedimentary rocks yield mainly younger ages ranging from 1,800 to 2,400 m.y. These anomalously young ages are presumed to have resulted from argon loss caused by heat effects of the Hudsonian Orogeny in the border zone of Superior Province. However, the granulitic rocks have yielded Kenoran ages and therefore appear unaffected by the orogeny.

SLAVE PROVINCE

Tectonic Summary

The oldest rocks in the province are Archean basic to acidic flows and pyroclastics and quartz-rich greywacke and shale. The volcanic rocks occur near or at the base of the generally conformable succession. They were probably extruded from various centres, possibly at different times, to form discontinuous belts so that, at times, volcanism and sedimentation proceeded simultaneously in different parts of the province. Most volcanism was subaqueous. The transition from volcanism to sedimen-

tation was gradual with local resumption of volcanism during sedimentation. The sediments were deposited in an eugeosynclinal environment.

The Archean rocks were folded, metamorphosed, and intruded by granite during the Kenoran Orogeny. A possible sequence of these events is: folding and low-grade metamorphism; granitization and intrusion of syntectonic granitic rocks and metamorphism overlapping in time, early folding, cross-folding and shearing; intrusion of high-level late tectonic granodiorites and pegmatites, and related metamorphism. The time separation of these events is unknown but to some extent they may be the result of roughly contemporaneous activity in different parts and levels of the orogen. K–Ar age determinations of these granites and metamorphosed equivalents of volcanic and sedimentary rocks give a mean age of 2,470 m.y.

After deep erosion, Aphebian sediments were deposited in a geosyncline and covered parts of Slave Province. Bathurst Subprovince includes Paleohelikian rocks. The Archean rocks were locally affected when the Aphebian rocks were deformed during the Hudsonian Orogeny. Three, and possibly four, diabase dyke swarms and at least as many fault sets of different ages and trends occur in rocks of Slave Province.



PLATE IV-5. Pillowed lavas of Archean Yellowknife Group, Slave Province, Yellowknife, Northwest Territories.

Archean

Volcanic and sedimentary rocks of the Archean Yellowknife Group occur in irregularly shaped areas, separated by masses of younger Kenoran granitic rocks. The Archean rocks underlie about a third of Slave Province. Volcanic rocks lie at the base of the Yellowknife succession in discontinuous belts along the margins of sedimentary units, except at Mesa Lake (Ross, 1959) where sediments occur below volcanic rocks. The thickness of the various volcanic sequences varies from about 1,000 to 40,000 feet and within individual belts the thickness also varies greatly. Basic flows form the bulk of the volcanic successions (Pl. IV-5), varying in proportion of massive, pillowed, and variolitic varieties and in amount of associated banded tuffs. Intermediate to acidic flows, tuffs, and breccias are commonly found in the upper part of the sequence and in the thickest parts of the volcanic belts. Acidic material forms less than 5 per cent of most volcanic sequences. Chemically, the common magma type is tholeiitic basalt with certain calc-alkaline characteristics that may result from modification by sialic contamination (Baragar, 1966). Sills, dykes, and irregularly shaped masses of massive fine-grained diorite or gabbro are found within the lavas and are their intrusive equivalents.

The lavas are overlain by thick monotonous sedimentary sequences. In most areas the contact is conformable and transitional with interbedding of flows or tuffs and sediments occurring in the contact zone. These rock types also interfinger along strike indicating some simultaneous sedimentation and volcanism. Relatively thin bands of volcanic strata occur within sediments above the main volcanic sequences, indicating local resumption of volcanism during sedimentation. The sediments are mainly quartz-rich greywackes or feldspathic greywacke with shale or argillite interbeds. Relative thicknesses of these interbedded and graded couplets vary but generally average less than a foot (Pl. IV-6). At some horizons argillite or greywacke occur as thick homogeneous units. Relative amounts of sand and shale as well as grain-size in the clastic units vary. Calc-silicate bands or lenses are common in the sandy portions in some areas. Primary structures such as graded bedding, groove casts, convolute lamination, flume structures, and parallel and ripple laminations are characteristic and suggest deposition by turbidity currents. Locally, thin beds of limestone or dolomite and dolomite breccia occur within the greywacke. They are commonly associated with feldspathic quartzites near the base of the sedimentary sequence and may have been deposited in shallow water around centres of volcanic activity. In some areas thin beds or lenses of paraconglomerate as much as 1,500 feet thick are interbedded with greywacke at or near the base of the sedimentary sequences above thick sections of volcanic rocks. Typically this rock consists of angular pebbles of basic and acidic volcanic rocks with some rounded cobbles of granitic rock and vein quartz with a greywacke matrix. Iron-formation is rare, but at Contwoyto Lake, bands of amphibolite, some of which are rich in iron, quartz, garnet, and iron sulphides, are interbedded with normal greywacke and may have been originally iron and aluminum-rich tuffs or sediments. The thick deposits of interbedded

greywacke–shale resemble the flysch facies of eugeosynclinal sequences. The source of these sediments is unknown, but the quartz-rich character of the greywackes and the presence of granitic cobbles and vein quartz in paraconglomerates that generally overlie the volcanics conformably suggest a source area consisting in part of granitic rocks and possibly metamorphosed sediments. Some material was also derived from the erosion of the underlying volcanic rocks.

East of Cameron River (Baragar, 1966) and in the Indian Mountain Lake area (W. W. Heywood) there is evidence to suggest the presence of a granitic basement beneath Yellowknife volcanic rocks. Metamorphosed basic dykes cut granitic and volcanic rocks but not the overlying sediments. The dykes may be temporally related to the volcanism and be feeders to the flows. However, Henderson (1941) has concluded that these granites were younger than lavas. Metamorphosed basic dykes also occur in the Yellowknife sediments in nearby areas. The metamorphism of dykes in the Cameron River area may be related to a young Kenoran granite that intrudes both the volcanic and granitic rocks.

The Wilson Island Group in the western part of the East Arm of Great Slave Lake is separated from the main part of Slave Province by younger Aphebian sediments. The age relationships of the Wilson Island and Yellowknife Groups accordingly are unknown but the Wilson Island is cut by a granite dated as 2,400 m.y. The group may represent a near-shore or shallow-water facies of the Yellowknife sediments. It is at least 11,000 feet thick and comprises acidic lava flows and pyroclastics with interlayered conglomerate and arkose that are overlain by cross-bedded quartzite, dolomite, and shale. These beds are succeeded by phyllite with thin interbeds of quartzites. Thin beds of iron-formation occur in the succession.

Kenoran Orogeny

Most of the K–Ar age determinations on minerals and rocks formed or metamorphosed during the Kenoran Orogeny fall in the range 2,300 to 2,600 m.y., and their mean value is 2,480 m.y. It has not been possible to date the rocks precisely enough to separate the various orogenic events.

During the Kenoran Orogeny the Yellowknife Group was folded, metamorphosed, and intruded by granitic rocks. The orientation of folds and the intensity of deformation vary throughout Slave Province but in most areas two directions of folding can be detected. The Yellowknife rocks were first deformed into northerly to south-northeasterly trending similar folds that plunge moderately north or south. The axial planes dip steeply. These folds in many areas were modifed by steeply plunging cross-folds, the axial planes of which generally strike in a northerly to northwesterly direction roughly at right angles to the axes of the early folds (Pl. IV-6). The sediments are tightly and intricately folded but the volcanic rocks are deformed differently probably because of their greater competency. Thin volcanic successions are thrown into broad folds that have the same orientation as the more intricate folding of the overlying sediments. In thick sequences the rocks generally form homoclinal successions that face away from the younger granitic rocks. Such sequences in some areas appear to be merely somewhat bent by the second folding. Shear zones in the volcanic rocks usually consist of chlorite–carbonate–sericite schists. These structures commonly are parallel with the strike of the lavas so that displacements cannot be measured. The shear zones probably occur during folding of the volcanic rock and may be caused by the second folding. Many shear zones in sediments result from shearing along limbs or axial planes of tightly closed folds, and may result from continuing compression after maximum closure is attained. In both rock types, shear zones may localize barren or mineralized quartz veins and, in some areas in the sediments, pegmatite dykes.

About two-thirds of Slave Province is underlain by granitic rocks of which possibly half are high-grade migmatite, mixed gneiss, banded gneiss, and granitic gneiss, while the remainder consists of batholiths of massive or slightly gneissic granitic rocks of various compositions. The gneisses are concentrated in the central and south-



PLATE IV-6. Graded beds of greywacke and shale of Archean Yellowknife Group, Yellowknife, Northwest Territories. The tight northwest-trending folds with vertical axes were produced in the late phase of Kenoran Orogeny.

eastern parts of the province and contain thin bands or septa of highly metamorphosed or partly granitized sedimentary or volcanic rocks. These rocks probably result from the granitization of Yellowknife rocks and form at a deep crustal level. Granitic rocks of at least two ages intrude rocks of the Yellowknife Group. The older intrusions are generally equigranular quartz diorite and granodiorite that contain biotite and commonly hornblende. They are faintly gneissic especially near their borders and are commonly bordered by migmatite and granitized sediments. These intrusions were probably emplaced during the metamorphism and early folding of the Yellowknife rocks at a higher crustal level than the granitic gneisses. The younger intrusions are granodiorite and quartz monzonite with biotite and muscovite as accessory minerals. They are generally massive and often porphyritic with megacrysts of potash feldspar. The younger intrusions are more potash rich than older plutonic rocks. The younger plutons commonly form oval or elliptical masses (Pl. IV-7) that in some areas are oriented roughly parallel with the axes of younger cross-folds. The stocks are partly concordant and partly discordant and in some, intrusion was achieved by pushing aside earlier deformed sediments. Contacts with older rocks are generally sharp, and migmatite zones, if present, are narrow or occur only along parts of the contact. The younger intrusions are surrounded by metamorphic aureoles and were emplaced after early folding and regional metamorphism. Pegmatites, commonly developed in the country rocks around the younger plutons, are usually simple, some bearing tourmaline, and lacking obvious internal zoning. East of Yellowknife, the pegmatites are complex and some are internally zoned. They contain such minerals as beryl, spodumene, ambligonite, garnet, and tantalite–columbite. They are arranged about the related granodiorites in zones defined by different mineral compositions.

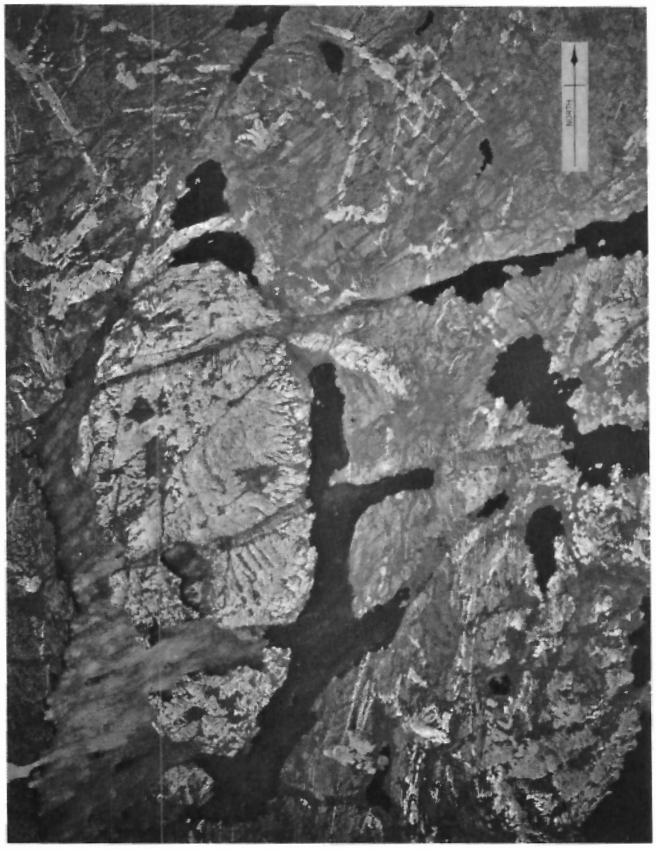

PLATE IV-7. Granite stock (light) intrusive into Archean Yellowknife meta-greywacke (dark), Staple Lake, Slave Province, Northwest Territories. Irregular white masses are pegmatite dykes and linear dark bands are diabase dykes. The east-trending diabase dykes are 2,150 m.y. old and the northeast one is 2,100 m.y. old.

All rocks of the Yellowknife Group have been metamorphosed. Degree of metamorphism varies from low greenschist to amphibolite facies. In the greywacke and shale in the lowest grade, chlorite, sericite–quartz, and plagioclase are the characteristic minerals. With increasing metamorphism, biotite is developed at the expense of chlorite and sericite and forms discrete grains and porphyroblasts. The transition to amphibolite facies is marked by the appearance of cordierite and andalusite metacrysts in the knotted schists. Staurolite and garnet are locally developed in this facies and chlorite is rare. Biotite is more abundant than in lower grades and muscovite is a common constituent. Cordierite and andalusite may occur together or separately. Sillimanite is formed in the upper amphibolite facies close to some granitic plutons and in areas where the sediments grade to migmatite. These rocks are coarse grained, and biotite, quartz, well-twinned plagioclase, and sometimes muscovite are also present. The basic volcanic rocks in the greenschist facies comprise plagioclase, actinolite, and chlorite. In the amphibolite grade, blue-green hornblende and plagioclase are the essential minerals. Initially, metamorphism accompanied deformation and some micas are oriented in minor structures related to folds. However, in some areas, in zones of higher metamorphic grade, metacrysts of andalusite and cordierite are not oriented and therefore metamorphism continued after deformation ceased. Some granitic plutons are surrounded by narrow zones of knotted andalusite- or cordierite-bearing schists but in other areas the development of these schists appears unrelated to the plutons, and is possibly a phase of progressive regional metamorphism. It seems likely that deformation, metamorphism, and intrusion of granites were elements of a sequence of overlapping events that began with folding and metamorphism and ended with intrusion of younger granitic rock in previously metamorphosed rocks. Older granitic plutons were probably intruded during metamorphism.

Post-Kenoran Events

Post-orogenic events in Slave Province include intrusion of diabase dykes, faulting, intrusion of an alkaline pluton, and, especially along the borders of the orogen,

some effects of the Hudsonian Orogeny. There were also some epeirogenic movements and erosion following the development of the present craton. Diabase dykes of three different ages and trends intrude the rocks of Slave Province (Fahrig and Wanless, 1963). The oldest dykes have an easterly trend and appear to be most abundant at Yellowknife and east of it, in the southern third of the craton (Pl. IV-7). Their age is about 2,150 m.y. A second group of dykes trend in a northeasterly direction and have been dated at about 2,100 m.y. At Indin Lake about 120 miles north of Yellowknife, a set of northwesterly and northeasterly trending, steeply dipping dykes are dated at about 2,100 m.y. These two sets of dykes appear to be restricted to Slave Province. The third set of dykes, the Mackenzie dyke swarm, trends north-northwesterly and occurs not only in Slave Province but also in the adjacent parts of Bear and Churchill Provinces. These dykes have been dated at about 1,200 m.y.

Archean rocks and rocks formed during the Kenoran Orogeny are cut by faults of several ages and orientations (Fig. IV-12). These faults are marked by narrow zones of gouge and mylonite within wider zones of crushed and brecciated rock; some are quartz veins and narrow stockworks. In the Indin Lake area, numerous vertically dipping faults trend northwesterly, northeasterly, or northerly. Largest displacements occur along the northwesterly faults that have left-hand strike slip. The diabase dykes that are dated at 2,100 m.y. have similar orientations and occur in fractures parallel with the faults. A few lie in the faults, and some are slightly sheared which suggests a close time relationship. A prominent set of faults in the southern and northern parts of the province strikes northeast to east. Those in the south are roughly parallel with the oldest diabase dyke swarm and could be related in time; most have right-hand strike-slip displacements. These older structures may be related to the development of the bordering Aphebian geosyncline. In the northern part of the province the northeast faults also cut Aphebian and Helikian rocks and are associated with northerly and northwesterly trending faults. They are probably about 1,200 to 1,300 m.y. old; those with the same orientation farther south are thought to be older. The most prominent fault set in the province strikes north-northwest to north and includes such well-known structures as the West Bay and other late faults near Yellowknife. Faults of this set extend well beyond the border of the province and cut rocks as young as Helikian. The main movement is left-hand strike slip, the horizontal component of which may be several miles. Within Slave Province these faults apparently are most abundant in the western part, and where they are numerous cross-faults of varying orientations connect the major structures. The Mackenzie diabase dyke swarm dated at about 1,200 m.y. is parallel with the fault system. Some dykes pass into the faults and locally are displaced by cross-faults related to the major structures. Accordingly, they have formed at about the same time. These faults and also the related

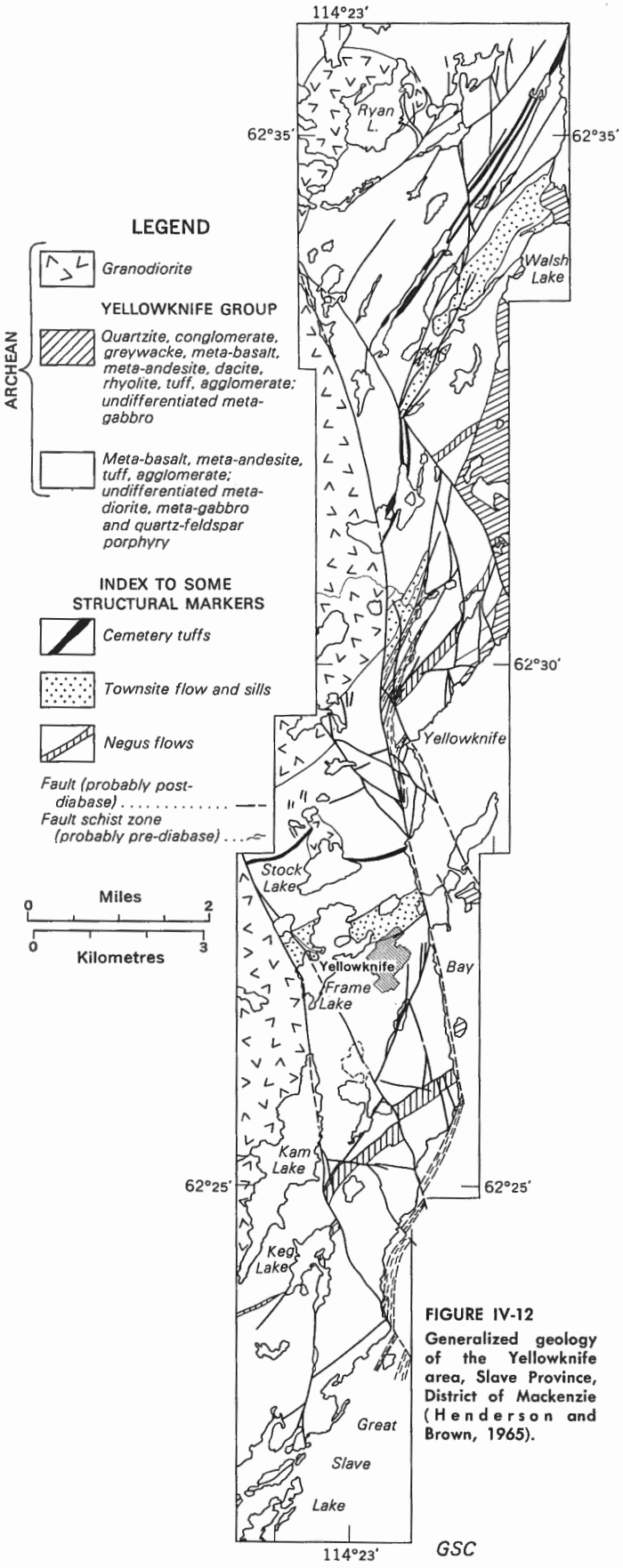

FIGURE IV-12
Generalized geology of the Yellowknife area, Slave Province, District of Mackenzie (Henderson and Brown, 1965).

diabase are later than all other faults in Slave Province including the northeast-trending faults in the northern section.

Important modifications of the rocks of Slave Province were caused by phases of the Hudsonian Orogeny. One, or possibly two, roughly circular masses of alkaline rocks that include nepheline- and sodalite-bearing syenite and possibly carbonatite were intruded during the Hudsonian Orogeny into the Kenoran granitic rocks at Big Spruce Lake, about 20 miles east of the boundary between Bear and Slave Provinces. Biotites from these rocks give ages of about 1,750 m.y. Many K–Ar age determinations have been made on rocks from northern Slave Province. Those for biotite give anomalous dates, less than the expected Kenoran ages, that range from about 1,800 to 2,300 m.y. In some samples muscovite that coexists with the anomalously young biotite gives Kenoran ages. These data suggest that the old rocks were heated during the Hudsonian Orogeny to an extent sufficient to drive argon out of some of the micas.

Other effects of the Hudsonian Orogeny are restricted to the borders of Slave Province. Along the western boundary of the province, Archean rocks have been involved in the folding of the younger Aphebian rocks. Much of the eastern border zone may be an extensive zone of shearing.

Bathurst Subprovince

Most of Bathurst Subprovince is underlain by the Goulburn Group that comprises a conformable sequence, about 15,000 feet, divided into five formations (Fraser, 1964). The sediments have been moderately folded about axes that strike north-northeast. Folding is more intense along Bathurst Inlet where fold axes strike in a northerly direction. The Goulburn Group lies unconformably on the Archean Yellowknife Group and Kenoran granites. These are cut by gabbro sills and diabase dykes that yield K–Ar whole rock ages of about 1,200 m.y. The basal formation of the Goulburn attains a maximum thickness of 3,500 feet and comprises, from the base up, grey to green quartzite and grit, dolomite with stromatolitic zones, and interbedded quartzite and maroon to grey argillite with local silty layers and stromatolitic dolomite. These rocks are overlain by as much as 7,500 feet of thick-bedded, massive, pink to red feldspathic quartzite and orthoquartzites with thin interbeds of conglomerate. These rocks are succeeded by about 2,000 feet of purplish green or grey argillites with thin interbedded grey limestone and dolomite overlain by about 800 feet of grey to pink dolomite with abundant stromatolitic zones. The uppermost formation is composed of interbedded red-grey or maroon siltstone and argillite with red crossbedded arkose near the top. The arkose contains granules of specularite.

The age of the Goulburn Group is not precisely known but it is thought to be Aphebian and to have been folded during a phase of Hudsonian Orogeny. The sediments are typical of those deposited in a shallow-water environment. Both marine and continental types occur in the sequence and several transgressions are recorded suggesting some tectonic instability during sedimentation. Formations are thickest near Bathurst Inlet and thin rapidly to the west and southwest. In the extreme western exposures near Contwoyto Lake only the lowest three formations are present and the total thickness is little more than 2,000 feet. The quartzite unit thins from 7,500 feet to 600 feet. The Goulburn Group is similar to the Epworth Group and a tentative correlation of these two sequences is suggested. J. A. Fraser points out that the basal unit of the Goulburn is lithologically similar to the lowest Epworth unit and part of the lower Epworth carbonate formation. The Goulburn quartzite formation is not well developed in the Epworth and may be absent as it thins rapidly to the west. It may be a facies of part of the lower Epworth carbonate unit. The upper three formations of each group represent similar lithologies and environments. The westward thinning of the Goulburn strata and distribution of basal sands along the eastern margin of the Epworth outcrop area suggest that the intervening basement may have been an arch during the time of deposition.

The Goulburn Group is overlain unconformably by cream, red, and buff quartzites of the Tinney Cove Formation that includes thin conglomerate beds with boulders of underlying strata. Kaolinite occurs in the matrix. Except near faults these rocks are very gently deformed. They are overlain, probably conformably, by the Parry Bay Formation that consists of at least 550 feet of thick-bedded, fine-grained, grey or buff dolomite and limestone with stromatolitic zones and minor amounts of interbedded brown dolomite and red or green shale and shaly dolomite. On islands in the mouth of Bathurst Inlet, but nowhere on the mainland, rocks of the Kanuyak Formation lie with slight angular unconformity on Parry Bay strata. Reddish brown dolomite and calcareous shales are the dominant lithologies. Basalts of the Coppermine Group with overlying sediments disconformably overlie rocks of the Kanuyak and Parry Bay Formations and are separated from older strata by an unconformity.

The Tinney Cove quartzites are lithologically similar to Hornby Bay quartzites and the Thelon Formation of the Dubawnt Group. On the basis of lithologic similarity and stratigraphic position, the Parry Bay dolomite is tentatively correlated with the Hornby Bay dolomite and the post-Thelon carbonate. Rocks of the Kanuyak Formation have no known equivalents in the region. These rocks are separated from Aphebian rocks deformed during a phase of the Hudsonian Orogeny by an angular unconformity and are cut by basic dykes and overlain by lavas that date from 1,000 to 1,400 m.y., indicating that the sediments are Paleohelikian in age.

Tectonic Summary

Bear Province is divided into four subprovinces. The Wopmay Belt contains the oldest rocks—Aphebian sedimentary and volcanic rocks that were folded and invaded by granitic ones during the Hudsonian Orogeny. Coppermine, Brock, and Minto Subprovinces are underlain by Helikian and possibly Hadrynian rocks that have been intruded by gabbro sills and dykes and are only gently folded.

The Aphebian rocks reflect two distinct sedimentary environments. The Snare and Epworth Groups are composed of miogeosynclinal shelf sediments deposited under shallow-water environments. They comprise quartzites, dolomites, shales, and siltstone–shales that occur in northerly trending belts along the margins of Slave Province, a craton of Archean rocks. These sediments were deposited in a geosyncline that probably also trended northerly. Snare sediments are overlain by basic volcanic rocks some of which are pillow lavas. These rocks are in a zone as much as 75 miles wide along the western margin of the craton. Farther west, the Echo Bay and Cameron Bay Groups of red beds and acidic volcanic rocks occur in north to east-northeasterly trending belts. These rocks are a late tectonic complex of molasse-type sediments consisting of conglomerate, arkose, red siltstone and shale, and volcanic piles of andesitic to trachytic compositions that contain great thicknesses of tuffs, breccias, and ignimbrites as well as flows. Intrusive equivalents of these volcanic rocks are common. Sedimentary rocks occur in basins that may be partly fault controlled. The Aphebian rocks are folded and intruded by granitic rocks; the intensity and complexity of the deformation in general increases away from the margin of the craton. In some areas the sediments have been converted to granitic gneisses after passing through zones of increasing metamorphism and the granitic intrusions are generally bordered by migmatites. The Cameron Bay rocks are less intensely deformed and metamorphosed and cut by high-level, late tectonic granitic rocks.

Within the part of the geosyncline affected by early orogenic events local zones of Archean basement may exist, but so far have not been identified except near its eastern margin. Likewise in the region south and east of Great Bear Lake in that part of the orogen involved in late tectonic activity, zones of granitic gneiss are present that probably were formed by earlier phases of the orogeny. Such rocks may be, in effect, basement to Echo Bay and Cameron Bay sedimentary and volcanic rocks. North of Great Bear Lake the deformed rocks are overlain unconformably by Paleohelikian sandstones and dolomites of the Hornby Bay Group which is conformably overlain by the thick Coppermine Group, a sequence of basalt with some red beds, succeeded unconformably by quartzite, dolomite, and shale. The Coppermine lavas are probably Neohelikian and the younger sediments may be either Neohelikian or Hadrynian. On Victoria Island the Neohelikian or Hadrynian Shaler Group of sediments and basalt rests unconformably on a basement of Archean rocks. These rocks are very gently folded about easterly to north-northeasterly axes. Faults, probably of different ages, and diabase dykes and sills of several ages cut the Hornby Bay, Coppermine, and Shaler rocks. The Muskox Intrusion, a differentiated gabbroic mass, intrudes Hornby Bay strata and is probably a little younger than the Coppermine flows. The Proterozoic rocks are overlain unconformably by Cambrian and Ordovician sediments that are broken by a few small normal faults.

Wopmay Subprovince

Aphebian

Aphebian rocks in Wopmay Subprovince are divided into the Snare, Epworth, Cameron Bay, and Echo Bay Groups. The Snare and Epworth rocks are separated by an area of granitic rocks. The Cameron Bay rocks are not in contact with the Epworth but may be in contact locally with Snare rocks.

Snare rocks occur in north-trending belts along the margin of Slave Province and as isolated patches within younger granitic rocks to the west. At Basler and Kwejinne Lakes, and to the southwest, Snare rocks lie with marked angular unconformity on the Archean Yellowknife Group and granitic rocks (Lord, 1942). Farther north the structural unconformity becomes less obvious due to gradual elimination of the angular discordance as a result of greater intensity of folding of the Snare rocks. The Snare Group consists of quartzite, dolomite, shale, and siltstone; the relative amounts of these sediments varies throughout the geosyncline. At Basler Lake and to the southwest the basal rocks are white or light grey, crossbedded quartzite with a few thin beds of shale. Near the base the quartzite is feldspathic and contains a few lenses of quartz-pebble conglomerate. At the base it contains a few fragments of basement rocks. Thickness varies from less than 100 to about 1,000 feet. At Basler Lake, a thin shale member near the top of the quartzite is overlain by sandy dolomite at least 300 feet thick. The quartzite at Kwejinne Lake is overlain by black shale, and west and southwest of Basler appears to be overlain by dark, finely laminated shale that contains another quartzite. North of Basler Lake the basal quartzite is overlapped by the dolomite that can be traced along the unconformity to the north end of the belt of sediments. The dolomite is overlain by alternating thin beds of siltstone and finely laminated shale of unknown thickness, and is probably several thousand feet thick. In many

places, the siltstone and shale beds are each less than an inch thick and the rock has a varved appearance. Locally the couplets increase in thickness to 3 or 4 inches. In the thicker couplets graded bedding and fine cross-laminations are evident. Within these rocks are units of quartzite, dolomite, and shales, each several hundred feet thick, the number and stratigraphic position of which are unknown because of the complexity of the structure. Some of the quartzite and dolomite units grade along strike into siltstone and shale.

The transition zone with dolomite is marked by zones in which calc-silicate lenses are numerous. West and south of Basler Lake quartzite, shale, and dolomite are present in various stratigraphic successions, whereas to the north metamorphosed equivalents of the siltstone–shale with minor quartzite or dolomite are dominant. At Grant Lake, the metamorphosed rocks are overlain conformably by basic lavas some of which are pillowed and that contain intercalated thin bands of cherty sediments and Snare-like sediments. Pink or red feldspathic sandstone and red or grey shale overlie the volcanic rocks and are tentatively included in the Snare Group. Snare rocks are of miogeosynclinal facies, deposited mainly in shallow water. The well-sorted, crossbedded quartzites suggest deposition in a shallow-water, rather turbulent, near-shore environment. The sandy nature of the dolomite also suggests a similar environment. The thick, finely laminated shales west of Basler Lake indicate a somewhat deeper-water environment. The facies change north of Basler to a sequence of siltstone and shale, and the wedging out of basal sands may also result from deposition in a somewhat deeper-water environment. In the original geosyncline near-shore conditions appear to have prevailed in the southern and southeastern parts, whereas, in the northern section, sediments were deposited into a slightly deeper-water, off-shore environment. Extrusion of basic submarine lavas in the north also indicates a transition to deeper-water environment.

The Epworth Group contains a conformable sequence of sedimentary and volcanic rocks with a thickness of at least 15,000 feet. The group lies on Archean volcanic rocks and granites with marked unconformity. Fraser (1966) has established five formations. The lowest consists of 2,000 to 3,000 feet of grey and pink quartzite with some beds of argillite and quartz-pebble conglomerate. It may be absent locally. The succeeding 2,000 to 5,000 feet is stromatolitic dolomite (Pl. IV-8) with a few interbeds of argillite. This formation locally forms the basal unit of the group. Thinly bedded, siltstone argillite and greywacke with thin bands of quartzite and dolomite, at least 3,000 feet thick, overlie the dolomite and are in turn overlain by 3,500 feet of limestone with interbedded grey and red argillite. The youngest formation, 1,200 feet thick, comprises red mudstone and siltstone that grades upwards into feldspathic, calcareous sandstone. Pillowed and massive andesite and related fragmental rocks are locally present within the sediments.

The siltstones and shales of the middle formation are metamorphosed to phyllites and andalusite-bearing or garnetiferous biotite schists where intruded by granite in the west. The metasediments also occur as patches within the granites and can be traced southward to where similar rocks are considered to be part of the Snare Group. It seems reasonable to assume that the middle unit of the Epworth Group is the stratigraphic equivalent of the Snare siltstones and shales and that the underlying dolomites and quartzites are also equivalent. The upper two Epworth formations may be equivalent to similar strata that lie above Snare volcanic rocks north of Grant Lake. The Epworth and Snare rocks appear to be approximately equivalent in time and to have been deposited in a single northerly to north-northeasterly trending geosyncline in water depths above wave base and probably just below it. The basal quartzites are probably wedge-shaped, near-shore deposits along the margin of the basin. Interbedding of shales and quartzites in the southern part of the geosyncline, and quartzite and dolomite strata within the siltstone shales in the north suggest some moderate tectonic instability during deposition of these sediments. Preservation of the thin silt–shale alternations, fine-lamination, cross-lamination, and graded beds suggests that deposition of the shale–siltstone sequences in quiet water below wave base took place.

The Cameron Bay and Echo Bay Groups are structurally conformable but may be separated by a disconformity. The Echo Bay rocks at Port Radium are usually separated into two units. The lower unit comprises tuff, green and red argillite, chert, arkose, and conglomerate with a few andesite flows and thin limy beds. Mudcracks, ripple-marks, and graded bedding occur in the arkose and shale. Conglomerates are present at several horizons. The upper formation consists of porphyritic andesite flows, breccia, tuffs, and agglomerates. The Cameron Bay rocks form a conformable sequence of conglomerate, purple and maroon arkose, siltstone, and shale. Pebbles in the conglomerate consist mostly of porphyritic volcanic rocks along with chert, argillite, and sandstone and occur in an arkosic matrix. About 30 miles north of Port Radium (Feniak, 1952) green and maroon, generally porphyritic andesite, dacites and trachytic flows, and agglomerate, and tuff are intercalated in typical Cameron Bay sediments. The volcanic rocks are similar to those of the Echo Bay Group at Port Radium.

Elsewhere in Bear Province, a separation into the two groups has not been made. For the most part the undivided volcanic complexes consist of maroon, green, or brown, commonly porphyritic andesites, dacites and trachytes with related tuffs, ignimbrites, and breccias. Included with the volcanic rocks, and often difficult to distinguish from them, are their hypabyssal, lithologically similar, intrusive equivalents. Usually associated with the volcanic rocks are sediments that form local, northerly trending belts. Typically these rocks consist of conglomerate, red, maroon, and purple arkose or feldspathic



PLATE IV-8
Stromatolites in dolomite of Aphebian Epworth Group, Takiyuak Lake, Bear Province, Northwest Territories.

sandstone, and shale. The clasts in the conglomerate are largely subrounded pebbles of the volcanic facies of the Cameron Bay and Echo Bay Groups, with some quartz, shale chips, and pink granodiorite or granite. The conglomerate occurs at several horizons within the sequence and in places forms the basal part of thick graded beds, the upper part of which is arkose. Crossbedding and chips from interbedded shales are abundant. The shales interbedded with the arkoses are red or maroon and commonly display mudcracks and sandstone dykes. The sedimentary sequences have been found below, above, and intercalated with the volcanic rocks. Exposed thicknesses of the various sequences vary from a few hundred to possibly as much as 10,000 feet, but precise correlation between the scattered exposures of these rocks is not possible. Basement to these rocks has not been identified with certainty. Some of the sedimentary basins may have been in part fault controlled.

The Cameron Bay and Echo Bay Groups are fundamentally different from the Snare and Epworth Groups, having formed in very different environments. The sediments, being largely red bed conglomerate and arkose were probably deposited in an environment ranging from continental to shallow marine. The mudcracks, shale chips, and scour channels suggest periodic subaerial exposure of the sea floor. The clastic units are products of rapid sedimentation from areas of considerable relief that are tectonically unstable. The volcanism that occurred at intervals consisted of andesite and trachyte flows and large amounts of tuffs and breccias from explosive activity. Rapid erosion of these volcanics produced many of the pebbles in the conglomerates and probably a considerable amount of the finer clastic material. Shallow intrusions related to the volcanic rocks cut both volcanic and sedimentary rocks. The sedimentary and volcanic rocks probably accumulated during a late tectonic stage of geosynclinal evolution, after the older miogeosynclinal Epworth and Snare rocks had been deformed in the early phases of the Hudsonian Orogeny.

Hudsonian Orogeny

During the Hudsonian Orogeny Aphebian rocks were folded, metamorphosed, and converted to granitic rocks, or intruded by them. Micas from these granitic rocks yield

K–Ar dates that range from about 1,700 to 1,870 m.y. The sequence of orogenic events is not known with certainty as the precision of the methods of dating is not great enough to isolate the various phases of the orogeny.

Folding of the Snare Group took place in two phases (Ross and McGlynn, 1965). The earlier and most important phase resulted in northerly to northeasterly trending folds. In the vicinity of Basler Lake, where deformation and metamorphism were not intense, and where competent thick quartzites occur in the succession, the folds are concentric, shallow, and open, plunging gently north, with northerly trending axial planes that dip steeply east. To the west, as the Snare rocks thicken, the trend of the folds gradually changes to northeast. North of Basler Lake, the folding is more intense and similar folds are present. The difference in style of folding may be partly due to facies changes in the sediments as the thick competent quartzite members grade to shale, but the change also coincides with an increase in metamorphic grade and probably is also related to more intense deformation at a deeper structural level. The axial planes strike north and dip steeply to the east; the folds plunge gently to the north. Andalusite and cordierite metacrysts developed during the metamorphism stage. The second phase folds are similar and have axial planes that strike southeasterly and dip steeply both southwest and northeast. They plunge at moderate angles to the northwest. Locally, they are related to the intrusion of the granitic rocks. The deformation affected basement rocks near the unconformity altering older structures within the Archean rocks and causing faulting and the formation of a schistosity in older granitic rocks. Early folds in Yellowknife rocks underwent further closure or were refolded about axial planes parallel with those in Snare rocks. The Epworth sediments are very gently folded along their eastern margin. To the west they are thrown into increasingly tight folds that trend northerly.

Aphebian rocks were converted to granitic gneisses and intruded by granitic rocks during the Hudsonian Orogeny. These plutonic events probably took place during the initial folding or just after it, and during the second phase of folding. During the intrusion of the granitic rocks and folding, Snare and Epworth rocks were metamorphosed to the point where andalusite, cordierite, and locally sillimanite metacrysts were developed. Gneissic to almost massive grey and pink granodiorites or quartz monzonites are the most abundant lithology in the granitic batholiths. These rocks grade into zones of varying widths of migmatites and mixed granitic gneisses that in turn grade to metamorphosed Snare or Epworth rocks. East of Wopmay River a broad zone of highly metamorphosed sediments, migmatites, and mixed gneisses intruded by a few, small granodiorite batholiths may represent the deepest level within the orogen. Farther south, east of Basler Lake, the grade of metamorphism of the sediments is lower and zones of migmatite rare. These factors along with the style of folding of sediments indicate that in this area a higher structural level within the orogen is exposed.

The phase of emplacement of granitic rocks was probably followed by a period of faulting, volcanism, and deposition of molasse-type sediments of the Cameron Bay and Echo Bay Groups. These sediments and the acidic volcanic rocks as well may have been deposited in basins bounded by faults such as the one along Wopmay River. This is a major, post-orogenic linear feature represented by zones of mylonite in the granitic rocks and by strong shearing in the sediments of the Snare Group. The Cameron Bay and Echo Bay Groups were intruded by feldspar porphyries of variable composition that probably are the intrusive equivalents of the volcanic rocks comprising the groups. These rocks are also intruded by nearly massive, hornblende-bearing, coarsely porphyritic granodiorites, and locally by massive pink granodiorite. Micas from these intrusive rocks yield K–Ar ages that range from 1,700 to 1,780 m.y. The latter intrusions are a high-level type with sharp contacts and narrow metamorphic aureoles. They are concentrated west of the Wopmay River in the area where the late geosynclinal and late tectonic Cameron Bay and Echo Bay rocks were deposited. A two-part division of granitic rocks, therefore, seems apparent. The oldest are syntectonic, rather deep-level granodiorites and attendant migmatites and gneisses that involve the Snare and Epworth rocks; these are followed by younger, late tectonic and possibly post-tectonic, high level granodiorites and granites that cut the Cameron Bay and Echo Bay and, to some extent, older rocks. The older intrusions are concentrated in the original miogeosynclinal environment whereas the younger ones occurring essentially in the late geosynclinal environment are probably closely related in time to the volcanic rocks.

Coppermine and Brock Subprovinces

Helikian

The Hornby Bay Group is best exposed north of Hornby Bay, Great Bear Lake. The basal units lie with angular unconformity or nonconformity on a deeply eroded surface underlain by rocks of the Epworth, Cameron Bay, and Echo Bay Groups and Hudsonian granitic rocks that cut these Aphebian sediments. Two formations are included in the group: a lower quartzite and conglomerate and an upper dolomite (Fraser, 1960). North of Great Bear Lake the lower unit may be as much as 4,000 feet thick but towards Coronation Gulf to the northeast it thins and finally disappears. The conglomerates are made up of pebbles of quartzite, quartz, and some quartz feldspar porphyry set in a white quartzite matrix. They are present as thin beds or lenses throughout the sequence of quartzites. The quartzites at the base of the succession are commonly thin bedded, reddish or purple, crossbedded, and ripple-marked. More typically, however,

the quartzites are white to light grey or pink, often cross-bedded, moderately coarse bedded, and coarse grained. They are composed of well-rounded grains of quartz with minor chert in a quartz cement. The upper unit of the group overlies the quartzites conformably. It consists of a shaly dolomite overlain by brown or reddish dolomite, and a buff to grey, finely crystalline dolomite with numerous stromatolitic zones. North of Great Bear Lake the dolomite is about 4,000 feet thick, but to the northeast towards Coronation Gulf it thins, and, where the lower unit is absent, lies unconformably on Epworth rocks and granite. To the west the dolomite is overlain probably unconformably by the early Paleozoic. Hornby Bay rocks are very gently dipping to horizontal except near some faults. The rocks appear to be folded into broad folds about northeast-trending axes.

The Hornby Bay rocks are conformably overlain by the Coppermine Group that on the basis of present knowledge is divided into two formations. The lowest formation is made up of basalt flows estimated as between 11,000 and 13,000 feet thick. The flows are brown, dark green, or purplish, fine-grained basalt and their composition appears to be fairly uniform throughout the sequence. Individual flows vary between 50 and 200 feet thick. They are largely massive but commonly the upper parts, and rarely the lower, are amygdaloidal. Calcite, quartz, chlorite, epidote, and locally native copper or chalcocite occur in the amygdules. The lavas dip very gently to the north at angles of 10° or less. Eight K–Ar age determinations on whole rock samples of these lavas yield dates that range from 735 to 1,200 m.y. There is no stratigraphic significance to this range of dates as some young ages have been obtained from flows near the base of the succession. Directions of natural remnant magnetization were measured on fifty-one oriented samples from twenty-four sites in these flows (W. A. Robertson, 1964), and yielded a pole position of 1°N and 176°W, placing the flows at a paleolatitude of 10°–15°. This pole position is very close to that of the Muskox Intrusion which is reasonably well dated at 1,150 m.y. It seems likely therefore that the Coppermine lavas are about 1,100 to 1,200 m.y. old and can be considered to be early Neohelikian.

The younger formation of the Coppermine Group comprises about 15,000 feet of sediments. Along Coppermine River the basal units are red sandstone, siltstone, and shale (Fraser, 1960). These rocks overlie the lavas conformably and may be locally intercalated as thin bands within the lavas. The red beds appear to be restricted to the Coppermine River and to be bevelled to the east and west. The basal units are overlain by grey shale which, according to Baragar (1967), is separated from the older rocks by a slight angular unconformity. The shale is overlain by buff, grey, or brownish dolomite with quartzite and shale beds near the top. North of Rae River, stromatolitic zones are present in a black limestone that may occur within the dolomite or overlie it. The stratigraphy of the Coppermine sediments underlying Brock Plain is not well known. The lowest beds are grey to black shales that probably are equivalent to those above the unconformity on Coppermine River. The shales pass upwards into a conformable sequence of pink, grey, and buff dolomite, and two beds of quartzite. Gypsum as much as 65 feet thick occurs in the carbonate unit on the Arctic coast and stromatolitic black limestone similar to that north of Rae River is found near Deas Thompson Point.

The basal red beds of the Coppermine sediments, because they are conformable with the lavas, and possibly interbedded with them, are probably closely related in age and may be Neohelikian. The sediments above the unconformity are cut by diabase sills from which three K–Ar whole rock age determinations yield dates that range from 445 to 718 m.y. They may be Neohelikian or Hadrynian.

Muskox Intrusion

The Muskox Intrusion is a layered basic intrusion that is dyke-like in plan and funnel-shaped in cross-section (Fig. IV-13). The intrusion cuts a basement complex of metamorphosed Epworth rocks, migmatite, granitic gneiss, and granodiorite, and the overlying quartzites of the Helikian Hornby Bay Group. The intrusion is not in contact with Coppermine rocks but the lavas and possibly some of the sediments probably formed the cover rocks. Age determination on micas and whole rock samples indicate an age for the Muskox Intrusion of 1,100 to 1,200 m.y. The remnant magnetism of the rocks has a pole position of 4°N, 175°W (W. A. Robertson, 1964). This position is very close to that of the Coppermine lavas and together with the radiometric data suggests time equivalence. Radiometric and paleomagnetic data also indicate that the diabase dykes of the Mackenzie swarm were emplaced at roughly the same time (Fahrig, *et al.*, 1965). There was accordingly an important period of basic magma intrusion in Bear Province in the early Neohelikian. According to Smith (1962), the shape of the intrusion is controlled to a large extent by the unconformity at the base of the Hornby Bay sediments, which dammed the magma moving up the northerly trending feeder dyke, causing it to spread out into a funnel-shaped form. A large volume of magma accumulated in which differentiation could take place. The feeder dyke is nearly vertical whereas the contacts of the funnel-shaped portion of the mass dip inwards at 57° in the south part, gradually decreasing to the northeast to 22°.

The intrusion has been divided into a number of structural and petrological units. The feeder forms the southern half of the mass and is about 37 miles long and from 500 to 1,800 feet wide. Bronzite, gabbro, and norite are the most abundant rock types, but as the dyke widens lenses of picrite occur and finally form an almost continuous zone in the centre of the dyke. The marginal zone of the intrusion is present in the northern part and varies in width from 200 to 1,200 feet. It consists of a gradational series of lithological types that, from the contact

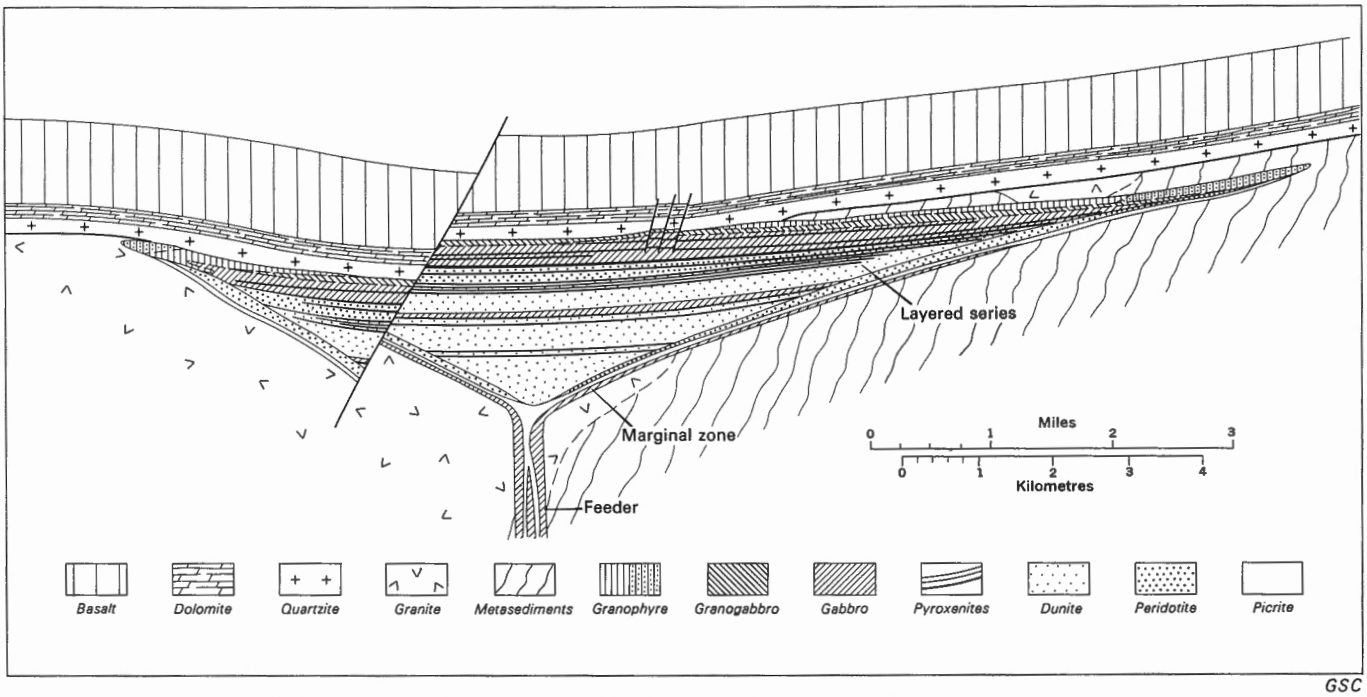

FIGURE IV-13. Typical cross-section of the Muskox Intrusion (by T. N. Irvine and C. H. Smith).

inward, include bronzite, gabbro, picrite, feldspathic peridotite, peridotite, and then dunite in places. The contact gabbro is chilled locally and thins northward to zero. The central layered series comprises alternating layers of dunite, peridotite, pyroxenites, and gabbro. Dunite is most abundant in the lower part of the series and gabbro in the top. The sequence is about 8,500 feet thick and is composed of thirty-eight main layers some of which are banded. The layers dip northwest at 10° about parallel with the dip of Coppermine flows. The layers continue to within 1,200 to 2,000 feet of the margin of the intrusion where they steepen slightly and then die out. The upper border zone is about 200 feet thick and consists of granophyre-bearing gabbro at the base that grades upwards into granophyre. Intrusive breccia is common along the top of the border zone, and consists of variably sized fragments of Hornby Bay quartzites and some fragments of the older basement rocks set in a granophyre matrix.

Diabase Dykes and Faults

The oldest known diabase dykes that cut rocks of Bear Province are northerly to north-northwesterly trending, vertical-dipping dykes of the Mackenzie swarm whose age is in the range 1,100 to 1,300 m.y. (Pl. IV-9). These dykes cut the Hornby Bay Group and probably also the Coppermine lavas. Dykes trending north of east are south of Great Bear Lake and possibly extend northeast to where they cut Epworth rocks. They have been tentatively dated at 875 m.y. (Fahrig, *et al.*, 1965). A few north-northeast striking dykes that occur south of the Coppermine lavas and also cut Coppermine sediments yield K–Ar dates that average 700 m.y. These latter rocks may be related to diabase sills in the Coppermine sediments that have similar dates. The Mackenzie dykes yield dates that compare with those of the Coppermine lavas and Muskox Intrusion, and paleomagnetic data from the three types of basic rock have given similar pole positions. Thus a major period of basic magma intrusion and extrusion took place in the middle Helikian. It was followed by a second period in the late Hadrynian, represented by diabase dykes, sills, and probably also the Natkusiak volcanic rocks.

Faults cut rocks of all ages in Bear Province but appear most common in rocks older than the Coppermine Group. Most strike northeasterly and northerly but some trend northwest. All are nearly vertical and marked by narrow zones of gouge, breccia, and mylonite. Quartz veins occur in related fractures and the "giant quartz-veins"—broad, long quartz stockworks—are localized along some faults, particularly those with a northeast trend. They contain massive quartz or crystal-lined vugs that are fractured and veined with younger quartz. The wall-rocks are silicified and cut by narrow, braided quartz veins. On some stockworks later faulting took place along one or both contacts. Localization of the stockworks is partly controlled by slight variations in strike, and possibly also dip, of the faults (McGlynn, 1957). In northern Bear Province, north- and northeast-trending faults displace the lavas and lower sediments of the Coppermine Group, as well as Muskox Intrusions and some diabase dykes of the Mackenzie swarm. However, many Mackenzie dykes are

known to be post-faulting and some cut the quartz stockworks. Either there are two ages of faults with the same orientation, or, more probably, the faulting and the intrusion and extrusion of basic magma are related events reflecting a period of tension during mid-Helikian times. It seems likely that many of the north- and northeast-trending faults in the southern part of the province are related to those in the north. However, there is movement on some of these faults which is later than the diabase dykes that are tentatively dated at about 875 m.y. Some old faults may have been reactivated at this time or new northerly and northeasterly trending faults may have formed at about the time of intrusion of the younger diabase dykes.

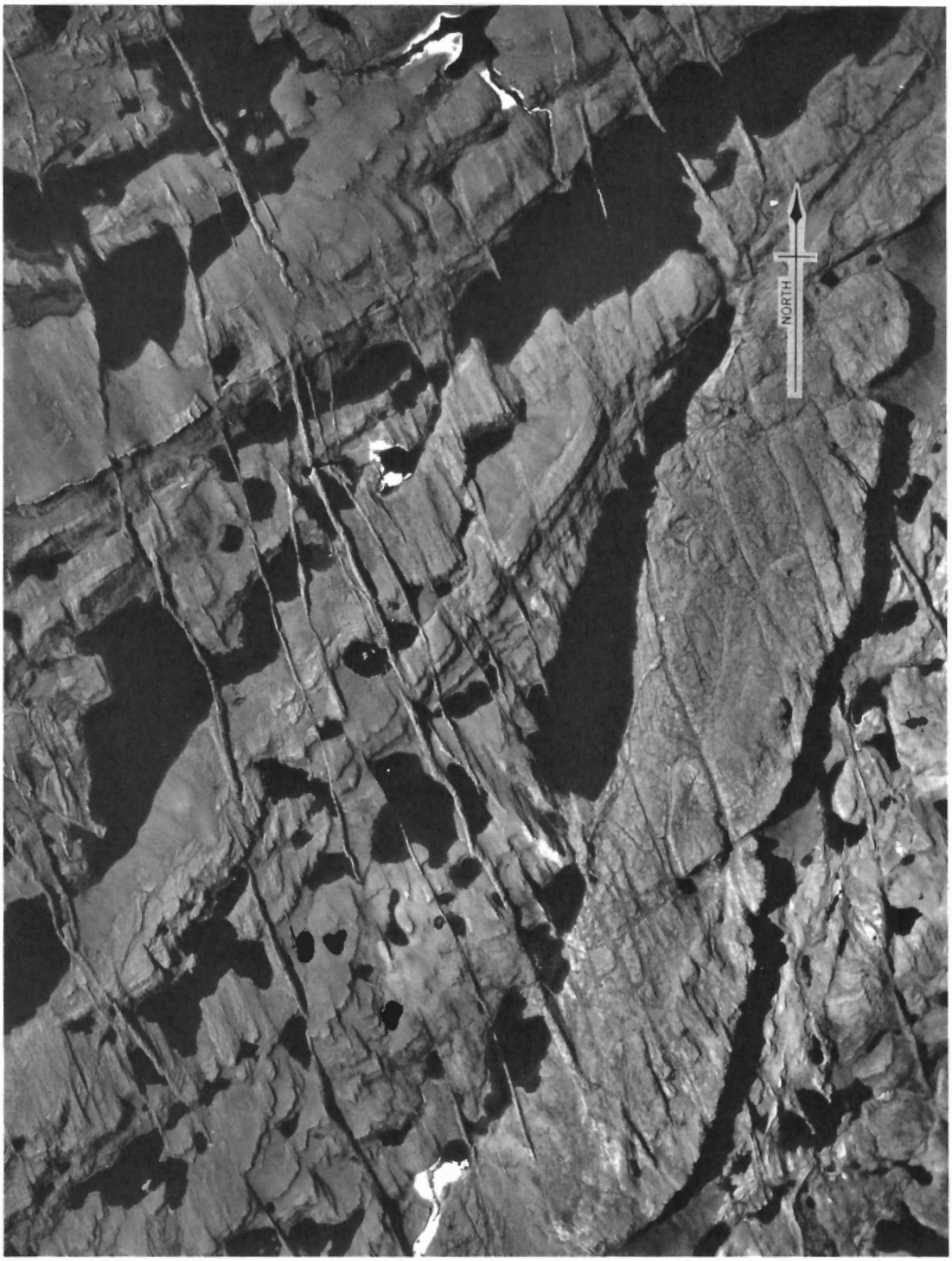

PLATE IV-9

Vertical air photo of northwest-trending diabase dykes of Mackenzie swarm cutting gently folded strata of Aphebian Epworth Group, Bear Province, Northwest Territories.

Minto Subprovince

The Precambrian rocks on Victoria Island range in age from the Archean to Hadrynian (Thorsteinsson and Tozer, 1962). The oldest rocks are exposed at the head of Hadley Bay and consist of north-northeasterly striking, moderate- to steep-dipping quartzite, schistose quartzite, and granodiorite thought to be younger than the sediment. Biotite from the latter rock yields a K–Ar date of 2,405 m.y., a Kenoran granite. The quartzite accordingly is Archean if cut by the granodiorite.

The thick conformable succession of the Shaler Group unconformably overlies the older rocks and is overlain disconformably by basic volcanic rocks of the Natkusiak Formation. The Shaler Group is intruded by numerous gabbro sills and dykes. The basal Glenelg Formation is divided into two members, the oldest of which consists of 2,000 feet of red, white, or grey, thin- to thick-bedded quartzite and sandstone, alternating with dark grey concretionary shale, siltstone, and very fine grained dolomite. The upper member comprises about 1,200 feet of thick-bedded, grey to light red quartzites with a thin unit of thick-bedded, reddish orange or brown, stromatolitic dolomite at the top. Tentatively correlated with the Glenelg are red, coarse-grained quartzites with quartz-pebble conglomerate on Richardson Islands and on the south shore of Victoria Island. On Banks Island, grey thin-bedded limestones, cherty limestone, and chert similar to the lower Glenelg in the type area are overlain by red sandstone or quartzite with minor grey shale and siltstone lithologically similar to the upper Glenelg.

The Glenelg Formation is overlain conformably by the Reynolds Point Formation that comprises a basal unit of grey, medium-bedded sandstone succeeded by 1,900 feet of grey, medium-bedded, argillaceous limestone with minor, interbedded sandstone, shale, and stromatolitic limestone. The conformably overlying Minto Inlet Formation varies from 300 to 1,200 feet in thickness. It is composed of white, thin-bedded gypsum and anhydrite with some grey or red sandstone, greyish green limestone, grey shale dolomite, and siltstone. The Wynniatt Formation rests with sharp contact on the Minto Inlet. It is composed of 2,700 feet of dark grey to black, thin-bedded limestone and argillaceous limestone with several stromatolitic and dolomitic horizons. The conformably overlying Kilian Formation is a mixed assemblage of thin-bedded, multicoloured sediments, about 600 feet thick. The lower member is composed of white, grey, or pink gypsum and anhydrite, grey gypsiferous or calcareous shale and reddish brown sandstone, siltstone and dolomite. Crossbedding, ripple-marks, and desiccation fractures are common, indicating a shallow-water environment of deposition. The upper member is similar but contains no gypsum or anhydrite. In the west the uppermost beds are buff, cross-bedded sandstone and conglomeratic sandstone that was eroded before the succeeding Natkusiak volcanics were extruded.

The Natkusiak Formation lies disconformably on the Kilian Formation. It has a maximum thickness of 1,000 feet and consists largely of basalt with minor amounts of tuff or agglomerate. Individual flows are as thick as 100 feet and are commonly massive with flow-top breccias and thin amygdaloidal zones at the base and top. Numerous diabase sills intrude all formations of the Shaler Group and possibly also the Natkusiak Formation. They are as much as 100 feet thick and extend for tens of miles along strike. They are probably closely related in time to the extrusion of the Natkusiak volcanic rocks. Whole rock K–Ar determinations from two samples from sills yield dates of 635 and 640 m.y. and those from three samples of the Natkusiak volcanics range from 390 to 483 m.y. These latter ages are probably too young; the volcanics are gently folded with the rest of the Shaler Group which is overlain unconformably by Middle Cambrian and Ordovician sediments. Except within a few feet of gabbro sills the Shaler Group is unmetamorphosed. The main exposures of the Precambrian rocks are along the Minto Arch, where the structural discordance between Paleozoic and Precambrian rocks clearly demonstrates that Shaler and Natkusiak rocks were folded before the deposition of Cambrian sediments. The Precambrian rocks have been gently folded about easterly-northeasterly axes. Rocks dip at 10 degrees or less and the folds plunge very gently to the east or west.

On the basis of present information correlations of Shaler strata with late Precambrian rocks on the mainland can only be speculative. The volcanic rocks and gabbro sills, although they yield K–Ar dates that range from 390 to 640 m.y. are probably pre-Middle Cambrian. The Shaler rocks that are separated from the volcanic rocks by a disconformity could be Neohelikian or Hadrynian in age. The range of dates from sills cutting Coppermine sediments is 445 to 718 m.y., which overlaps with dates from Natkusiak volcanics and related sills, suggesting an age equivalence. Quartzites on the Richardson Islands and adjoining mainland of Victoria Island that have tentatively been included in the Glenelg are on strike with Coppermine sediments and cut by sills that intrude Coppermine sediments; therefore they may be part of the Coppermine succession. Similarly, quartzites at the head of Wellington Bay on the south shore of Victoria Island may be part of the Coppermine or even related to older quartzites on the mainland in the Bathurst Inlet area such as the Tinney Cove Formation. The main part of the Shaler succession has the characteristics of sediments deposited in a shallow-water, marine environment with gypsum and anhydrite extensively developed. That part of the Coppermine sediments above the unconformity also appears to be a typical, marine, shallow-water succession but gypsum beds are only local. Coppermine sediments, accordingly, may correlate with at least part of the Shaler Group.

Tectonic Summary

The oldest rocks, probably Archean, in Churchill Province are composed predominantly of basic to intermediate volcanic flows and pyroclastics, overlain by flysch-type sediments of greywacke and shale with thin bands of volcanic rocks, quartzite, and dolomite. The various belts are isolated and trend north to northeast to easterly. The rocks are folded about northerly to northeasterly axes and metamorphosed to varying degrees, transformed into migmatite or granitic gneiss, and cut by massive or slightly foliated granitic rocks. Only in a few places is there positive evidence for the age of these strata and no basement has been found beneath them. Most K–Ar age determinations yield Hudsonian dates but a number of older Kenoran ages are obtained. All these probable Archean rocks have been affected to varying degrees by phases of the Hudsonian Orogeny. Before the deposition of Aphebian sediments, all of what is now Churchill Province is thought to have been underlain by deformed Archean supracrustal and granitic rocks.

Aphebian rocks occur in a number of separate belts in Churchill Province. In places these rocks are known to rest unconformably on older, presumably Archean rocks and Kenoran granitic gneisses. Deposition of the sediments probably took place in separate geosynclines possibly of significantly different ages. All Aphebian rocks have been folded, metamorphosed very slightly to intensely, and some have been intruded by granitic rocks and converted to granitic gneiss. This orogenic activity is referred to as the Hudsonian Orogeny. It may have separate phases as it is probable that several separate orogenic belts occur in Churchill and Bear Provinces that roughly coincide with the several geosynclines in which Aphebian sediments were deposited (Fig. IV-14). The structural trends vary from belt to belt and intensity and style of deformation, degree of metamorphism, amount of granitization, and number of granitic intrusions vary along individual belts as well as between belts. In the areas between the orogenic zones, the rocks are for the most part granitic genisses that commonly yield K–Ar ages in the range 1,700 to 1,900 m.y. However, a number of older dates have been obtained from these gneisses. These gneisses and granites are thought for the most part to have formed during the Kenoran Orogeny and, therefore, to represent basement to the Aphebian sediments. They were affected by phases of the Hudsonian Orogeny as indicated by modifications of the K–Ar ages and then were cut by major fault zones some of which were probably initiated during late phases of the Hudsonian Orogeny.

Late or post-tectonic sediments and volcanics were deposited in isolated basins that were probably largely fault controlled. The sediments include conglomerate, feldspathic sandstone, lithic sandstone, arkose, and shale. The volcanic flows and pyroclastics and their intrusive equivalents range in composition from basalt to trachytes. The moderate to slight folding of these rocks took place possibly during movement along the faults. They are slightly metamorphosed but are not cut by granitic rocks. These rocks are late Aphebian if the related tectonic activity is considered to be a late phase of the Hudsonian Orogeny, or Paleohelikian if the tectonic activity is post-orogenic. The major period of intrusion of basic rocks is about 1,200 m.y. at which time the Mackenzie swarm of north-northwesterly to northerly trending dykes were intruded in the western part of the province. Other smaller swarms of about this age were intruded in the eastern part of the province. South of Great Slave Lake the dykes trend west of north and yield ages in the range 1,500 to 1,600 m.y.

Helikian and possibly Hadrynian sandstones and dolomites are flat lying or very gently folded. They were deposited on deeply eroded basement underlain by Kenoran and Hudsonian gneisses and granitic rocks, and Aphebian sediments in western Churchill Province.

Aphebian

Aphebian sediments in Churchill Structural Province were deposited on a basement of deeply eroded Archean rocks and granitic gneisses formed during the Kenoran Orogeny. Known exposures of Aphebian strata appear to be rather small segments of formerly more extensive geosynclines. There may have been three separate geosynclines, the relative ages of which are not known. They may have been of quite different ages and unrelated tectonically.

In the western part of the province in the East Arm Belt, Aphebian rocks were probably deposited in a geosyncline that included Aphebian rocks in Slave and Bear Provinces. This geosyncline had a northerly to north-northeast strike. Part of what is now Slave Province was a positive cratonic area during Aphebian time and a source of some of the sediments. The stratigraphic order of the various groups thought to be remnants of the original geosyncline is grossly similar, which suggests that they may be time equivalents. Basal quartzite strata are largely fluvial but partly marine and include dolomite in some sequences. They are succeeded by argillite and dolomite. Thick stromatolitic sequences occur in the dolomites in parts of the sections. The limy rocks are overlain and locally grade into siltstone, shale, and greywacke that represent a somewhat deeper water facies. Red, non-marine sandstone and shale are the youngest rocks of the various groups. Basic to intermediate volcanic rocks are present in the lower part of some sequences and at or near the top of others. In general the sequences are miogeosynclinal, composed of rocks deposited in a shallow-water, shelf environment. Non-marine and red bed strata indicate a number of marine regressions. There are differences between the

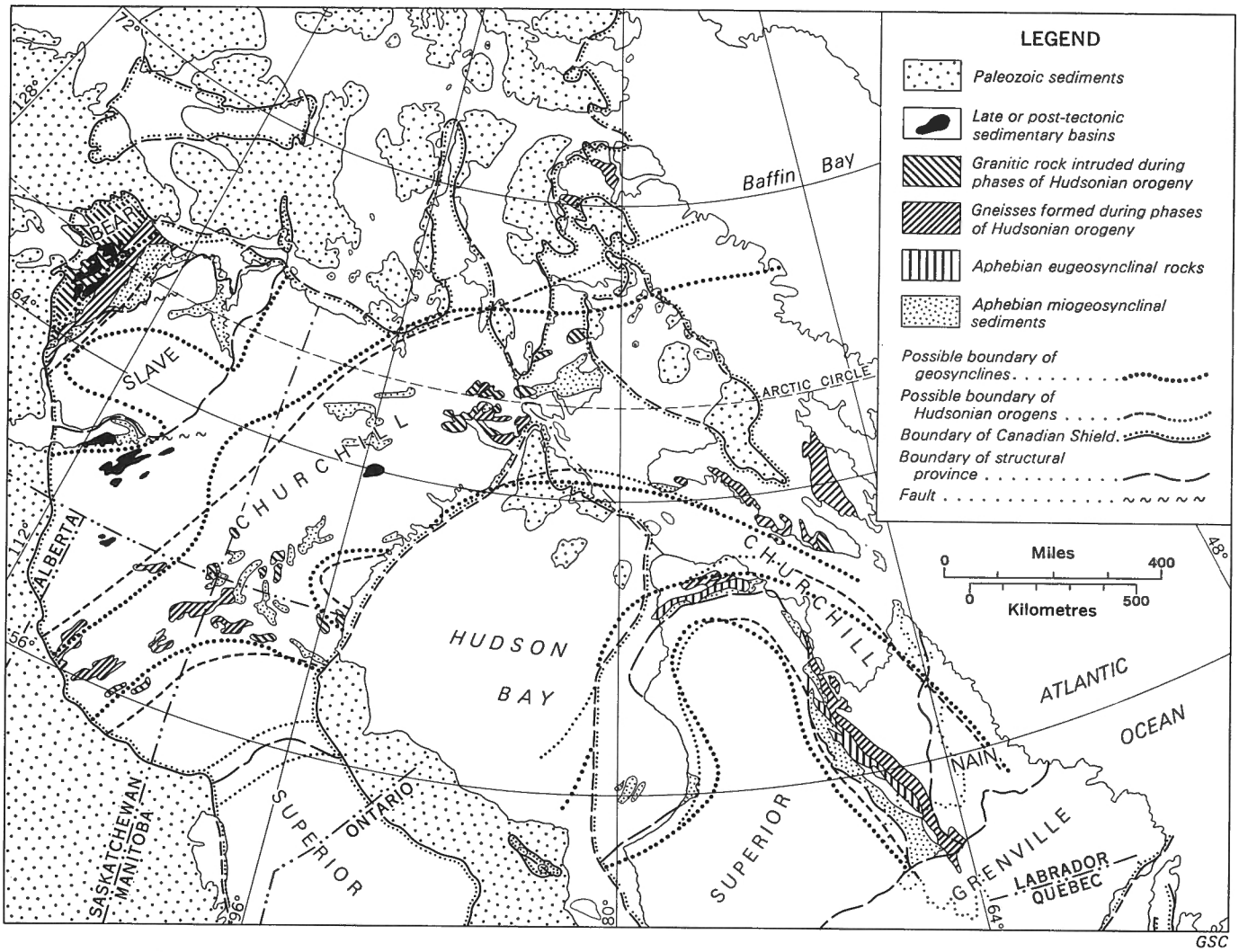

FIGURE IV-14. Possible extent of Aphebian geosynclines and Hudsonian orogens in Churchill and Bear Provinces of the Canadian Shield (by J. C. McGlynn).

various successions such as thickness of formations, variation in thickness of stromatolitic units, relative amounts of marine and non-marine sediments, and presence or absence of volcanic rocks.

In the central part of Churchill Province west of Hudson Bay, Aphebian rocks were probably deposited in a north-northeasterly to east-northeasterly trending geosyncline. The succession generally consists of a basal quartzite with lenses of conglomerate and greywacke overlain by argillite, carbonate, thinly bedded greywacke, and impure quartzite. Basic volcanic rocks occur locally near or at the top of the sequence. Where they are not in contact with younger granitic rocks the basal units rest unconformably on Archean rocks or granitic gneisses formed during the Kenoran Orogeny. The unmetamorphosed Aphebian sediments now exposed are generally miogeosynclinal rocks, generally shallow water or shelf facies, and largely marine types. Some of the clastic rocks are probably fluvial. The upper part of some sequences represents a transition to somewhat deeper water environments. Typical eugeosynclinal sequences, however, do not occur along the margins and if they were present in the interior parts of the geosyncline they have since been eroded or completely converted to gneisses. The limited evidence from paleocurrent and other studies indicates that the sediments were derived from the east, but the westernmost Aphebian rocks have not been studied in detail and their source is unknown.

East of Hudson Bay, Aphebian rocks occur that were probably deposited in a third geosyncline that includes rocks in the Labrador and Cape Smith Fold Belts and possibly also the Belcher Fold Belt in eastern Hudson Bay extending around the stable craton part of Superior Province. However it should not be considered to be a geosyncline marginal to a continental mass as Archean rocks are exposed north of the Cape Smith Fold Belt and east of the Labrador Fold Belt. Rocks of the Labrador Fold Belt consist of a western sequence of miogeosynclinal sediments that rest unconformably on Archean gneisses of Superior Province and an eastern sequence of eugeo-

synclinal sediments and volcanic rocks. Gabbro and ultrabasic sills probably related to the volcanism occur in both sequences. The eugeosynclinal sequence is partly a time equivalent of the miogeosynclinal formations, but is mostly younger than them. Basal unconformities are found in several places along the eastern margin of the belt but generally the Aphebian rocks are separated from mixed gneisses or granitic gneisses by a fault. The gneisses are probably metamorphosed equivalents of the sediments but locally may include Archean basement rocks. In Nain Province, Aphebian sediments occur which may mark the eastern boundary of the geosyncline. A western source area is indicated for the miogeosynclinal sediments and for at least some of the eugeosynclinal sediments. These rocks can be traced into Grenville Province where they become involved in the Grenvillian Orogeny. The geosyncline probably extended southwest to include the Aphebian rocks of the Mistassini Subprovince and some in Southern Province. In the Cape Smith Fold Belt the basal units along the southern margin of the belt that lie unconformably on Archean gneiss of Superior Province are miogeosynclinal sediments comprising quartzite, shale, dolomite, and iron-formation. They are overlain by thick basic volcanic rocks. In Belcher Fold Belt the sequence is essentially conformable and very thick, ranging up to about 30,000 feet. Most of the sediments are typical miogeosynclinal facies and record deposition in a shallow-water shelf environment; some are deposited in a fluvial environment.

The general distribution of Aphebian rocks suggests that they were deposited in three separate geosynclines that were not marginal to continent but were intracratonic. The sediments are for the most part miogeosynclinal facies but eugeosynclinal rocks are developed in some parts. An obvious difference between the various sequences is the occurrence of important thicknesses of iron-formation in Quebec and Labrador as opposed to the relatively insignificant development of these rocks in other areas. This reflects a difference in the nature of the source area or different conditions of deposition of the sediments. The presence of red beds and thick carbonate sequences that include significant stromatolitic zones in rocks that are probably more than 2 million years suggests that the chemistry of sea water and atmosphere has not changed in a fundamental way since Archean time.

In western Churchill Province several sedimentary basins have thick deposits of polymictic conglomerates, arkose, feldspathic or lithic sandstone, and shale. Basic to intermediate volcanic lavas and pyroclastics and hypabyssal intrusive equivalents occur in some. The sediments are largely non-marine or continental, often reddish and relatively immature. They are products of rapid erosion of a tectonically unstable terrain of considerable relief. The basins of deposition appear to be fault bounded with movement along the faults occurring during sedimentation. The rocks are moderately to gently folded and except near intrusions are unmetamorphosed. Volcanic rocks, where dated, yield maximum ages of about 1,730 m.y. The sediments and igneous rocks are typical of late tectonic, synorogenic activity and of post-orogenic tectonic activity. They may therefore be late Aphebian or Paleohelikian and may possibly correlate with similar rocks that occur in Bear Province.

Hudsonian Orogeny

The Hudsonian Orogeny in the late Aphebian affected the greater part of Churchill Structural Province as well as Bear and Southern Provinces. Within Churchill and Bear Provinces, it is possible to very tentatively outline three orogenic zones (Fig. IV-14) in which Aphebian rocks are folded, metamorphosed, converted to granitic gneisses, and intruded by granitic rocks. These orogens coincide approximately with the Aphebian geosynclines outlined in the preceding section. Although micas from rocks formed during these orogenies yield K–Ar dates that range between 1,650 and 1,850 m.y., the individual orogenies may not be precisely the same age. Accordingly, they are referred to as phases of the Hudsonian Orogeny.

The most westerly of the possible orogenic zones occurs essentially in Bear Province and is described in that section. Aphebian rocks in East Arm Subprovince are very slightly metamorphosed, but not cut by granitic rocks, and are moderately to gently folded about axes that trend northeast. The deformation may be of about the same age as the orogenic activity to the west and related to it, but these rocks, which are thought to be direct correlatives of Aphebian rocks in Bear Province, are thought to be east of the zone of intense orogeny.

In central Churchill Province, Aphebian strata are folded about north-northeasterly to east-northeasterly axes. In some areas these structures are refolded about north to northwesterly trending axes. Intensity of deformation, metamorphism, and granitization increases from the eastern margin to the interior of the orogen and along the axes of the orogen is most intense in the southwest and northeast sections. In the central part of the orogen, the exposed granitic gneisses may be largely Archean gneiss little affected by the orogeny. To the northeast in the Melville Peninsula and possibly on Baffin Island folding and metamorphism is again more intense. Granitic rocks are intruded along the length of the orogen and high-level intrusions of fluorite-bearing granodiorite are characteristic.

In Quebec and Labrador a third orogenic zone partly corresponds to the Aphebian geosyncline. Folding and metamorphism increase in intensity to the east and north. Fold axes in the Labrador Fold Belt trend northwest to north and are inclined to the east or northeast. Metamorphism increases to the east towards a belt of granitic gneiss and migmatite that include highly metamorphosed remnants of the Aphebian sediments and volcanic rocks. Similar gneisses occur along the northern margin of the Cape Smith Fold Belt. Zones of remobilized Archean basement rocks may be within these gneisses. The

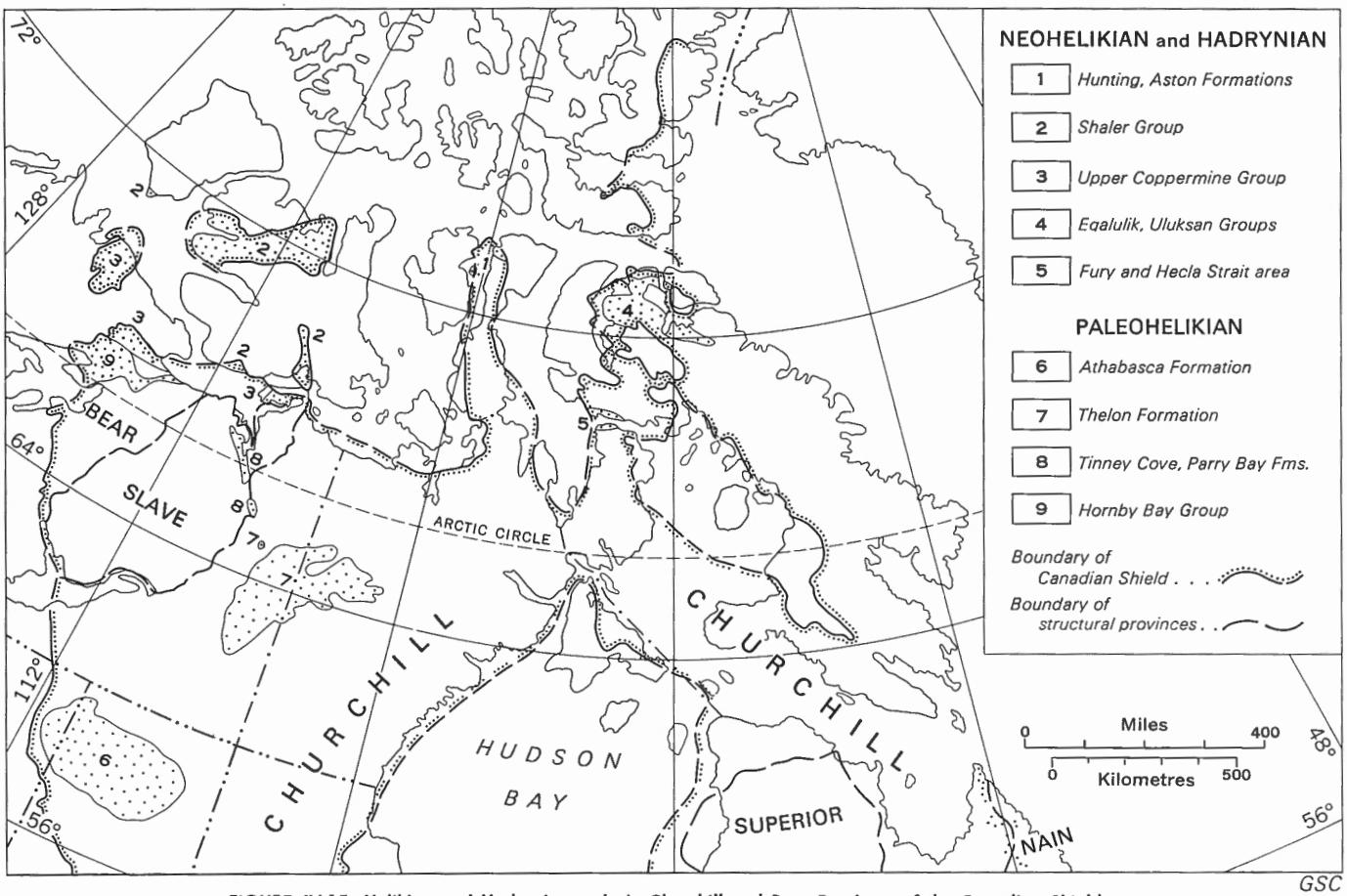

FIGURE IV-15. Helikian and Hadrynian rocks in Churchill and Bear Provinces of the Canadian Shield.

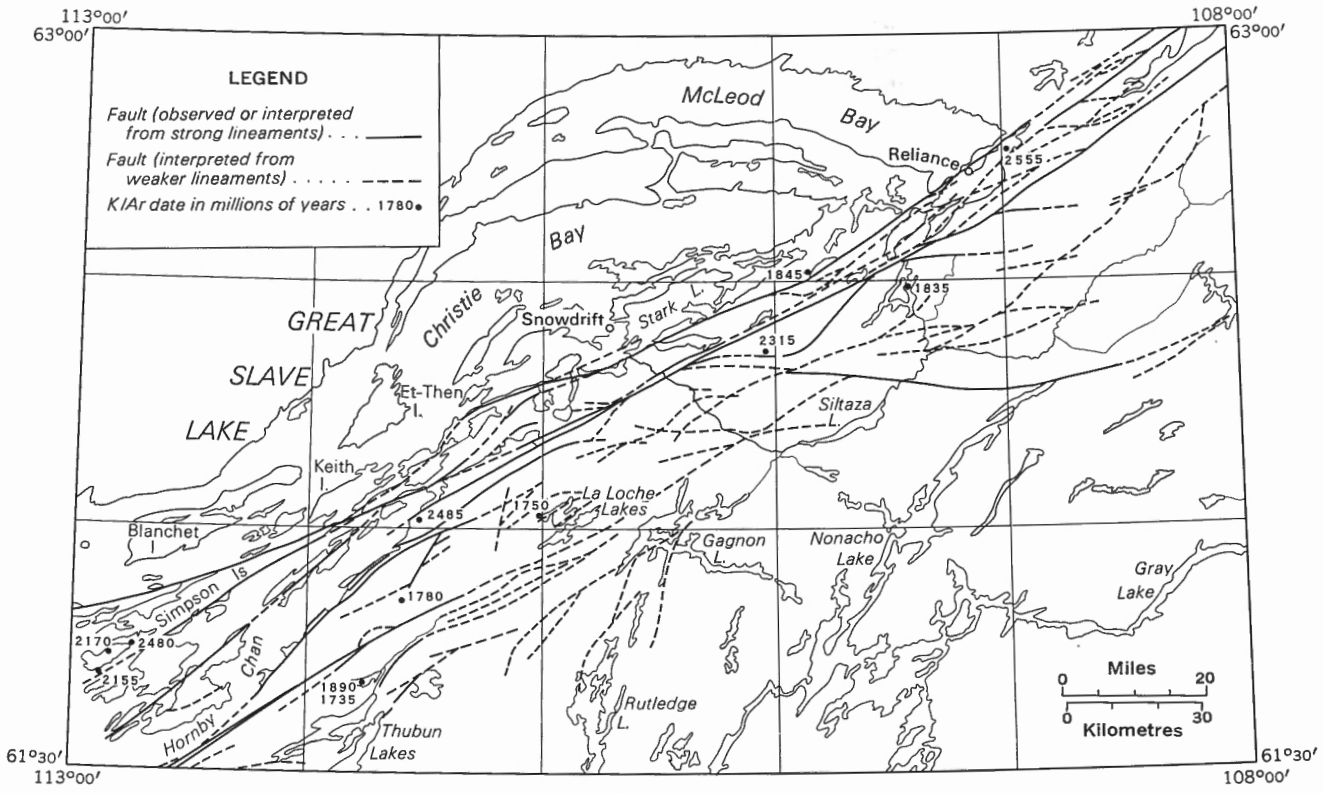

FIGURE IV-16. The McDonald fault system south of Great Slave Lake, Northwest Territories (by E. W. Reinhardt).

Aphebian strata of Belcher Fold Belt are moderately folded but little metamorphosed.

In the areas intervening between the orogenic zones in the Churchill Province rocks are granitic gneisses with volcanic and sedimentary rocks of possible Archean age. The granitic gneisses probably are also for the most part Archean rocks that have been little affected by the Hudsonian Orogeny. They have been cut by major fault zones that strike north, northeast, and northwest. Some are complex zones along which wide bands of sheared, brecciated, or mylonitized rocks are developed. They probably were initiated in the Hudsonian Orogeny but movement continued during the Helikian. Some zones occur near the boundaries of the structural province and others near the margins of the orogens. Syn- or post-tectonic sediments were deposited in basins developed along some of these fault zones.

Helikian and Hadrynian

Helikian rocks outcrop in several areas of Churchill Province (Fig. IV-15). The character of the strata in various basins is similar, being generally composed of quartzite, conglomerate, siltstone, shale, and stromatolitic dolomite. The basal quartzites were deposited on a deeply eroded basement during a period of tectonic stability in a fluvial environment as indicated by the sedimentary structures. The other strata were probably deposited in a near-shore marine environment. Remnants between the main exposures suggest that these sediments were more extensive, covering much of the western part of the shield particularly in Paleohelikian time, but parts of Slave and Bear Provinces may have been positive areas. Neohelikian and Hadrynian sedimentary and volcanic rocks are rare in Churchill Province. On northwestern Baffin Island and Somerset Island the sequences are grossly similar, comprising thick basal quartzites that grade upwards to dolomite, and then in the thickest sequences into silty shale and quartzites. Some basic volcanic rocks are near the base. They may be correlative with the Coppermine sediments of Bear Province, and the Shaler Group on Victoria Island which are shallow-water marine deposits that may have been deposited in the same basin.

Faults and Diabase Dykes in Churchill Province

The most prominent swarm of diabase dykes that cut the rocks of Churchill Province are north-northwesterly trending, vertical-dipping dykes of the Mackenzie swarm whose age is about 1,200 m.y. These rocks are about the same age and probably are related genetically to the Coppermine lavas, the sills in the Goulburn and Great Slave Groups, and possibly also volcanics in the upper part of the Dubawnt Group. A swarm of northwesterly striking dykes in Baffin Island yield dates in the range 1,000 to 1,200 m.y., and the volcanic rocks near the base of the Eqalulik Group may be related to these dykes. Other dykes and sills of gabbro and basic volcanic rocks that occur throughout the Proterozoic rocks in the Arctic Islands and northern mainland give K–Ar ages that range from 400 to 750 m.y. There seems to be two periods of major basic magma intrusion, at about 1,100 to 1,200 m.y. and 600 to 750 m.y.

Faults cut all rocks of Churchill Province but may be more abundant in the older rocks. They trend north, northeast, and northwest and are commonly steep-dipping. Relative ages of these structures or absolute ages of major movement are uncertain. Some of the major structures, such as the McDonald *(Frontispiece;* Fig. IV-16) and Bathurst fault zones, those near Lake Athabasca, and along the Slave–Churchill boundary, were initiated early, about 1,700 to 1,760 m.y. ago before and during deposition of late Aphebian or early Paleohelikian red bed sequences. Wide zones of intense crushing, brecciation, and mylonitization were formed during this early period of movement. Later movement took place and many new faults were formed. These structures are characterized by narrow breccia or shear zones and probably occurred at a higher crustal level. They are numerous in the gneissic terrain but also cut Paleohelikian and Neohelikian rocks. The faults may be temporally related to that period of tectonic instability during which the older diabase dykes and Neohelikian lavas were generated. After deposition of Neohelikian and possibly Hadrynian sediments some new faults formed and movement occurred along older structures. These structures may be related to the Hadrynian gabbroic intrusions and volcanic rocks. Along a few faults such as the McDonald, there is evidence of movement in Paleozoic time (Chapter VIII). Most faults appear to be strike-slip types with left-hand displacement being more common than right-hand. However, some of the older faults with a long history of movement appear to have a considerable vertical component.

Thompson Belt

The Thompson Belt includes a rather narrow part of Churchill Province along the boundary with Superior Province in northern Manitoba (Bell, 1966; Davies, *et al.,* 1962). The oldest rocks are greywacke, argillite, feldspathic quartzite, minor conglomerate, skarn, metavolcanic rocks, and amphibolite. These rocks occur as structurally conformable sequences in belts that trend northeast to east. The strata appear to be folded isoclinally and may be overturned. The sediments locally are called the Assean Lake Group. With increasing metamorphism they grade into well-banded, biotite quartz feldspar gneiss that may also contain hornblende, garnet, magnetite, or cordierite. The gneisses grade into or are intimately mixed with less well-banded granitic gneiss that grades into or is cut by massive to faintly gneissic grey to pink granodiorite, massive red porphyritic granite, and pegmatite. Ultrabasic and basic rocks intrude the sediments and volcanic rocks and their metamorphosed equivalents. They

occur as lenticular, concordant masses several hundred to a few thousand feet long and are restricted to a narrow zone of shearing and faulting, that strikes northeasterly to parallel with the Churchill–Superior boundary. The ultrabasic rocks are serpentinized, but where not sheared, textures indicate that they were originally peridotite composed of olivine and pyroxene. In places these rocks appear to be metamorphosed and pyroxene, amphibole, olivine, clinochlore, and spinel are developed. Some ultrabasic bodies are cut by pegmatite.

The zone into which the ultrabasic rocks are intruded coincides with a belt of low gravity which to the northeast is coincident with the boundary with Churchill–Superior Provinces. In Superior Province a belt of high gravity is coincident with a belt of high-grade metamorphic rocks, largely in the granulite facies. The various units have also their own magnetic expressions (Kornik and MacLaren, 1966).

The age of the volcanic and sedimentary rocks and the ultrabasic intrusions is a matter of uncertainty—they may be Aphebian or Archean. Bell (1966) suggests that they represent the redeformed western extension of the Fox Lake Belt, the age of which is also uncertain. The small ultrabasic masses, he suggests, were derived from differentiated basic to ultrabasic sills and were reintruded during the second deformation. The ultrabasic masses are considered to be Alpine-type intrusions by Davies, *et al.* (1962). The nature of the Hudsonian Orogeny in this area is not completely known; some of the presently observable effects may result from the older Kenoran Orogeny. The sediments and volcanics have been folded and probably refolded; they were granitized and intruded by some granitic rocks and pegmatites and certainly were metamorphosed, intensely sheared, and faulted. Northeasterly to easterly trending fault zones and faults are concentrated in the boundary area of the two provinces. The precise location of the boundary is a matter of some controversy. Stockwell (1961) places it along the zone where the easterly trending structures of Superior Province are abruptly cut off by northeasterly striking structures of Churchill Province (Fig. IV-11). Bell (1966) suggests that the boundary be placed along the north contact of the zone of high-grade gneisses of the granulite facies. Kornik and MacLaren (1966) place the junction along a narrow continuous magnetic low (Pl. IV-4). The latter boundary almost coincides with that of Stockwell and may be considered a refinement of it.

Flin Flon–Lynn Lake Region

The oldest rocks of the Flin Flon–Lynn Lake region are basic to acidic volcanic flows, related tuffs and agglomerates, arkose, greywacke, and shale. These rocks are in two belts that trend east through Lynn Lake and Flin Flon. The intervening area is underlain by a complex of granitic gneisses and intrusions. In the Flin Flon Belt the oldest rocks are generally considered the Amisk Group, which contains thick massive or pillowed basalt and andesite that are overlain by intercalated basic flows, dacite and rhyolite, tuffs, breccias, and agglomerates. In some areas these rocks are conformably overlain by greywacke and shale, or grade along strike into them. Gabbroic sills or plutons occur in the volcanic rocks and locally also in the sediments and are probably intrusive equivalents of the lavas. Similarly, intrusive masses of quartz feldspar porphyry and rhyolite porphyry are considered equivalents of the acidic lavas. Minimum thickness of the group varies from a few thousand to at least 20,000 feet.

In the vicinity of Flin Flon, Amisk rocks are overlain by the Missi Group. At Flin Flon, Stockwell (1960) considered the base of the Missi to be an unconformity. West of Flin Flon, Byers and Dahlstrom (1954) also found an unconformity at one locality but considered the two groups are regionally structurally conformable and that no major time break exists in the sequence. The basal unit of the Missi Group is normally composed of a conglomerate that consists of angular to subrounded boulders or pebbles of the various lithologies in the underlying Amisk along with minor amounts of chert, jasper, and quartz and very rare cobbles of granite in a greywacke or arkosic matrix. The unit grades upwards to greywacke and subgreywacke with thin beds of conglomerate. The minimum thickness of the Missi appears to be at least 5,000 feet. Sediments that have been given various local names in the area between Flin Flon east to Wekusko Lake have been tentatively correlated with the Missi by Harrison (1951) and others. They are essentially greywacke and shale with local limy beds and, in places, considerable thicknesses of volcanic rocks. Conglomerates are commonly not present. The sediments are structurally conformable with the underlying Amisk and in places interbedded with Amisk flows. Locally sediments and Amisk volcanics are separated by faults. All sediments were probably deposited near the end of the Amisk volcanism, or just after it, and volcanic activity continued in places during the deposition of sediments. The Amisk and Missi rocks may have occurred originally in easterly trending belts.

The Kisseynew Complex occurs in easterly trending belts north of the Amisk and Missi Groups, extending from east-central Saskatchewan easterly to the Paleozoic cover of Hudson Platform. East of Sherridon, Robertson (1953) has subdivided the Kisseynew into the Nokomis and Sherridon Groups, based largely on structural interpretation. The Nokomis Group is a hornblende gneiss consisting essentially of hornblende and plagioclase with variable amounts of quartz, garnet, biotite and epidote and quartz–plagioclase–biotite gneiss with variable amounts of garnet, hornblende, microcline, cordierite, and sillimanite. The hornblende gneiss occurs as thin bands interbedded with the more abundant biotite gneiss and

grading into it. The Nokomis is overlain conformably by the Sherridon Group which consists of quartz-rich gneiss, hornblende gneiss, limestone, and a band of anthophyllite-bearing rock. The transition between the two groups is marked by graphite in the upper fine-grained parts of the Nokomis succession, and by discontinuous orthoquartzite and limestone near the base of the Sherridon sequence, which grade upwards into quartz-rich gneiss. Quartz forms more than half these latter rocks along with some plagioclase, garnet, and usually biotite. The hornblende gneiss is composed of plagioclase and hornblende with lesser amounts of quartz, biotite, garnet, and epidote. Relict pillows occur in some bands but other hornblende gneisses grade into the quartz-rich gneiss which suggests a sedimentary origin. Chemical analyses indicate that the Nokomis sediments are close to the average composition of greywacke and that the Sherridon sediments are approximately the average composition of subgreywacke. This stratigraphic division of the Kisseynew Complex has been extended only short distances beyond the original map-areas. Elsewhere Kisseynew rocks are commonly divided into hornblende gneisses and biotite gneisses but the stratigraphic sequence is usually unknown. Robertson (1953) suggests that Nokomis represents eugeosynclinal deposits of greywacke and shale and that the Sherridon sediments were more mature and possibly deposited in a shallow-water shelf environment.

The relationship between rocks of the Kisseynew Complex and the Amisk and Missi Groups is not entirely understood. The unresolved problem has been ably treated by Harrison (1951) and more recently by D. S. Robertson (1951, 1953), Kalliokoski (1953), Byers and Dahlstrom (1954), and McGlynn (1959). Four hypotheses have been proposed: (1) the Kisseynew gneisses include Amisk and Missi rocks and possibly also older and younger strata; (2) the gneisses are younger than the Amisk, but conformable with it, and older than the Missi; (3) the gneisses are the metamorphosed equivalents of Missi strata only; (4) the gneisses are separated from Amisk and Missi strata by a major fault and the relative ages are, therefore, unknown. It is certain that in some areas Kisseynew rocks are separated from Amisk and possibly also Missi strata by faults. In other parts of the contact zone, rocks that can be traced into Amisk volcanics are structurally conformable with Kisseynew gneisses and in still other segments, Amisk and Missi strata have been traced through zones of increasing metamorphism to rocks indistinguishable from Kisseynew gneisses. Conglomerates have been found in the Kisseynew Complex that are lithologically similar to Missi conglomerates (Byers and Dahlstrom, 1954). Kisseynew rocks are generally more highly metamorphosed and granitized than the Amisk and Missi rocks but the latter attain an equivalent grade of metamorphism near areas of Kisseynew gneisses. Much of the Kisseynew metasediments and hornblende gneisses could be derived from greywackes and volcanic rocks similar to the Amisk and Missi strata. The Sherridon Group of more siliceous and limy sediments in the Sherridon area has no known lithologic equivalents. It seems likely that the Kisseynew Complex is in large part the metamorphosed equivalent of the Amisk and Missi Groups and their equivalents, and that the Sherridon Group may represent a facies variation or a sequence younger than the Missi sediments.

In the Lynn Lake Belt the oldest rocks belong to the Wasekwan Group and consist of a conformable sequence of basic, intermediate, and acidic volcanic rocks, related tuffs, agglomerates and breccias, and greywacke, and shale. These rocks are very similar to Amisk strata. In the Lynn Lake area the volcanic rocks form the oldest exposed units, but to the southeast around Granville Lake, the oldest exposed strata are sediments that have intercalated thin volcanic horizons; east of Lynn Lake the sediments and volcanic rocks are also interbanded. These rocks occur in easterly trending belts. Gabbro sills and plutons and acidic porphyry intrusions that are probably intrusive equivalents of lavas occur in the Wasekwan sequence. Rocks of the Sickle Group overlie Wasekwan strata. Regionally the sequences appear conformable although locally some evidence of unconformity has been found. The basal unit of the Sickle is usually conglomerate, consisting of subrounded to subangular pebbles or cobbles of sedimentary or volcanic rocks of the Wasekwan Group with very minor amounts of quartz and granitic rock in a greywacke matrix. The conglomerate varies in thickness from a few feet to more than 1,000 feet and grades upward to arkose and greywacke through a transition zone of fine clastics with lenses of conglomerate. Arkose and subgreywacke with minor amounts of shale constitute the upper unit of the Sickle Group. Cross-bedding, and less commonly graded bedding, are present throughout the sequence. Calc-silicate lenses and irregularly shaped zones occur in the arkose. Around Granville Lake several bands of massive and pillowed basic volcanic rocks, indistinguishable from the older Wasekwan volcanics, occur within the Sickle sediments. Minimum thickness of Sickle rocks is 10,000 feet.

Both the Amisk–Missi and Wasekwan–Sickle sequences appear to be typical eugeosynclinal deposits. Volcanic rocks are most abundant in the lower part and recur within the flysch deposits of the upper part of the sequences. Local unconformities appear to exist in the Flin Flon area. Volcanic rocks provided most of the clasts in the conglomerates but the proportion of potash feldspar and quartz in the finer clastics, particularly the Sickle sediments, suggest a source area partly underlain by granitic rocks. Sickle strata appear to have been deposited in a shallower-water environment than the Missi greywackes. Age relationships between these widely separated rocks are not known but they are usually considered correlative. The Amisk and Missi Groups extend to the west and then northerly. It is possible that they could be traced through the belt of high-grade meta-

morphism and granitization into recognizable Wasekwan and Sickle rocks.

In the Lynn Lake region the older rocks and their metamorphosed equivalents have been folded about eastward to southeastward and locally northward striking axes. These folds have moderate plunges and axial planes dip steeply, mainly south. These structures have been refolded about northwestward and, locally, northeastward trending axes. Amisk and Missi rocks are folded about northward to northeastward striking axes that commonly are westward dipping and the folds generally plunge to the north. As the structures are traced north towards the Kisseynew Complex the foliation and relict bedding strike easterly and the folds are refolded and wrapped around granitic intrusions. Kisseynew rocks are complexly folded and various interpretations of the structures have been made. A generally applicable interpretation involves folding about westerly trending axes, overturned to the south, and refolding about northerly trending axes.

All rocks of the Wasekwan and Sickle Groups are regionally metamorphosed to the amphibolite facies with local zones of epidote–amphibolite or even high greenschist facies. Basic volcanic rocks are, therefore, converted to hornblende–plagioclase schist or finely banded gneisses with variable but minor amounts of epidote and biotite; the sediments are converted to biotite–muscovite–feldspar–quartz schists and gneisses that locally contain garnet and sillimanite. The rocks grade southward into coarse-grained, biotite–quartz–feldspar gneiss, hornblende–plagioclase gneiss, and granitoid gneiss containing pegmatites and extensive zones of migmatite. With increasing amounts of granitic material the rocks grade into granitic gneisses. This gneiss complex is lithologically similar to Kisseynew Complex of the Flin Flon area. The complex is intruded by and grades into batholiths of gneissic to massive grey biotite–hornblende or biotite tonalite and granodiorite, and pink granodiorite.

In the Flin Flon region Amisk and Missi rocks and their assumed equivalents are metamorphosed to the epidote–amphibolite or amphibolite facies with smaller areas of rocks in the greenschist facies. As these strata are traced to the north they grade into rocks in the high amphibolite facies similar to Kisseynew gneisses and which, in some areas, are grouped with the Kisseynew Complex. The Kisseynew Complex is consistently in the high amphibolite facies and locally may reach hornblende–granulite facies. Included in the Kisseynew Complex are granitoid rocks, migmatite, and gneissic granodiorite. The Amisk and Missi Groups and their equivalents are intruded by granitic rocks, commonly granodiorite, and locally grade into these rocks through zones of migmatites or granitized gneiss. Most granites that intrude Amisk–Missi units also are younger than Kisseynew rocks. The large batholiths are biotite or hornblende–biotite granodiorite or quartz diorite. Smaller masses that cut these rocks and appear to be higher level intrusions consist of pink biotite granodiorite and granite, or diorites and syenodiorite. These rocks are late, in part post-tectonic, and limited in distribution (McGlynn, 1959). "Quartz-eye" granites appear to be the oldest granites in the region and are quartz diorite in composition. They are found only in Amisk volcanic rocks and are distinguished by the common occurrence of blue quartz eyes. Some appear to be metamorphosed. The "quartz-eye" granites are probably of diverse origins and probably include intrusive quartz diorite batholiths, metamorphosed or granitized acidic pyroclastics, and intrusive equivalents of acidic volcanic rocks.

With one exception minerals from all granitic rocks and metamorphosed older rocks yield Hudsonian dates in the 1,700 to 1,800 m.y. range. Muscovite from metamorphosed Sickle sediments gives an age of 2,670 m.y.; a coexisting biotite from the same sample yielded an age of 1,650 m.y. On the basis of the older date and because of lithological similarities to rocks in Superior Province, Wasekwan–Sickle and Amisk–Missi strata are tentatively classified as Archean. The granitic rocks may be Hudsonian but also may be Kenoran granites that have been updated by effects of the Hudsonian Orogeny.

Southwestern Churchill Province

This region includes that part of Churchill Province between Great Slave Lake and Lake Athabasca and south and southwest of the latter lake. This area is underlain chiefly by granitic rock types but both older and younger sedimentary and volcanic rocks occur.

Archean. The oldest rocks are the Tazin Group north of Lake Athabasca. At Uranium City the Tazin (Tremblay, 1957) consists of greywacke, shale, quartzite, limy quartzite, dolomite, iron-formation, and amphibolites of uncertain origin. The sediments are metamorphosed to the amphibolite facies or granitized to banded gneiss, quartzitic granitic gneiss, and gneissic granite. Greywacke and shale were probably the dominant rock types in the original sequence which may have been at least 12,000 feet thick. Some of the amphibolites may have been basic flows or sills but the numerous narrow bands interbedded with the metasediments were probably tuff or sediments. In northwest Saskatchewan and southern District of Mackenzie, rocks tentatively correlated with the Tazin Group are thinly bedded greywacke and shale with some chert. They have metamorphosed to granitic gneiss, or graded into it, and only small remnants remain. West of Nonacho Lake and south of Great Slave Lake metamorphosed basic volcanic rocks, subgreywacke, and shale grade into migmatite and granite. Correlation with the Tazin is uncertain. Similar rocks and their granitized equivalents occur northeast of Uranium City in the Snowbird Lake area (Taylor, 1963). All these deposits are similar lithologically and appear to have been deformed and altered in a

similar style so that they can be considered as possible equivalents. Tazin rocks and possible equivalents to the northeast are highly folded about northeast striking axes that in places are curved suggesting later cross-folding. Northwest of Uranium City folds trend north to northeast, whereas at Uranium City easterly trends predominate. The rocks of granitic composition form belts of migmatite, banded granitoid gneiss, and porphyroblastic granitic gneisses. These rocks, although variable, consist essentially of plagioclase, microcline, quartz, and biotite, with or without hornblende, and with variable amounts of muscovite, chlorite, and garnet. Pegmatites are common. The gneisses are intruded by batholiths of massive granodiorite, partly porphyritic. East of the Nonacho Belt are pyroxene-bearing gneisses of the granulite facies. Foliation is parallel with bedding in the metasediments and trends northeasterly to northerly. K–Ar dates range from 1,700 to 1,850 m.y. but a few are in the range 2,000 to 2,300 m.y.

Hudsonian Orogeny. The main expression of the Hudsonian Orogeny in the region may be faulting. Extensive zones of mylonites and brecciation occur at Uranium City and to the northeast and northwest. In Alberta and south of Great Slave Lake, numerous northerly trending fault zones are marked by brecciation and mylonization. A single zone of complex faulting may extend from northeast of Uranium City to northwest Saskatchewan and then north to the Nonacho Belt. A similar wide zone occurs along the McDonald fault south of Great Slave Lake *(Frontispiece)* and another at Snowbird Lake. The zones appear to have formed before the late Aphebian, Nonacho, and Martin sediments were deposited. Later movement took place along many of the faults displacing the sediments and the mylonite zones. The late faults are not restricted to older zones of movement but have the same trends, that is, northeast, northerly, and northwest. They are steeply dipping and marked by narrow zones of brecciation and fracturing. Relative ages of the three sets of faults have not been clearly established. The early phases of faulting with the development of mylonites and breccia zones may have occurred during the Hudsonian Orogeny possibly as a late phase. K–Ar ages on muscovite along the McDonald fault, and from the sheared basement of the Nonacho sediments are in the 1,700 to 1,800 m.y. range. Movement on the later faults is probably post-orogenic because Helikian rocks are cut by some of these faults.

Aphebian or Paleohelikian. At Uranium City, the Martin Formation was deposited on an erosion surface of considerable relief formed on the Tazin Group and Hudsonian intrusions. The northeast-trending belts of Martin rocks are probably fault-preserved remnants of more extensive basins. The formation (Tremblay, 1957) consists of a minimum of 13,000 feet of conglomerate, arkose, siltstone, and basaltic or andesitic flows and sills. The basal conglomerates comprise angular to subangular boulders of basement rock in an arkosic matrix (Pl. IV-10). The arkose is orangish red and consists of feldspar, quartz, and rock fragments. It is crossbedded and ripple-marked. The sand grains are angular to subrounded and poorly sorted. Numerous conglomerate lenses and thin beds are intercalated. The boulders in these units are smaller and better rounded than in the basal conglomerate, but also consist largely of recognizable basement lithological types. The siltstone beds are deep red, feldspar-rich, and mud-cracked. The volcanic rocks are basaltic in composition, usually amygdaloidal, and commonly porphyritic. K–Ar whole rock samples are as much as 1,700 m.y. old. These

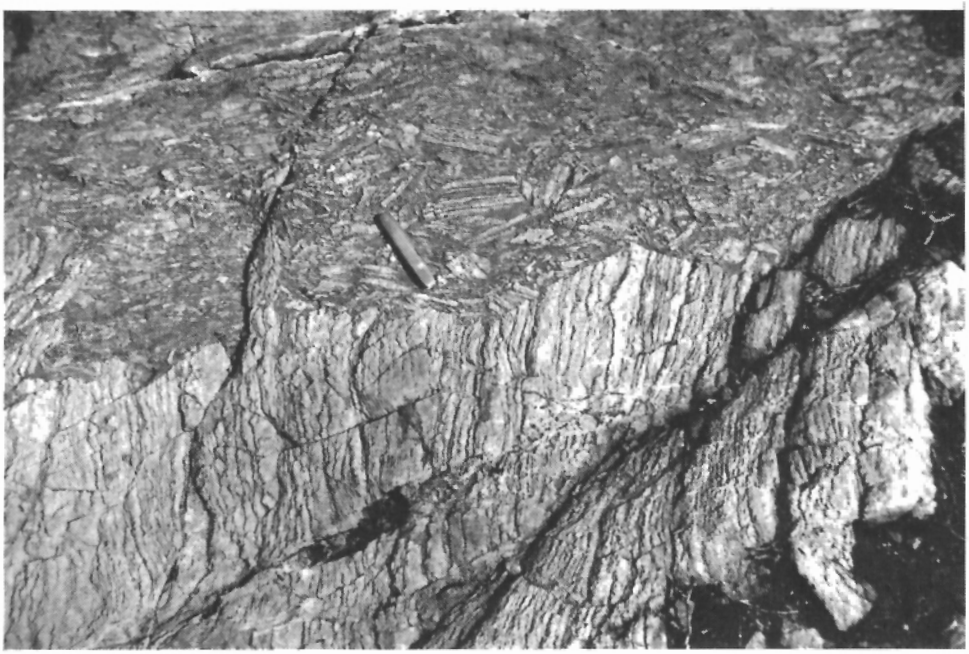

PLATE IV-10
Unconformity between late Aphebian Martin Formation and gneisses of the Archean Tazin Group, Churchill Province, Uranium City, Saskatchewan.

continental sediments are typical red bed deposits, the products of rapid erosion and deposition under tectonically unstable conditions deposited in a fluvial environment. The basins of deposition may have been partly fault controlled. Deposition took place after or possibly contemporaneous with fault movements that produced the early mylonite and breccia zones in the basement rocks. Paleocurrent studies by L. P. Tremblay indicate a transport direction that varies from southeast to southwest and suggests a source to the north. The Martin has been gently folded about northeasterly trending axes. The rocks are not cut by granites but are displaced by late faults with northeasterly strikes. They are thought to be late Aphebian sediments that were deposited during a late phase of the Hudsonian Orogeny.

Rocks of the Nonacho Group are exposed south of Great Slave Lake. They are lithologically similar to the Et-Then and Martin Groups and were deposited in similar environments. The Nonacho Group comprises a conformable sequence of polymictic conglomerate, and conglomeratic sandstone that include lithic sandstone, arkose, and feldspathic quartzites, siltstone, and shale. Along the southern contact with the basement the basal units are sedimentary breccia composed of angular boulders of granitic rocks from the nearby gneisses in an arkosic matrix. The breccias grade upwards into arkose with scattered boulders of granite and beds and lenses of disintegrated granite. Below the unconformity the basement rocks are broken into large blocks sealed by seams of muddy arkose. The lower arkose is overlain by coarsely bedded polymictic conglomerate containing closely packed subrounded boulders, chiefly of granitic rocks and gneisses, but with some quartz, quartzite, and basic to acidic volcanic rocks. The conglomerate is overlain by buff and grey lithic sandstone, arkose, and feldspathic quartzite in which laminations and crossbedding are common. These rocks consist of variable amounts of quartz, feldspar, and lithic grains that are subrounded to subangular and cemented by carbonate. Thin red or green shale beds occur and lenses or thin beds of conglomerate or clusters of pebbles and shale chip conglomerate are abundant. Within the sandy unit is a fairly thick band of conglomerate distinguished by large boulders of white quartzite as well as granite. Siltstone and shale units may be fairly thick, particularly in the north. In shaly units mudcracks and sandstone dykes are found. Crossbed orientation (McGlynn, 1966) in the southern part of the belt indicates that paleocurrents were from north to south along the eastern margin and south to north along the western margin of the belt. They probably do not correspond to sedimentary transport directions. Source areas, in addition to nearby basement highlands, may have been to the north as quartzite boulders in the conglomerates are lithologically similar to the quartzites of the Archean Wilson Island Group exposed on islands in Great Slave Lake.

Nonacho rocks rest on a basement of granitic gneiss, migmatites, and massive granitic rocks. In the south they are not cut by younger granites and are unmetamorphosed. The strata are moderately folded about axes that strike north-northeast to northeast and plunge to the north. The folding is most intense in the western part of the belt. Basement rocks are cut by northerly trending faults along which extensive breccia and mylonite zones are developed. Nonacho rocks, especially near contacts with the basement, are cut by northerly and northeast-striking faults, movement on which is post-folding.

The Athabasca Formation, locally overlain conformably by the Carswell Formation, outcrops over a large area south of Lake Athabasca. The Athabasca is separated from the Tazin and granitic gneiss by an unconformity and is not in contact with the Martin Formation. The Athabasca Formation consists of orthoquartzite with rare interbeds of shale and quartz-pebble conglomerates (Fahrig, 1961). The sandstone is composed almost entirely of rounded, moderately well sorted quartz grains in a quartz cement. Both trough and planar crossbedding are abundant throughout the sequence and indicate sedimentary transport from the east and southeast, or locally, from the northeast. Grain-size distributions and ripplemark orientations corroborate this conclusion. The Carswell Formation consists of a pink or buff, finely laminated dolomite, grey dolomite with many stromatolitic zones, and dolomite breccias. It is strongly deformed and forms the margin of a circular structure, within which are outcrops of the orthoquartzite and granitic basement. The Athabasca sandstone is considered to be a product of slow erosion of a tectonically stable source area to the east and southeast of present outcrops. Sediments were deposited in fluvial environment. Part of the sequence may be marine and probably the Carswell Formation was deposited in a shallow-water, marine environment. They are post-Hudsonian and seen to be separated from the orogenic event by a period of erosion. They are then probably younger than the Martin strata. Except for the deformation in the circular structure the Athabasca rocks are unfolded but may have been gently tilted. They have been cut by a northeast-striking fault and intruded by a few diabase dykes, one dated at 1,230 m.y. The Athabasca Formation is considered Paleohelikian and tentatively correlated with the Thelon Formation of the Dubawnt Group.

East Arm Subprovince

In the East Arm Fold Belt are Archean rocks of the Wilson Island Group and the Kenoran granitic rocks that have been described under the section on Slave Province. Aphebian rocks are from the Union Island, Great Slave, and Et-Then Groups. The Union Island Group outcrops in only a small area on Union and adjacent islands. It consists of black and grey shale, pillowed basalt, dolomite, and greywacke. The group is older than

the Great Slave Group and separated from it by an unconformity (Hoffman, 1967).

The Great Slave Lake Group has been divided by Stockwell (1936) into six formations, the basal units of which rest on a deeply eroded surface. It is about 20,000 feet thick and the strata occur in an easterly trending synclinorium. Along the north flank the strata dip very gently south but are folded in the central and southern parts. The rocks are cut by sills, laccoliths, and possibly also stocks of hornblende–biotite diorite or quartz diorite. A biotite from one of these intrusions yielded a K–Ar age of 1,845 m.y., which indicates that the rocks are Aphebian and were folded during an early phase of the Hudsonian Orogeny.

The basal Sosan Formation is from 3,000 to 5,000 feet thick and composed of feldspathic quartzite, a sandy stromatolitic dolomite member, and, near the top, red or pink siltstone and shale. The Kahochella Formation is as much as 7,000 feet thick and consists of red shale, siltstone, arkose, carbonate, and a few beds of oölitic iron-formation. In the central part of the basin, andesite flows and pyroclastics with intercalated red sandstone and shale occur within the sequence. The Pethei Formation comprises about 1,500 feet of limestone and dolomite with abundant stromatolitic zones (Pl. IV-11) along the north limb of the synclinorium. Lateral equivalents (Hoffman, 1967) in the southern part of the basin are fine-grained limestone and interbedded limestone and argillite and in the southwest are graded beds of greywacke and shale, with interbeds of chert and limestone (Pl. IV-12). Conformably overlying the Pethei are red mudstone, dolomite, and limestone forming the Stark Formation. Breccias, probably of different origins, are common in the carbonate rocks. The Tochatwi Formation comprises about 3,000 feet of non-marine, molasse-type red sandstone and shale. The youngest rocks of the Great Slave Group form the Pearson Formation which consists of 500 feet of basalt, commonly vesicular and amygdaloidal, with related tuff and thin beds of argillite. Orientations of crossbedding and other current structures suggest a source area to the west and north of the present basin for most clastic sediments

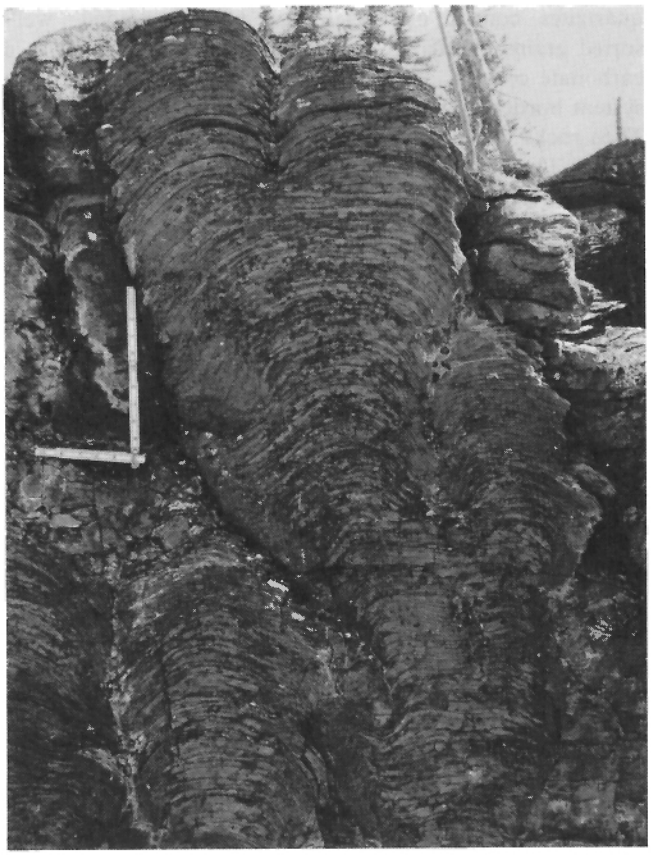

PLATE IV-11. Algal stromatolites in the Aphebian Pethei Formation, Great Slave Group, East Arm Fold Belt, Great Slave Lake, Northwest Territories.

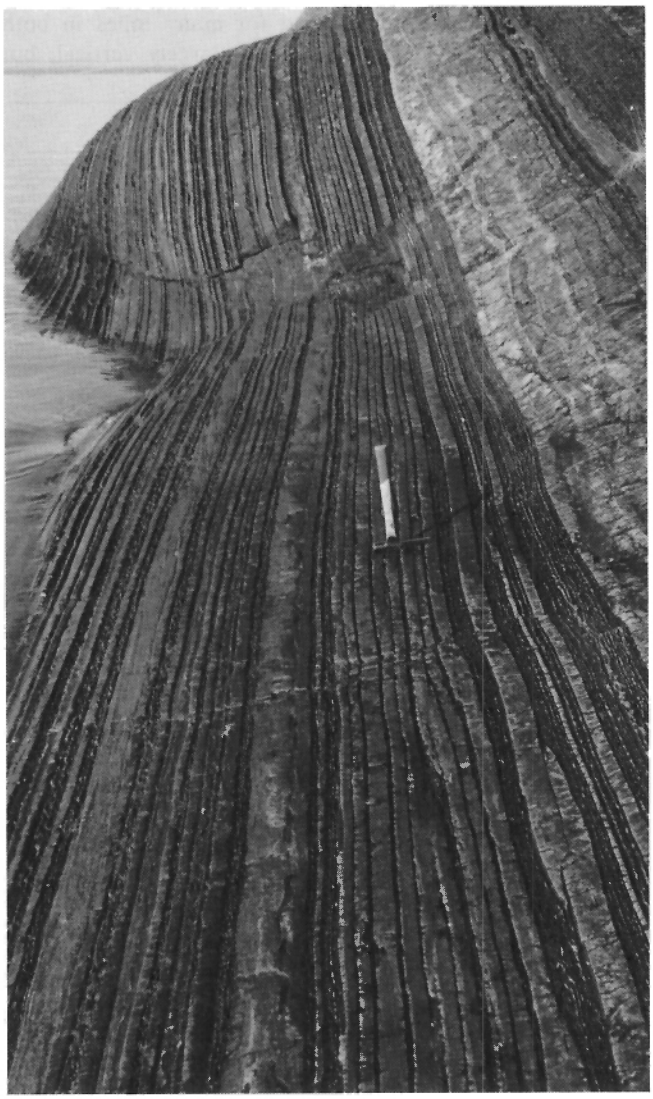

PLATE IV-12. Thin, graded beds of greywacke and argillite of Aphebian Pethei Formation, Great Slave Group, Great Slave Lake, Northwest Territories. The thick light bed is chert.

(Fig. IV-17). The sequence includes near-shore sediments deposited in fluvial non-marine, shallow marine, and deep-water marine environments. The youngest reddish sediments were deposited in a shallow-water, near-shore, partly marine environment and were subject to subaerial exposure.

The Great Slave is overlain unconformably by the Et-Then Group, which rests on truncated folds in Great Slave strata and is later than the diorites that intrude the older rocks. Et-Then rocks are cut by diabase sills and dykes. The dykes are part of the north-northwesterly striking Mackenzie swarm that yield K–Ar whole rock dates of about 1,200 m.y. but the sills which are cut by these dykes are not dated. Et-Then rocks are gently folded about easterly trending axes and along with older rocks are cut by the northeast-trending McDonald fault and related parallel faults (Fig. IV-16). The McDonald fault *(Frontispiece)* forms the southeastern margin of the subprovince and extends beyond for many miles in both directions. Movement appears to be largely vertical, but the history is more complex as both transcurrent and normal movements may have taken place. Most displacement appears to have been completed before the intrusion of the Mackenzie diabase dykes and probably began during the late stages of the Hudsonian Orogeny.

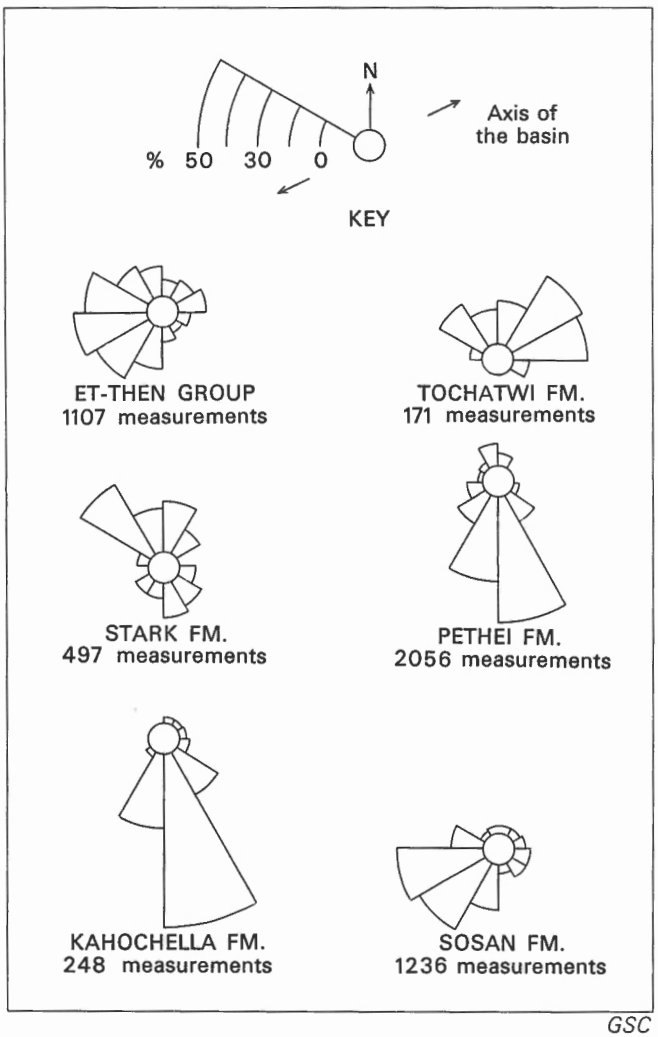

FIGURE IV-17. Paleocurrent orientations, Great Slave and Et-Then Groups, East Arm Subprovince, Northwest Territories (Hoffman, 1967).

The Et-Then Group is divided into two formations, the oldest of which, the Murky, is composed of several thousand feet of conglomerate with minor arkose and shale. The sequence is made up of cyclical alternations of coarse boulder-conglomerate that grade upwards, with decreasing pebble-size into thin, commonly lenticular arkose and rare shale beds. The cycles vary in thickness up to 50 feet. The boulders and pebbles are rounded and vary in composition. Near McDonald fault, granitic rocks and granitic rocks and granite gneiss predominate, whereas to the north, clasts of the Great Slave Group are most abundant. The conglomerate may thin to the north and the clasts appear to decrease in size in the same direction suggesting a source to the south. These strata pass through a transition zone in which the amount of sandstone increases into the Preble Formation which consists of at least 4,000 feet of buff, pink, or red lithic sandstone and feldspathic quartzites with intercalated thin beds of pebble-conglomerate, shale-chip conglomerate, and red shale. Crossbedding and thin laminations are common and mudcracks are found throughout the sequence. The quartzites consist of rectangular to subrounded, well-sorted grains of quartz, rock fragments, and feldspar in a carbonate cement. It is possible that the McDonald fault system bordered the original sedimentary basin. The Et-Then rocks were deposited in a non-marine fluvial environment. They are typically late or post-tectonic clastics and were probably deposited during a late phase of Hudsonian Orogeny, or after it.

North-central Churchill Province

The north-central part of Churchill Province extends to latitude 66° and includes northern Manitoba and northeastern Saskatchewan. The oldest rocks occur in a northeasterly trending belt that extends through Ennadai Lake. They unconformably underlie the Aphebian Hurwitz Group and are considered Archean. They consist of a conformable sequence of basic to intermediate volcanic rocks and greywacke argillites and tuffs. The volcanic rocks are massive or pillowed andesite with minor amounts of intercalated dacites and quartz biotite rhyolite. Tuff and agglomerate are associated with the acidic phases and irregular masses of medium-grained gabbros occur within the basic flows. Greywacke, subgreywacke, and shale occur both below and above the volcanic rocks (Eade, 1966b). The sediments are thin- to thick-bedded and graded beds are abundant. Locally, a considerable thickness of lithic or crystal tuff occurs within the sedimentary sequence. One thin band of impure shaly dolomite has

been mapped and iron-formation, usually a quartz–magnetite type, occurs as discontinuous lenses up to 150 feet thick within the greywacke.

The Archean rocks are folded about north-northeast to northeasterly trending axes. The folds generally plunge north. Near or within large masses of granitic gneiss, the basic volcanic rocks are converted to massive or well-foliated amphibolites and the sediments to quartz–biotite–feldspar schist and gneiss. The rocks commonly grade to granitic gneiss through a zone of migmatites and mixed gneisses. The granitic rocks are either well-foliated, massive, or porphyritic, pink or grey, and contain biotite or less commonly hornblende. Their composition is generally that of a granodiorite. The volcanic and sedimentary rocks and possibly the granitic gneisses are cut by high-level nearly massive granodiorite and quartz diorite plutons that are bounded by sharp contacts or narrow migmatite zones and metamorphic aureoles. Cordierite, garnet, and in places andalusite are developed in the sediments of the aureoles. In the Kognak Lake area (Eade, 1966b) the granitic rocks and the Archean strata are overlain unconformably by the Aphebian Hurwitz Group. Some granitic rocks are thought to intrude Hurwitz strata and probably two ages of granites exist. Micas from granitic rocks in the region yield K–Ar dates in the range 1,675 to 1,800 m.y. One date of 2,375 m.y. suggests Kenoran intrusions.

The Hurwitz Group underlies scattered areas along a northeasterly trending belt that roughly coincides with the belt of Archean rocks and also occurs north and northwest of Baker Lake (Wright, 1955). The group, as defined by Eade (1966), comprises five conformable formations. The lowest, a local conglomerate, contains rounded pebbles or cobbles of granitic rocks, andesite, greywacke, quartz, and tuff in a greywacke matrix, grading upwards into greywacke with a few scattered pebbles. The overlying orthoquartzite is fine to medium grained, glassy, white, thin to thick bedded, and contains scattered lenses of quartz-pebble conglomerate. Crossbedding is rare but ripple-marks are common at several horizons and have a preferred northeasterly orientation. The formation varies in thickness from 700 to at least 4,000 feet commonly over short distances. Poorly exposed slate, shale, and siltstone overlie the quartzite. They are grey, black, or red, finely laminated, and contain lenses of greywacke. These rocks are succeeded by dolomite, argillite, and siltstone. The dolomite is cream or grey laced with quartz or chert veinlets, and contains a few, thin, stromatolitic horizons. It grades upwards into black or grey, finely laminated argillite with siltstone interbeds. Locally quartz–jasper–hematite or quartz–magnetite iron-formation occur as lenses within the dolomite. These rocks are generally overlain by a variable thickness of greywacke composed of thin-graded sandstone and shale couplets. The youngest exposed formation of the Hurwitz Group consists of grey to buff impure quartzite composed of quartz and sericite, thinly bedded, ripple-marked and crossbedded. A few thin horizons of siliceous dolomite are intercalated. North and northwest of Baker Lake Hurwitz rocks have a similar stratigraphic sequence. They are locally overlain by basic volcanic rocks and intruded by gabbro sills of unknown age.

The Montgomery Lake Group unconformably underlies the Hurwitz Group and unconformably overlies the Archean in the Kognak Lake area (Eade, 1966b). It consists of quartzite, dolomite, siltstone, and argillite. The group is not extensively exposed and little is known about it, but in northern Manitoba there may be some correlatives.

The Aphebian sediments and igneous rocks are folded moderately about northeastward and locally northwestward trending axes. The sequence of folding is uncertain but the northeasterly trend may be the oldest. Contacts between the Hurwitz and Archean rocks are often sheared indicating movement. Hurwitz strata are for the most part only very slightly metamorphosed. In the Kognak River area Eade found one granodiorite pluton bounded by a narrow metamorphic aureole in which cordierite and garnets are developed in the sediments and, therefore, he considers the granodiorite to be post-Hurwitz.

In northeastern Saskatchewan and northern Manitoba are sediments similar to the Hurwitz. In northern Manitoba pyroxene-bearing granulites appear to be older (Davison, 1966). The sediments comprise quartzite, feldspathic quartzite, argillite, greywacke, and dolomite. Orthoquartzites so characteristic of the Hurwitz are much less common and do not form basal units. The oldest rocks appear to be feldspathic quartzite and greywacke. These rocks are intensely folded about northeasterly to easterly trending axes. Grade of metamorphism varies but usually is rather high, commonly in the amphibolite facies. The metasediments grade into biotite–quartz–feldspar gneiss that contains potash feldspar, masses of pegmatitic granite, and migmatite. These rocks grade to foliated and locally massive granodiorite. All rocks are cut by high-level, intrusive, porphyritic, or fluorite-bearing granodiorite. Micas from all granitic rocks yield K–Ar ages for the range 1,700 to 1,850 m.y. as does a Rb–Sr whole rock isochron of the fluorite-bearing granodiorite and they are considered to have been formed during the Hudsonian Orogeny.

On Seal River in northeastern Manitoba the Great Island Group comprises a conformable sequence of shale, quartzite, dolomite, and minor greywacke (Davison, 1966). These rocks are known to be exposed in only a small area. They are fresh to very slightly metamorphosed, are not known to be cut by granitic rocks, and are gently folded about north-northwesterly fold axes. They appear to lie unconformably on the surrounding granitic rocks and older sediments. If older sediments are Aphebian this sequence is probably of Helikian age. An alternate correlation is that of Great Island rocks with the Hurwitz, whereas the rocks below the Great Island sequence are equivalent to the Montgomery Lake strata.

In Thelon Subprovince, the Dubawnt Group comprises relatively undisturbed sedimentary and volcanic rocks that rest with profound unconformity on a deeply eroded basement underlain for the most part by granitic rocks of Kenoran and possibly Hudsonian age. Donaldson (1965) has divided the group into six formations. The oldest, the South Channel Formation, consists of conglomerate with a few interbeds of sandstone, varying rapidly in thickness to a maximum of 5,000 feet. It is composed of rounded to angular pebbles and cobbles of granitic rocks with minor quartzite, quartz, granulite, mylonite, basalt, and felsite in a red sandy matrix. The overlying Kazan Formation comprises about 12,000 feet of maroon pink to reddish brown feldspathic sandstone with calcareous cement and minor amounts of interbedded maroon siltstone and mudstone. The sandstones are well

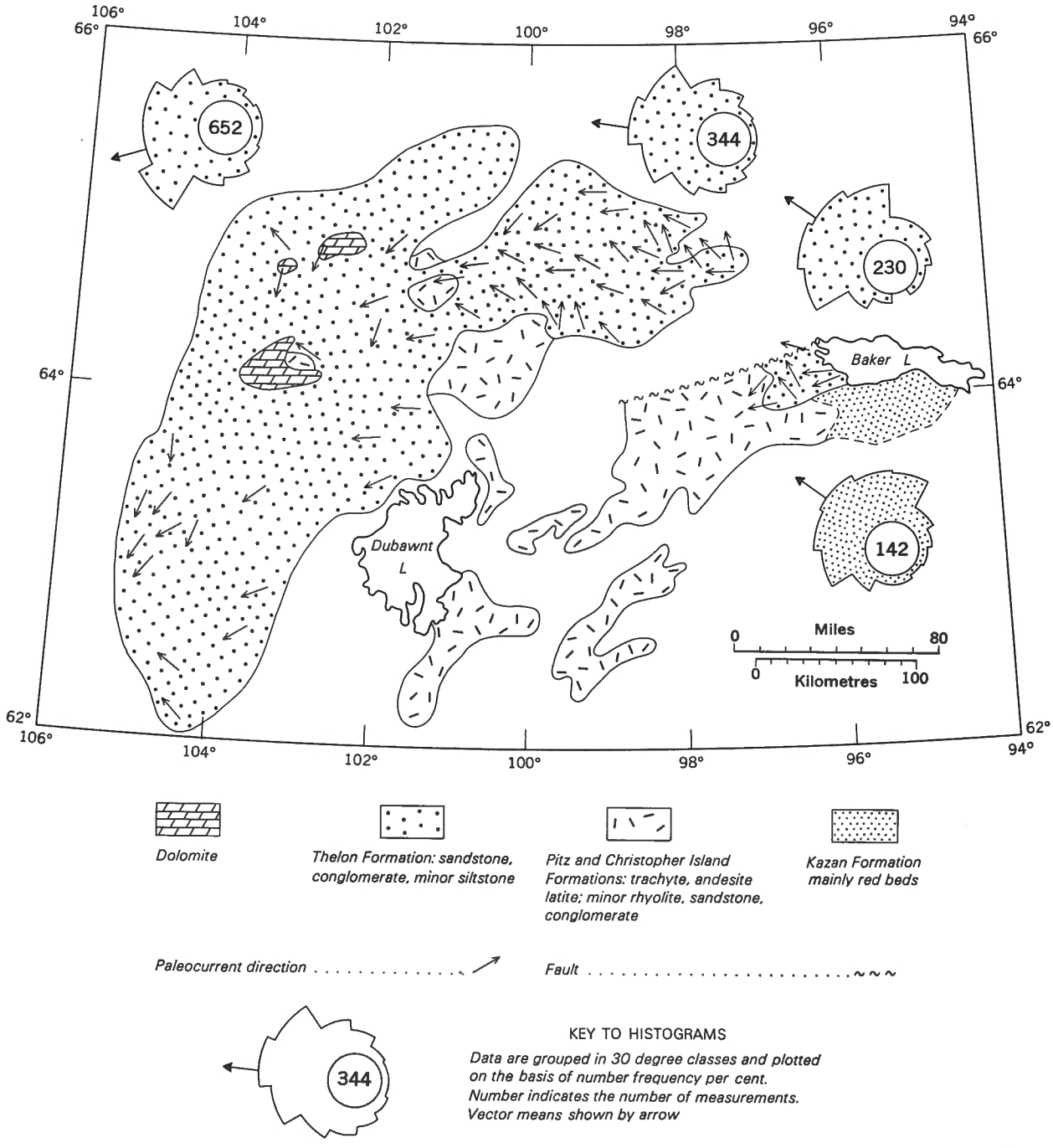

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FIGURE IV-18. Paleocurrent orientations, Dubawnt Group, north-central Churchill Province, Northwest Territories (J. A. Donaldson).

sorted and display thin laminations and crossbedding, the orientations of which indicate northwesterly transport of sediment (Fig. IV-18). Mudcracks and ripple-marks are abundant in the silty shale. South Channel and Kazan strata occur only south of Baker Lake. These rocks were gently folded and eroded before volcanic rocks of the Christopher Island Formation were extruded. Reddish brown or purplish grey porphyritic andesite, latite, and trachyte form the bulk of this formation which attains an exposed thickness of about 200 feet. Basal tuff and agglomerate contain cobbles of older formations. A second volcanic unit, the Pitz Formation, is composed of trachyte characterized by abundant chalky white phenocrysts of feldspar and quartz phenocrysts in a purple to red or maroon groundmass. The volcanic rocks have been intruded by sills or laccoliths of massive, fine- to medium-grained, grey to reddish black syenite. The volcanic rocks appear to be only slightly deformed. Whole rock samples of these rocks yield K–Ar and Rb–Sr whole rock isochrons up to 1,735 m.y., a minimum age for the sequence. These rocks are overlain disconformably by the relatively flat lying sequence of the Thelon Formation. Thin basal conglomerates contain pebbles of the Kazan, Pitz, and Christopher Island Formations. The Thelon sandstone is composed of quartz but near the base contains feldspar and interstitial white clay. The sandstone is cream, light grey, buff or pink, and in contrast to the Kazan, is friable, ripple-marked, and festoon crossbedded. Orientations of crossbeds indicated westerly and northwesterly transportation of sediment (Donaldson, 1965). Well-rounded pebbles and bands only one pebble thick are common in the lower part. The Thelon Formation is overlain by dark grey, thinly bedded dolomite, partly stromatolitic, and locally by a thin layer of basalt. The stratigraphic relationship between the two younger units is uncertain.

The sediments of the lower part of the Dubawnt south of Baker Lake comprise a red bed sequence and were probably deposited in a continental and possibly partly shallow marine environment and subjected to frequent subaerial exposure. Westward thinning of the South Channel conglomerates and the crossbedding orientations suggest a source to the east and southeast with northwesterly and westerly sediment transport. The immature nature of the clastic sediments suggests rapid deposition from a tectonically unstable area of rather rugged relief. Faults may have partly controlled the location of the southern margin of the basin and the rate of sediment accumulation. The sedimentary and volcanic rocks are typical of late or post geosynclinal and tectonic environments. The age of the volcanic rocks suggests that sedimentation and volcanism occurred either late in the Hudsonian Orogeny or just after the orogeny and, therefore, these rocks are considered to be late Aphebian or Paleohelikian. The Thelon sediments, which were deposited after the above rocks were eroded to some extent, are compositionally and texturally mature, and probably were deposited on a surface of low relief and under stable tectonic conditions. The sandstones probably were deposited in a fluvial environment. Sedimentary structures indicate a source area to the east or southeast. Possibly the upper part of the Thelon and the carbonate sediments were formed in a shallow-marine environment. These rocks are cut by diabase dykes of Helikian age and are, therefore, considered to be Paleohelikian.

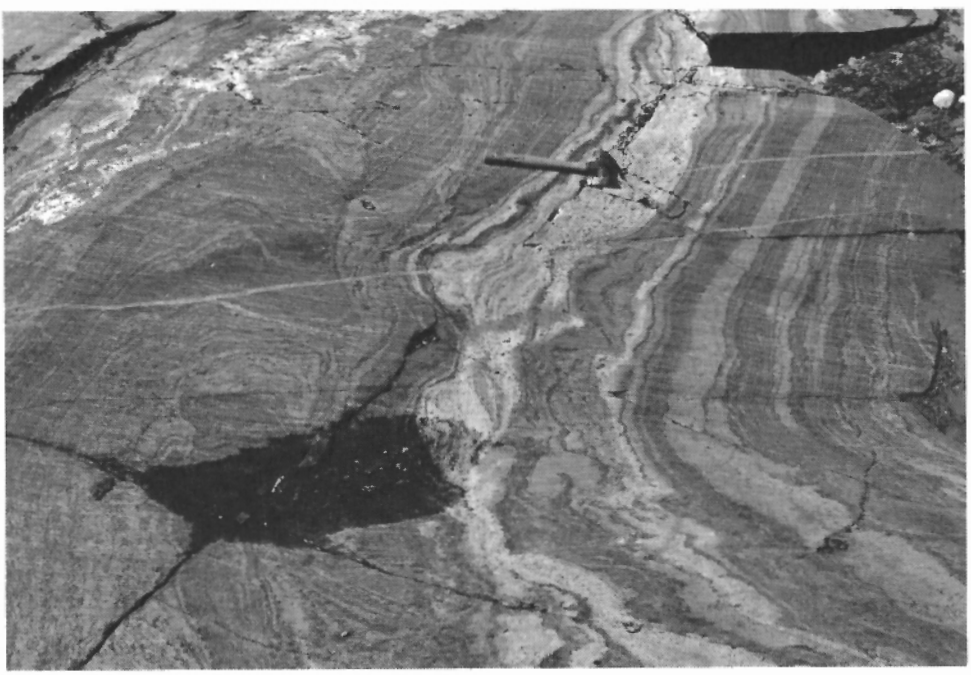


PLATE IV-13

Banded metasedimentary gneiss, possibly Archean, near mouth of Back River, northern Churchill Province, Northwest Territories.

Northern Churchill Province

The northern part of Churchill Province includes the Precambrian rocks of the eastern Arctic Islands, Boothia Peninsula, and the northeastern mainland west of Hudson Bay. Geology of the mainland part of this area has been studied by Fraser (1964), Heywood (1961, 1966), and Blackadar and Christie (1963).

Northeastern mainland. Much of the region is underlain by granitic rocks of various types that include pink or grey granitic gneiss with zones of migmatite, mixed gneiss, and highly metamorphosed volcanic and sedimentary rocks (Pl. IV-13). The gneissic rocks are variable in composition and may be rich or poor in hornblende or biotite, and otherwise consist of plagioclase and quartz with varying amounts of microcline partly porphyroblastic. These rocks are intruded or grade into foliated to massive granodiorite or quartz diorite that are equigranular and less commonly porphyritic. Foliation trends north to north-northeast in the western part of the area but to the east gradually changes to northeast to easterly. East of the Bathurst Inlet and on Wager Bay and Melville Peninsula are ill-defined zones of pyroxene-bearing gneisses or granulites of uncertain origin. Small masses of gabbro, peridotite, and anorthositic gabbros are scattered throughout the region. K–Ar age determinations on micas from the granitic rocks of all types yield ages that fall in the range about 1,650 to 1,850 m.y.

Within the gneisses are zones of metamorphosed volcanic and sedimentary rocks of various types and probably different ages. South of Queen Maud Gulf northerly to north-northeasterly trending bands of amphibolite with associated biotite gneisses are probably metamorphosed equivalents of basic volcanic and sedimentary rocks. Calc-silicate, quartzite, and iron-formation are present locally. These rocks may be polymetamorphosed equivalents of the Archean Yellowknife Group. North of Wager Bay and on Melville Peninsula are variably metamorphosed basic volcanic rocks that only locally contain primary structures. Metamorphosed sediments, now biotite schists, tuffs, and locally quartzite and iron-formation are intercalated. These formations seem to be remnants of an eugeosynclinal sequence and are similar to rocks in the Ennadai belt to the south that may be Archean. The rocks are folded about northeasterly trending axes.

The Chantrey Group occurs in a northeasterly trending belt south of Boothia Peninsula and consists of a succession of crystalline limestone and quartzite with lesser amounts of conglomerate, shale, greywacke, and chert. These rocks are deformed into northeasterly trending folds and are slightly metamorphosed. Contacts with granitic rocks are obscured by shearing, which suggests separation of the two units in many places by faults. Pegmatites cut the sedimentary rocks. Chantrey rocks are lithologically similar to rocks of the Hurwitz Group to the south and may be correlative. On Melville Peninsula is an apparently conformable sequence of sediments in which quartzite and limestone predominate in the lower part and quartzite, argillite, and greywacke in the upper. These rocks are folded about easterly to northeasterly trending axes, and may have been refolded. The sediments are metamorphosed and cut by or altered to granitic gneiss. Locally the metasediments are present in domal structures, mantling cores of foliated granitic gneisses that range in composition from granodiorite to granite (Heywood, 1966). Contacts between core gneisses and the sediments are generally sharp but locally migmatite occurs at the contact. The sediments are cut by granitic rocks that yield Hudsonian dates and were probably deformed and metamorphosed during this orogeny. They are assumed to be Aphebian in age. Granitic rocks in the area are partly older than the sediments and may possibly be Kenoran intrusions that have been involved in the Hudsonian Orogeny.

On northern Boothia Peninsula and Somerset Island, unconformably overlying the mixed gneisses, is a conformable sequence of unfossiliferous sedimentary strata comprising the Aston and Hunting Formations. The Aston Formation consists of as much as 2,000 feet of well-sorted red sandstone with minor pebble bands and shale-chip conglomerate. Ripple-marks and crossbedding are common. A thin basal member of white quartzite is overlain by a thin stromatolitic dolomite bed. The contact with the overlying Hunting Formation is gradational and conformable. It comprises 8,000 feet of fine-grained dolomite with minor interbedded shale and some stromatolitic horizons. These sediments are thought to have been deposited in a marine, shallow oxidizing environment. Orientation of crossbeds suggest transport to the northeast away from basement rocks on Boothia Peninsula. These rocks are cut by numerous sills and dykes of diabase particularly in the Aston rocks. Blackadar and Christie (1963) suggest a Proterozoic age because of lithologic and stratigraphic similarities to other Proterozoic sequences in the Arctic Islands, and because they are cut by diabase dykes not found in the fossiliferous strata. Tuke, *et al.* (1966) suggest a Paleozoic age because they appear to be overlain conformably by Ordovician rocks and are lithologically similar to Cambrian rocks on Boothia Peninsula. The evidence for either age is inconclusive.

Eastern Arctic Islands. The oldest Precambrian rocks in Baffin Island are a heterogeneous assemblage of complexly folded granitic gneisses that include bands of highly metamorphosed sediments and volcanic rocks, and which are intruded by relatively small masses of faintly foliated to massive granodiorite. Those in the north are composed of plagioclase, quartz, microcline, varying amounts of biotite and hornblende, and locally some garnet. Zones of migmatite and mixed gneisses occur within the gneisses. Amphibolites that are metamorphic equivalents of basic to intermediate volcanic flows and related tuffs, and biotite schists that are probably metamorphosed greywacke, occur as thin discontinuous bands. Iron-formation and small masses of ultrabasic rocks are found in a number of places.

Foliation has a regional trend of north to northeast. The grade of regional metamorphism is amphibolite facies. The gneisses on Devon and Ellesmere Islands are similar to those in northern Baffin Island. The gneisses on southern Baffin Island are more variable and include ill-defined zones of pyroxene- and hornblende-bearing granulite. They are intruded by massive granitic rocks. Metaquartzite, crystalline limestone, and metamorphosed pelitic rocks are common. Regional trends of foliation are variable but in a general way trend southeast. In large areas the grade of metamorphism is amphibolite facies.

Micas from granitic gneisses both in the north and south yield K–Ar dates that range from 1,590 to 1,975 m.y. but most samples range from 1,700 to 1,800 m.y., indicating that the gneisses were formed during the Hudsonian Orogeny, or affected by it. The age of the metamorphosed volcanic and sedimentary rocks is uncertain. Jackson (1960) has pointed out the similarity between the iron-formations in northern Baffin Island and those on Melville Peninsula. The associated volcanics and sediments of both areas are also similar and the structural trends are consistent over the region. Such rocks may be Archean and the younger gneisses may have been formed during the Kenoran Orogeny and then affected by the Hudsonian Orogeny. In southern Baffin Island the remnants of metamorphosed sediments include much quartzite and limestone and this sequence may correlate with similar Aphebian rocks on Melville Peninsula. The granitic gneisses would then have formed during the Hudsonian Orogeny.

Gently deformed and relatively unmetamorphosed Helikian and Hadrynian rocks occur in the northwestern part of Baffin Island in Borden Subprovince and north of Fury and Hecla Strait. The Eqalulik Group (Lemon and Blackadar, 1963; Blackadar, 1965) is composed of a basal formation of massive and amygdaloidal basalt about 1,000 feet thick, with thin beds of tuff and quartzite and an upper formation of more than 4,000 feet of pale orange to reddish brown quartzite with arkose and quartz-pebble conglomerate. The uppermost 50 feet is shale and quartzite. The group is conformably and transitionally overlain by the Uluksan Group, a conformable sequence 23,000 feet thick. The Arctic Bay Formation is the oldest of seven formations in the group and consists of 1,000 feet of black argillaceous limestone and dolomite that grade laterally into the Fabricius Fiord Formation of black shale, siltstone, and sandstone. These formations are succeeded by 1,000 feet of massive grey dolomites of the Society Cliffs Formation. The Victor Bay Formation is composed of dark to light grey dolomite, chert, and dark grey limestone and mudstone and contains units of edgewise conglomerate in which the dolomite fragments occur in finely laminated dolomite. Red mudstone, shale, siltstones, and sandstones form the Strathcona Sound Formation conformably overlying the Victor Bay. The Elwin is the youngest formation of the Uluksan Group and comprises a sequence of reddish or orange quartzites, siltstones and shales. Near Tremblay Sound, Victor Bay strata are overlain by grey and black limestones and argillaceous limestone of the Athole Point Formation. The relation of these rocks to other units younger than the Victor Bay strata is unknown.

Along the shores of Fury and Hecla Strait, 10,000 to 15,000 feet of sediments is exposed. The lowest unit that unconformably overlies basement gneisses is a thick succession of orange to pale and dark red sandstone with minor shale and conglomerate. These rocks may be equivalent to the Eqalulik Group. They are conformably overlain by about 1,500 feet of black siltstones, shales, and ferruginous dolomite, similar to the lower Uluksan Group.

The Eqalulik and Uluksan Groups are cut by diabase dykes, most of which trend northwest, and by diabase sills. The basic rocks do not cut overlying Cambrian and Ordovician strata. K–Ar age determinations on whole rock samples of these rocks yield ages of 475 and 693 m.y. in Fury and Hecla Strait area and 635, 915, and 1,140 m.y. near Arctic Bay. One sample of the Eqalulik volcanic rocks gives an age of 903 m.y. The range of these dates is puzzling but when considered along with ages of other gabbroic and basaltic rocks in the Arctic region a possible interpretation is that there were two ages of diabase intrusion; an older with a probable age of 1,100 to 1,200 m.y. followed by younger late Hadrynian. The Eqalulik strata are probably Paleohelikian in age and the Uluksan Group is Neohelikian or Hadrynian.

Labrador Subprovince

Labrador Subprovince is a belt of folded Aphebian rocks, as much as 65 miles wide, extending south of Ungava Bay for more than 600 miles. The belt probably represents only a small part of the original Aphebian geosynclinal deposits. Aphebian rocks in the south that have been involved in the Grenvillian Orogeny are described in the section on Grenville Province. The western boundary of the geosynclinal sequence in most places is a well-defined unconformity along which relatively unmetamorphosed sedimentary rocks rest on Archean crystalline rocks of Superior Province that have yielded K–Ar ages of 2,400 to 2,500 m.y. The eastern boundary is less well defined, and, although a basal unconformity has been recognized at the northern end of the belt, and for short distances in several areas in the south, the eastern boundary along much of the central region is marked by either a major fault or a zone within which metamorphic grade increases eastward. K–Ar ages of 1,600 to 1,800 m.y. predominate in the metamorphic terrain to the east and similar ages have been obtained for low-grade metamorphic rocks within the geosyncline. These ages indicate that the major deformation of the Labrador Geosyncline took place during the Hudsonian Orogeny, and that the deformed rocks are Aphebian. Early accumulation of the geo-

synclinal sequence followed closely by metamorphism prior to the main orogenic event is suggested by a date from the Aphebian strata of 2,060 m.y. (Beall, *et al.*, 1963). Other dates that fall between 1,400 and 1,500 m.y. may record a final period of low-grade metamorphism. The Aphebian rocks comprise a western sedimentary sequence, and an eastern volcanic sequence that contains numerous gabbroic and ultrabasic sills. Gabbroic sills also occur in the western sequence, but there they are thinner and rather rare. In the vicinity of Schefferville the major sedimentary, volcanic, and intrusive units are, respectively, the Knob Lake, Doublet, and Montagnais Groups. Stratigraphic nomenclature for these and other units has recently been reviewed by Frarey and Duffell (1964), who proposed the formal rock-stratigraphic name, Kaniapiskau Supergroup, to designate the assemblage of all sedimentary and extrusive units of the Labrador Geosyncline.

The Knob Lake Group is a miogeosynclinal sequence consisting mainly of sandstone, siltstone, shale, dolomite, chert, and iron-formation. It has a maximum thickness of more than 5,000 feet, but thins westward, and several units wedge out near the western boundary of the geosyncline. To the east is the Doublet Group, a eugeosynclinal sequence consisting mainly of basaltic flows and basic pyroclastics, that is estimated to be more than 15,000 feet thick. Volcanic rocks are intercalated with sedimentary units of the Knob Lake Group, indicating some volcanism concurrent with sedimentation, but the Doublet Group apparently formed subsequent to the main period of sedimentation. In the vicinity of Ahr Lake, Baragar (1960) has demonstrated that the gabbros and basalts are predominantly tholeiitic, and this composition may prevail. Gabbro and peridotite sills of the Montagnais Group were emplaced prior to, and during the early stages of, the orogenic episode that resulted in folding about axes parallel with the trend of the geosyncline. Composite sills have been recognized and many of the sills are differentiated.

In the Schefferville area the Knob Lake Group comprises the Attikamagen, Denault, Fleming, Wishart, Ruth, Sokoman, and Menihek Formations. Of these, the Denault Formation, consisting largely of dolomite, and the Sokoman Formation, containing the main units of iron-formation, are particularly distinctive and serve as markers for regional correlation. Stratigraphic sequences throughout the geosyncline show close similarity to the sequence established for the Schefferville area. Differences, variously attributable to non-deposition, local unconformities, and facies changes, have been recognized, but regionally uniform conditions of deposition are indicated for most units. Some clastic units show marked variability of thickness, and for basal arenaceous strata this characteristic appears to be related to initial topographic relief. Southeast of Schefferville the Attikamagen Formation is underlain by the Seward Formation composed mainly of arkose and conglomerate. Near Cambrian Lake to the north, similar sedimentary rocks form a basal sequence probably more than 5,000 feet thick. Faults that flank areas underlain by these rocks may have been active during deposition of the immature sediments, and nearby exposures of basement within the geosyncline suggest the existence of rugged landmasses during the early stages of geosynclinal deposition. Younger arenaceous units such as the Wishart Formation are orthoquartzitic, reflecting stable shelf conditions and probable low relief of adjacent landmasses. The Wishart sandstones are typically well sorted, which suggests deposition in shallow agitated water, and similar environments are indicated for the locally arenaceous and stromatolitic Denault Formation, penecontemporaneously reworked Wishart cherts, and the granule- and oölite-rich beds of the Sokoman Formation. Three main facies are recognized in the Sokoman Formation: a magnetite–silicate facies, a magnetite–hematite, and a silicate–carbonate (Gross, 1961). The Sokoman Formation, more than 600 feet thick over wide areas, is predominantly cherty, and is the host rock for the extensive iron ore deposits that developed by natural enrichment processes, probably for the most part during the Mesozoic Era. Enrichment, possibly related to Cretaceous or post-Cretaceous faulting, is recorded near Schefferville by small trough-fillings of fossiliferous iron-formation comprising the Redmond Formation. This formation, more than 300 feet thick, consists of reworked iron-formation rich in jasper fragments enveloped by iron-oxide layers, poorly consolidated pink to pale grey clay, and minor fossiliferous argillite (Blais, 1959). The formation is bounded by faults and rests unconformably on Kaniapiskau strata.

The lowermost unit of the Doublet Group is the Murdoch Formation. A basal unconformity is indicated by lenses of polymictic conglomerate that contain fragments of Knob Lake strata. The predominant lithologies are basic tuff and agglomerate that commonly are altered to chloritic schists. Sedimentary rocks of the Doublet Group are included mainly in the Thompson Lake Formation, a unit probably less than 2,000 feet thick. Although texturally and compositionally immature greywacke characterized by graded bedding occurs in this formation, moderately sorted lithic sandstones containing primary structures indicative of deposition in shallow water are locally abundant (Donaldson, 1966). These rocks may have been partly derived from the cratonic landmass to the west. The overlying Willbob Formation, more than 15,000 feet thick, consists of basaltic lavas that are commonly pillowed, thin sedimentary interbeds, and intercalations of tuff, flow breccia, and agglomerate.

Axial planes of folds typically are inclined east or northeast, and some folds, particularly within the central part of the belt, are overturned. In contrast to the less competent sedimentary rocks, the stratified igneous rocks are characterized by simple, large scale folds. Thrust faults parallel with the axial planes, formed both during and after the period of folding, are particularly abundant in areas of intensely folded sedimentary strata. As many as five generations of folds and cleavages have been

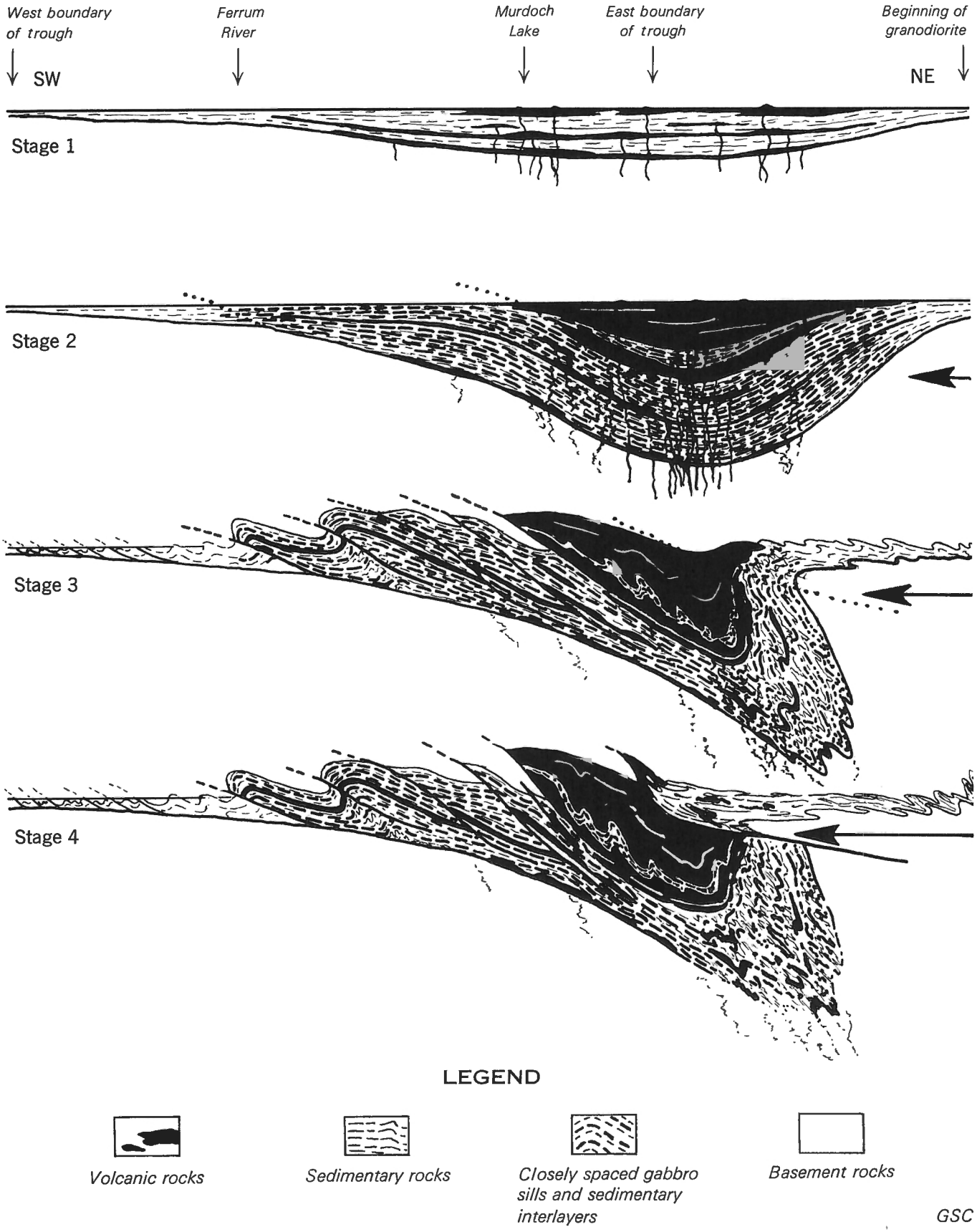


FIGURE IV-19. Schematic cross-section of the Labrador Fold Belt at successive stages in its development (by W. R. A. Baragar).

recognized in some areas. The probable stages of development of the orogenic belt are shown in a series of schematic cross-sections (Fig. IV-19). Metamorphic grade increases northward, as the Kaniapiskau rocks west of Ungava Bay are mainly in the epidote–amphibolite facies. In the north, the deformed Aphebian rocks terminate in a broad syncline in which beds of iron-formation can be traced continuously around the structure (Béland and Auger, 1958). In the south the geosynclinal sequence has been involved in deformation and metamorphism related to the Grenvillian Orogeny.

Because the eastern part of the region has not been studied in sufficient detail, the relative abundance of migmatized and granitized Kaniapiskau strata in relation to basement rocks is unknown. Northeast of Schefferville, metasedimentary rocks of the Laporte Group, predominantly biotite schists, are separated from Kaniapiskau strata by the eastern boundary fault, and in all probability are Kaniapiskau equivalents (Fahrig, 1957). Biotite-, amphibole-, and pyroxene-bearing quartzo-feldspathic gneisses and granites that occur east of the Laporte rocks and which in some areas are in direct fault contact with low-grade Kaniapiskau strata may represent basement rocks that were uplifted during the Hudsonian Orogeny, or may be high-grade metamorphic equivalents of the Kaniapiskau Supergroup. Taylor (1968) has recently indicated that metamorphosed rocks derived from Kaniapiskau strata may extend far to the east of the presently defined western segment of the geosynclinal sequence.

Folded strata of the Kaniapiskau Supergroup are unconformably overlain by flat-lying sandstone and conglomerate of the Sims Formation at Sims Lake, 75 miles southeast of Schefferville (Wynne-Edwards, 1961). The Sims consists mostly of white, pink, and yellow orthoquartzite containing local layers and lenses of rounded jasper and vein-quartz pebbles. The formation, at least 700 feet thick, contains some crossbedded units, but characteristically is massive, consisting of well-rounded and well-sorted quartz grains in a microcrystalline matrix of quartz and sericite. Quartzites possibly correlative with the Sims are present south of the main outcrop areas, and show evidence of Grenville deformation, suggesting a Helikian age for the Sims. In this region, large areas are underlain by the Shabogamo gabbro, both north and south of the Grenville Front. Diabasic gabbro and anorthositic gabbro of this unit form discordant bodies texturally and compositionally distinct from sills of the Wakuach gabbro, and intrusive relationships demonstrate that at least some of the Shabogamo plutons are younger than the Sims Formation. The gabbro typically is fresh, containing labradorite, augite, hypersthene, and olivine, but where altered amphibole and garnet are abundant. South of the Grenville Front the gabbros have been transformed to augite–hypersthene gneisses in zones of cataclastic deformation.

In northeastern Superior Province, widely scattered outliers of relatively undisturbed sandstone, conglomerate, and mudstone rest unconformably on the Archean. These outliers, named the Sakami Formation by Eade (1966a), may be equivalent to either the Sims Formation or the basal arkosic beds of the Kaniapiskau Supergroup. Between Cambrian and Chakonipau Lakes about 140 miles northwest of Schefferville, according to E. Dimroth, carbonatites intrude Kaniapiskau strata in diatremes as much as half a mile in diameter, and in north-trending dykes. Their age is unknown. Calcareous tuffs that may be related to these intrusions occur nearby.

Cape Smith Subprovince

The Cape Smith Fold Belt of Proterozoic rocks in New Quebec extends from Smith Island in the northeast corner of Hudson Bay to the shore of Hudson Strait near Wakeham Bay. The 230-mile belt has a relatively linear southern boundary but a sinuous northern boundary, and attains a maximum width of about 50 miles near the middle of the belt. The south and central parts contain two lithological sequences separated by an unconformity (Bergeron, 1959). The lower sequence, the Povungnituk Group, unconformably rests on Archean basement. The overlying Chukotat Group that composes the central part of the belt is separated by a major fault from the northern part of the belt which appears to be an upthrust segment of the Povungnituk Group. The northern limit of the belt is an irregular boundary beyond which migmatitic gneisses are abundant. In the west, parallel thrust faults occur between the Povungnituk Group and the basement, and between the Povungnituk and Chukotat Groups. Much of the eastern part of the belt has not been investigated.

The Povungnituk Group is more than 10,000 feet thick, and consists of basic and intermediate lavas, basic sills, rare ultrabasic sills, and sedimentary rocks that occur both at the base of the sequence and as thin interbeds within the volcanic units. Basalt, locally pillowed, is the most abundant volcanic rock, greatly exceeding andesite and other less basic lavas. The sediments include argillite, black shale, sandstone, limestone, dolomite, minor iron-formation, and schistose, low grade metamorphic equivalents of these rocks. The Chukotat Group has a thickness greater than 15,000 feet, and consists of abundant pillowed basalt and andesite, sills of gabbro and peridotite, and sedimentary rocks that largely are confined to a thin basal unit. The Chukotat sediments include arkose, quartzite, shale, chert, greywacke, and a basal conglomerate rich in fragments of quartzite and dolomite. Folds and faults in the Cape Smith Fold Belt show two main trends. One trend is parallel with the belt and is most pronounced south of the major thrust fault that separates the Chukotat and Povungnituk Groups. Faults and axial planes of folds that show this trend typically are steeply inclined to the north. The other trend is nearly normal to the main trend, and is marked by cross-folds that increase in

development and abundance northward. North of the major fault this second structural trend is as prominent as the first, or more so. The gabbro and peridotite sills were altered largely to meta-gabbro and serpentinite during folding.

The Chukotat Group is generally less metamorphosed than the Povungnituk Group, with greenschist facies prevailing; rocks in the northern part of the belt are predominantly paraschists and amphibolites that, where least metamorphosed, show lithological similarity to units of the Povungnituk Group. The paraschists and amphibolites may be Povungnituk equivalents that have been more intensely metamorphosed and upthrust along the north-dipping fault that marks the northern boundary of the Chukotat Group. In the north these rocks are intruded by small irregular bodies of granite, diorite, and syenite. Migmatitic gneisses that underlie the area extending northward to Hudson Strait are largely of sedimentary origin, and in part may represent a more intensely metamorphosed extension of the Povungnituk Group.

Eight K–Ar ages on rocks of the Cape Smith Fold Belt fall in the range 1,430 to 1,650 m.y. (Beall, *et al.*, 1963). Four samples from the Povungnituk Group give the oldest values, about 1,600 m.y., and, because geological evidence indicates deformation of the Povungnituk Group prior to deposition of the Chukotat Group, this value serves to provide a maximum age for the Chukotat as well as a minimum age for the Povungnituk. Determinations for the other four samples are about 1,450 m.y., the probable age of metamorphism of the Chukotat Group. The Chukotat Group is, accordingly, Paleohelikian. Although K–Ar ages of basement rocks near the basal contact reflect recrystallization concurrent with Povungnituk metamorphism, whole-rock Rb–Sr ages for two samples reveal that the basement rocks immediately south of the belt were involved in the earlier Kenoran Orogeny, and are therefore Archean.

Belcher Subprovince

Along the east coast of Hudson Bay, extending north and south of Richmond Gulf for a total distance of more than 350 miles, Proterozoic sedimentary and basic igneous rocks underlie parts of the coast and nearby islands in a belt as much as 20 miles wide. These rocks dip gently towards the Belcher Islands, and may be older than the thick succession there, or more likely are equivalent to it. Other Proterozoic rocks occur in the Sutton Inlier and other small inliers within the Paleozoic rocks of Hudson Platform.

R. Bell in 1879 recognized an unconformity within the succession bordering Richmond Gulf and the sequences below and above this unconformity subsequently were named the Richmond Gulf and Nastapoka Groups. A second unconformity was reported within the lower sequences by Woodcock (1960), who introduced the name Pachi Group for rocks below this unconformity. Eade (1966) used the name Manitounuk Group for the undivided succession, and also discussed the problems of stratigraphic nomenclature. The Richmond Gulf and Pachi strata are recognized only in the vicinity of the type area. The Pachi Group consists of arkose and conglomerate overlain by a massive andesite less than 100 feet thick. The sediments, more than 500 feet thick, rest on Archean basement rocks along an uneven surface locally mantled by a regolith. The Pachi rocks were faulted and slightly folded prior to deposition of the overlying strata, and appear to be restricted to the vicinity of Richmond Gulf. The overlying Richmond Gulf Group consists of more than 1,500 feet of arkose, argillite, and a local basal conglomerate rich in boulders of the Pachi volcanic rocks. The unconformity at the top of the Richmond Gulf Group is partly obscured by faulting, but in most places, an angular unconformity is evident (Pl. IV-14). Arkoses and associated immature sediments of the Pachi Group probably were deposited rapidly in a continental environment. Kranck (1951) suggested that Richmond Gulf occupies a graben with an east-trending axis, and Woodcock (1960) mapped faults that have this trend within the Pachi and Richmond Gulf sequences, noting their greater abundance in the former. Because none of the major faults extend into the younger Nastapoka rocks, contemporaneous faulting during the interval of red bed sedimentation is suggested.

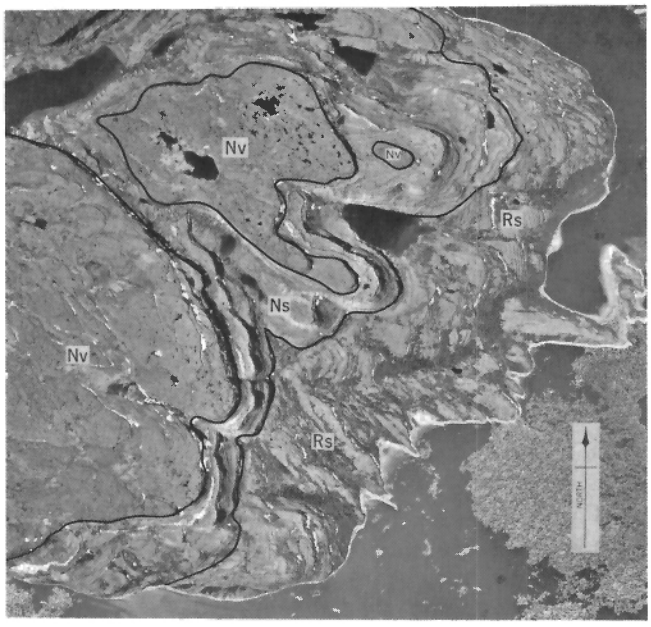


PLATE IV-14. Aphebian Nastapoka Group (Nv–basalt; Ns–sediments) unconformably overlying southwest-dipping Richmond Gulf Group (Rs), Belcher Subprovince, Richmond Gulf, Quebec.



PLATE IV-15
Orbicular diorite sill, in Aphebian Nastapoka Group, Richmond Gulf, Quebec.

The Nastapoka Group extends far to the north and south along the east shore of Hudson Bay, resting nonconformably on Archean crystalline rocks. The trace of this unconformity is relatively linear and commonly marked by a carbonate-cemented conglomerate. Nastapoka lithologies include, in general order of decreasing age, sandstone and siltstone, dolomitic and cherty limestone, quartz sandstone, sporadic lenses of iron-formation, and volcanic rocks with associated gabbro sills, one of which formed phases of orbicular diorite (Pl. IV-15). The Nastapoka strata, in contrast to those of the older groups, must have been deposited in a stable tectonic environment following an interval of erosion sufficient to create an extensive shelf. Sandstone, shale, dolomite, chert, iron-formation, and interlayered basalt compose the Nastapoka Islands, dipping gently westward in apparent conformity with the mainland strata, but at a lower angle. They may represent younger rocks of a continuous sequence, part of which lies beneath Nastapoka Sound, or they may be part of the mainland sequence repeated by vertical uplift or overthrusting. If the sequence is unfaulted, the minimum total thickness of the Nastapoka sequence is more than 2,500 feet.

Proterozoic rocks folded about north- to northeast-trending axes form the Belcher Islands, a cluster of arcuate ridges that outline gently plunging synclines and anticlines in southeastern Hudson Bay (Pl. IV-16). The main group of islands outlines four synclines and four anticlines that are intersected by only a few faults of relatively minor displacement. The synclines are more open than the anticlines, and the folds range from upright-symmetrical in the west to asymmetrical about an eastward-inclined axial plane for the easternmost fold. Grade of metamorphism apparently is in the greenschist facies, but many of the rocks show no evidence of metamorphism. All strata belong to the Belcher Group, within which Jackson (1960) recognized sixteen units of formation rank, including the earlier-named Kipalu Formation. The known thickness of the sequence ranges from about 20,000 feet to almost 30,000 feet but because the base of the sequence is not exposed, the total thickness may be greater. No major unconformities occur, and the lateral continuity of individual units is impressive. The units are relatively uniform in thickness except for the upper volcanic unit, which shows a marked increase in thickness westward. Basic intrusions occur mostly as sills, one of which occurs above the iron-formation and varies little from an average thickness of about 100 feet. The maximum thickness of the other sills is 900 feet, and the thicker parts of these sills are composite. Some basic dykes cut the sills, but because neither sills nor dykes occur in the upper sedimentary sequence, the intrusions probably were emplaced when the flows of the upper volcanic unit were extruded.

The sedimentary rocks for the most part record deposition in shallow water on a stable shelf. Crossbedding and intraformational conglomerates occur in several

units, and stromatolite zones are abundant. In the sedimentary sequence below the iron-formation dolomite and orthoquartzite together greatly exceed the amount of greywacke and argillite. Thick units of basalt and greywacke above the iron-formation reflect less stable tectonic conditions and probably accumulated in a deep-water environment. Hematite-cemented arkose caps the greywacke, suggesting continental deposition, the last-recorded sedimentary event. Three whole rock K–Ar ages, two from cores of basalt pillows and the third from a basic sill, fall in the range 1,600 to 1,700 m.y. Two samples of pillow rims analyzed by the same method gave ages of 830 and 1,019 m.y. G. D. Jackson (1960) regards the younger ages as the result of alteration related to emplacement of young basaltic dykes, one of which has been dated as 880 m.y., and interprets the older values as indicating the minimum age of the Belcher Group. Thus, although the Belcher Group is probably Aphebian, a Paleohelikian age cannot be ruled out.

Although not unequivocally established, correlation of the Belcher and Manitounuk Groups is generally accepted on the basis of information now available. Units of the Richmond Gulf succession, particularly those of the Nastapoka Group, are lithologically similar to units of the Belcher Group. Differences in stratigraphic detail may be attributed to lateral facies changes. Lithological similarities include the presence of iron-formation and stromatolitic carbonate, the abundance of quartzose sandstones, and the widespread flows and sills. The eastward convergence in thicknesses suggests proximity to the border of the geosyncline and possibly also the shore line. The upper basaltic unit of the Nastapoka Group yielded a K–Ar whole rock age of 1,385 m.y. Because the basement to this group is Archean, the single age determination suggests an Aphebian or Paleohelikian age. Bergeron (1957a) noted the lithological similarities and the possibility of continuity of the Aphebian rocks of the Belcher Islands, the Cape Smith Fold Belt, with those of the Labrador Fold Belt and he suggested that they once formed a continuous geosyncline.

In the Hudson Bay Lowland, Proterozoic rocks are exposed as inliers within the unconformably overlying Ordovician and Silurian strata (Sanford, *et al.*, 1968). These Proterozoic rocks are lithologically similar to the succession on Belcher Islands and bordering Richmond Gulf and are probably Aphebian. They dip gently towards Hudson Bay and may be correlative with the latter succession, outcropping also on several small islands in James Bay. Strata of the inliers, more than 200 feet thick, comprise a lower unit of dolomite, in part stromatolitic, and an overlying unit of greywacke, argillite, quartzite, chert, carbonate, and conglomerate. These strata are intruded by sills of gabbroic diabase and rest unconformably on gneisses and granitic rocks, presumably Archean.

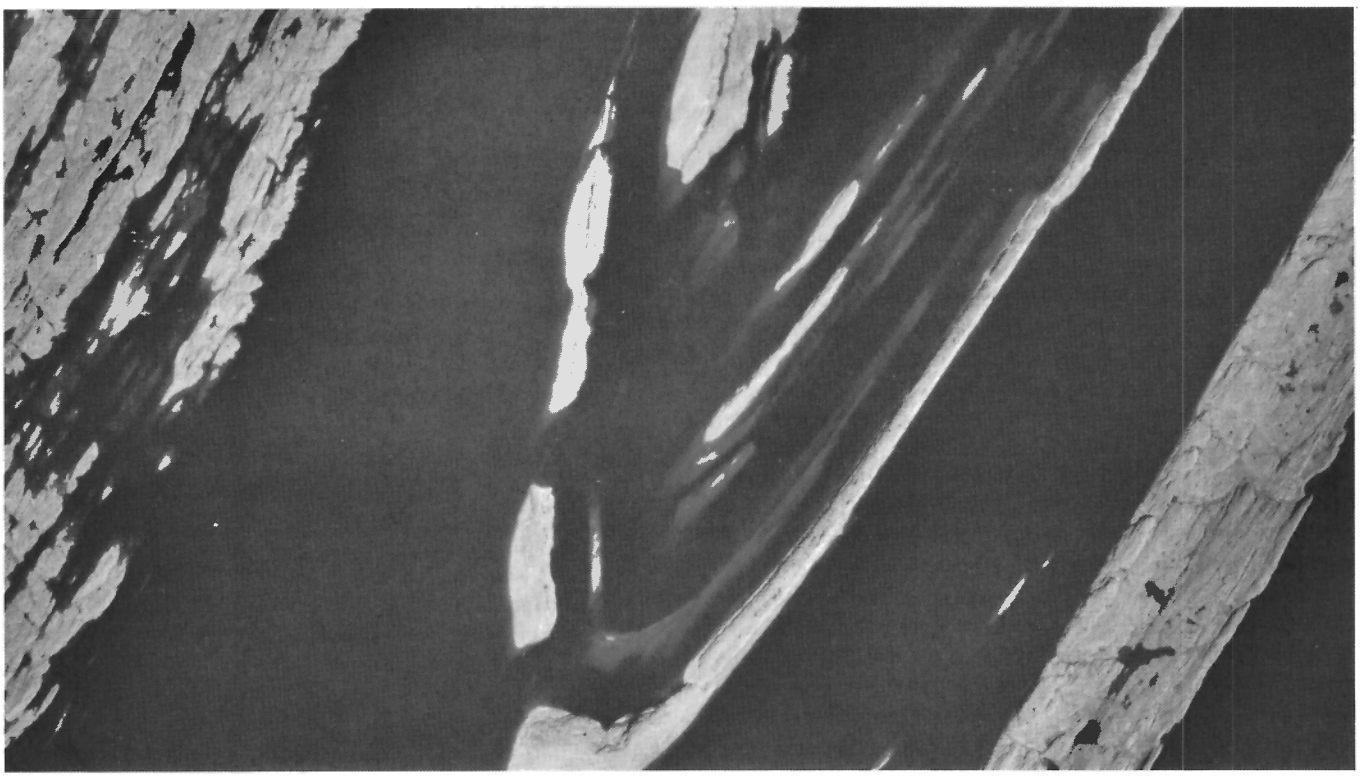


PLATE IV-16. Gently south-plunging anticline in Aphebian Belcher Group, Belcher Islands, Northwest Territories.

Tectonic Summary

Southern Province, south of Superior Province, is divided into three subprovinces: Penokean Fold Belt, Port Arthur Homocline, and Lake Superior Basin. Southern Province comprises Aphebian or younger sedimentary and volcanic rocks that lie unconformably on Archean rocks of Superior Province. The Aphebian rocks have been intruded by basic rocks of several ages and in the Penokean Fold Belt they have been folded, locally metamorphosed, and intruded by small granitic batholiths and then affected by the later Hudsonian or Penokean Orogeny. Three separate sequences of Aphebian rocks are found in Southern Province. In the Port Arthur area the Animikie Group comprises a conformable sequence of chert, jasper, iron-formation, carbonate, interbedded carbonate and chert, argillite, and tuffs. These rocks were deposited in a shallow-water, near-shore environment during a time of relative tectonic stability. Some chert and iron-formation may have been partly deposited in rather restricted near-shore basins and the argillite and slate in somewhat deeper water below wave base. These rocks are very little deformed and have a shallow southerly dip.

Two sequences of Aphebian rocks occur in the Penokean Fold Belt, the most extensive of which is the Huronian Supergroup consisting of four groups. The oldest, the Elliot Lake Group, comprises oligomictic conglomerate, argillite, siltstone, and greywacke or subarkose. The younger Hough Lake, Quirke Lake, and Cobalt Groups each represents a similar cycle of sedimentation that produced a basal assemblage of paraconglomerates, conglomeratic greywacke or subarkose succeeded by argillite, siltstone with some greywacke or subarkose and terminated with the deposition of subarkose and quartzite. The geosyncline in which these rocks were deposited was oriented in an easterly direction and the source of the sediments was to the north. The paraconglomerates are probably tillites and indicate that glaciation took place in early Aphebian times. The clastic sediments reflect rapid sedimentation in a fluvial environment from tectonically unstable areas. Sedimentation appears to have been continuous in the geosyncline but local disconformities occur. The presence of sediments that are probably Huronian equivalents in the Lake Mistassini area of Superior Province and in Grenville Province indicates that the geosyncline was originally very extensive. Sediments of the Whitewater Group are found only within the Sudbury Basin and include a thick sequence of tuffs, tuff-breccias, and bedded tuffs followed by argillite and siltstone, and then arkose. The Whitewater rocks are younger than Huronian strata and separated from them by an unconformity.

Correlation of the three Aphebian sequences in Southern Province with each other and with other Aphebian sedimentary sequences in the Canadian Shield is not well established. The glacial environment that resulted in some of the Huronian sediments must have been widespread during Huronian time. As evidence of glacial conditions is found only in the Huronian sequence, then the other Aphebian sediments in the Shield must be of a different age but whether they are older or younger is not known. Whitewater sediments are younger than Huronian strata and are commonly tentatively correlated with rocks of the Animikie Group on the basis of certain common features such as presence of carbonaceous sediments, tuffaceous rocks, and algae structures, but the evidence is not conclusive. Animikie rocks are also correlated with the Kaniapiskau or Belcher sequences chiefly on the basis of the occurrence of iron-formations. It is possible that the Animikie, Whitewater, Mistassini, and Kaniapiskau sediments were all deposited in a continuous geosyncline and that they are younger than Huronian rocks.

In Canada, Animikie rocks are very gently deformed but their equivalents to the south and west in Minnesota have been folded and mildly metamorphosed. Goldich, *et al.* (1961) define this event as the Penokean Orogeny and the K–Ar ages of metamorphic minerals presumably formed during this orogeny average about 1,700 m.y. In the Iron Mountain district of Michigan the major metamorphic event affecting Animikie rocks or their equivalents took place at about 1,900 m.y. Stockwell (1964) points out that the range and mean of K–Ar ages in Southern Province is about the same as for the Hudsonian Orogeny in Churchill Province, and suggests that the orogeny in Southern Province be referred to as Hudsonian. Whitewater rocks in the Sudbury Basin are older than the Sudbury Irruptive dated at 1,720 m.y., but the age of folding of the sediments depends on whether the irruptive was emplaced along the base of deformed or undeformed Whitewater rocks. Rocks of the Huronian Supergroup were folded about easterly trending axes, then intruded by gabbro sheets that date about 2,150 m.y. and then intruded by several masses of granitic rocks and locally metamorphosed. Ages of post-Huronian granites near Sudbury appear to be over 2,000 m.y. The oldest K–Ar age from the Cutler granite on the north shore of Lake Huron is 1,750 m.y. but Fairbairn, *et al.* (1965) suggest that this granite may be older because a sample of it along with samples of other granites in the region define a Rb–Sr isochron of over 2,000 m.y. K–Ar ages of metamorphic minerals range between 1,500 and 1,900 m.y. It is possible, therefore, that Huronian rocks were affected by a series of orogenic events, the earliest of which was folding followed by metamorphism and intrusion of small granitic masses possibly of differing ages but probably over 2,000 m.y. old. The range in radiometric ages may be due to intermittent movements along the major faults that cut Huronian rocks or the intrusion of the Sudbury Irruptive, or, near the Grenville Front, to the effects of

the Grenvillian Orogeny. It is possible, however, that Huronian rocks were affected by an early orogeny at about 2,000 m.y. and that Animikie rocks were involved in a later orogeny, the Penokean or Hudsonian, at about 1,700 to 1,900 m.y., that also affected Huronian rocks to some extent.

Several periods of basic magma intrusion took place in Southern Province. The Nipissing and Sudbury gabbros occur in Huronian rocks as sheets, sills, and dykes. These rocks yield radiometric ages of about 2,150 m.y. The Sudbury Irruptive is dated at about 1,720 m.y. and is of major economic importance. The Logan sills intrude Animikie rocks, probably are about 1,000 m.y. old, and may be related to the Duluth gabbro and to Keweenawan flows. On the margins of Lake Superior sediments and flows of the Keweenawan Group rest unconformably on the Aphebian and Archean rocks. The sediments consist of sandstones, conglomerates, limestone, and shale overlain disconformably by thin conglomerate, sandstone, and a great thickness of basaltic flows. The flows have yielded K–Ar dates that range from 900 to 1,100 m.y. indicating a Neohelikian or possibly Hadrynian age.

Huronian Supergroup

The name Huronian has been applied to a sequence of sedimentary and volcanic rocks of Aphebian age that lies unconformably on the Archean which is deeply eroded and has a rugged topography. The Huronian occurs in a belt at least 270 miles long and as much as 80 miles wide that extends from Lake Superior along the north shore of Lake Huron from Sault Ste. Marie to Sudbury and then north to Kirkland Lake and Noranda (Fig. IV-20). Remnants of equivalent rocks probably occur in the Chibougamau and Lake Mistassini regions previously described under Superior Province. The Huronian rocks were first systematically studied by Collins (1925) who defined the stratigraphic nomenclature that was accepted for many years and made regional correlations, some of which have been disputed. Difficulties were created because much subsequent geological work was concentrated in the vicinity of Sudbury and near Blind River after the discovery of uranium in that area. Collins' work was modified in each locality but lack of information from the intervening area made correlations difficult. Recent work by the Ontario Department of Mines between Blind River and Sudbury has provided new data for a stratigraphic nomenclature that can be applied with some confidence throughout the Huronian sequence. Problems have also been created by the usage of the term Huronian as it has been applied as a rock-stratigraphic, a time-stratigraphic, and a time term. It is used herein as a rock-stratigraphic term with the rank of a supergroup (Table IV-4), divided into groups and formations following Roscoe.

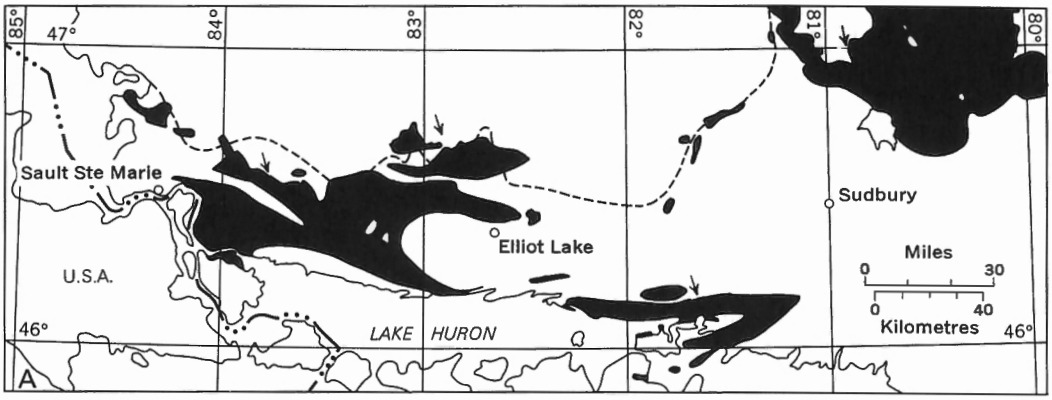
The Elliot Lake Group is the oldest in the Huronian Supergroup and is composed of two formations, the Matinenda and McKim. The group was originally defined in the Elliot Lake area and the upper formation was named the Nordic but this unit has since been correlated with the McKim in the Sudbury area and abandoned. The thickness of the group varies from about 100 feet to at least 13,000 feet. The basal formation, the Matinenda, is composed of subarkose or feldspathic quartzite, quartzite, and oligomictic and polymictic conglomerate. In the Elliot Lake area, the formation varies in thickness from a few feet to roughly 800 feet, which is probably the maximum. The thicker sections are in topographic lows in the basement that trend southeast and may correspond to valleys eroded in the Archean volcanic and sedimentary rocks. The subarkose is greenish yellow, thin to coarse bedded, crossbedded and coarse grained. It is composed of immature, moderately well sorted, subangular to subrounded grains of quartz and feldspar (mostly potassic) with minor amounts of sericite and chlorite. The oligomictic conglomerates, some of which are uraniferous and pyritic occur in beds that range in thickness from an inch to 10 feet. The conglomerates are thicker, more persistent and abundant in the lower parts of the thickest sections. Well-rounded quartz pebbles form 95 per cent of the clasts; chert is the next most abundant rock type and minor amounts of acidic volcanic and feldspar pebbles also are present. The clasts range in size from pebble to cobble and the size decreases down the paleoslope in the Elliot Lake area (Pienaar, 1963). The matrix is composed of subangular quartz and potash feldspar sand grains with some white mica and pyrite. In the ore horizons, pyrite, brannerite, uraninite, and monazite occur as subrounded grains in the matrix. The upper part of the Matinenda Formation is composed of subarkose with lenses or beds of polymictic conglomerate that grade upwards to quartzite. Pienaar reports local disconformities between the upper and lower parts of the formation and the upper units overlap the basement. The polymictic conglomerates occur at the base and are composed of rounded, tightly packed pebbles or boulders of quartz, metavolcanic rocks, chert, and minor granite in a subarkosic matrix. These rocks are not uraniferous.

The McKim Formation of the Elliot Lake Group comprises interbedded argillite, siltstone, greywacke or subgreywacke, and feldspathic quartzite. In the Elliot Lake area the formation is as much as 300 feet thick but it appears to thicken to the south and east in the Sudbury area to the point where it may attain a thickness of 10,000 feet. The argillites are thinly bedded, often laminated, and commonly greenish or black. The greywacke and feldspathic quartzite are thinly bedded and often interbedded with argillite. Both graded bedding and crossbedding occur. Volcanic rocks occur near the base of the Matinenda Formation or underlie it. They are considered part of the Elliot Lake Group but have been assigned various names: the Duncan Formation near Sault Ste. Marie, the Thessalon in the Bruce Mines area, the Pater in the Espanola area, and the Stobie, Copper Cliff, Frood,

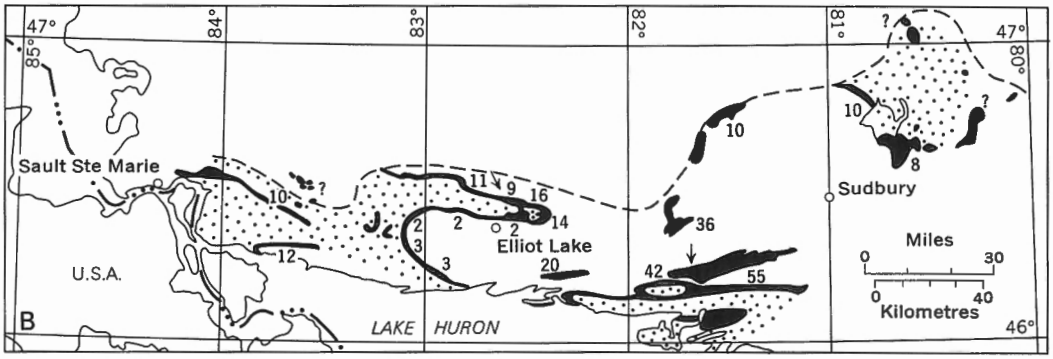
TABLE IV-4 *Comparison of stratigraphic nomenclature of Huronian rocks, Southern Province*

	Collins (1925) North Shore of Lake Huron and Sudbury		Coleman (1914) Sudbury		Cooke (1946) Sudbury		Thomson (1962) Lake Temagami, Sudbury, Blind River		Roscoe (1960) Quirke Lake, Elliot Lake		Robertson (1963) Elliot Lake, Blind River		Roscoe (personal communication, 1967) Sault Ste. Marie, Blind River, Sudbury, Lake Temagami	
	Series	Formation	Series	Formation	Series	Formation	Group	Formation	Group	Formation	Group	Formation	Group	Formation
PROTEROZOIC			Whitewater		Whitewater	Chelmsford Onwatin Onaping							Whitewater	
HURONIAN	Cobalt	Upper white quartzite Banded chert quartzite Lorrain Gowganda			Cobalt	Gowganda	Cobalt	Lorrain Gowganda	Cobalt	Lorrain Gowganda	Cobalt	Lorrain Gowganda	Cobalt	Bar River Gordon Lake Lorrain Gowganda
	Bruce	Serpent Espanola Bruce Mississagi		Ramsey Lake	Bruce	Serpent Espanola Bruce Mississagi Ramsey Lake McKim Copper Cliff	Bruce	Serpent Espanola Bruce Mississagi	Quirke	Serpent Espanola Bruce	Bruce	Serpent Espanola Bruce Upper Mississagi Middle Mississagi Lower Mississagi	Quirke Lake	Serpent Espanola Bruce
									Hough	Mississagi Pecors Whiskey			Hough Lake	Mississagi Pecors Ramsey Lake
									Elliot	Nordic Matinenda			Elliot Lake	McKim Matinenda[1]
	Unconformity		Unconformity			Fault	Unconformity		Unconformity		Unconformity		Unconformity	
ARCHEAN	Sudbury	McKim Copper Cliff	Sudbury	Wanapitei McKim Copper Cliff	Stobie Hill Coneston		Whitewater	Onwatin–Chelmsford Onaping						
							Copper Cliff							
							Sudbury	Scadding Falcon Wanapitei Ramsey Lake McKim						
	Keewatin	Schist complex	Keewatin				Keewatin Type	Stobie and other volcanic-sedimentary rocks	Pre-Huronian Basement		Keewatin		Pre-Huronian Basement	

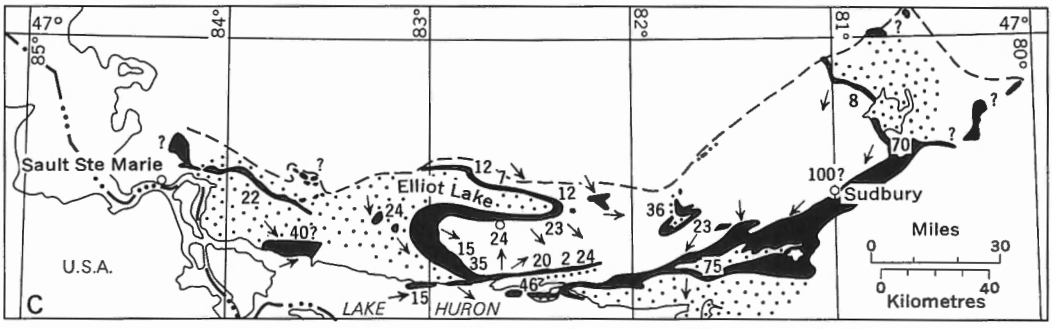
[1] Within or below the Matinenda are volcanic rocks named Thesalon in the Sault Ste. Marie area, Pater in the Blind River area and Stobie and Copper Cliff in the Sudbury area.



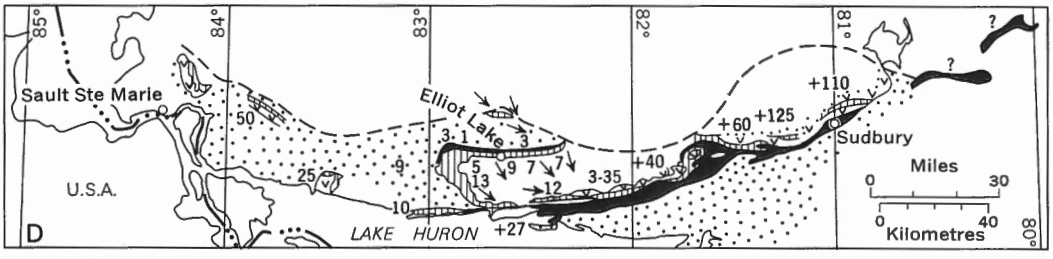
Cobalt Group: Gowganda, Lorrain, Gordon Lake, and Bar River Formations



Quirke Lake Group: Bruce, Espanola, and Serpent Formations



Hough Lake Group: Ramsay Lake, Pecors, and Mississagi Formations



Elliot Lake Group: Livingstone Creek, Thessalon, Pater, Stobie, Matinenda, and McKim Formations

Outcrop areas of groups in A, B, C, and McKim Formation in D

Matinenda Formation

Volcanic rocks of Livingstone Creek, Pater, and Stobie Formations

Subsurface extent of groups

Paleolimits of groups . . . . . ------

Paleolimit of Gowganda Formation . . . . . ----

Predominant direction of primary inclination of crossbeds in major arenaceous units . . . . . →

Estimated thickness of groups in hundreds of feet . . . 12

GSC

FIGURE IV-20
Distribution of the Huronian Supergroup, Southern Province, Ontario (by S. M. Roscoe).

and Elsie Mountain in the Sudbury area. In the Bruce Mines area the volcanic rocks occur within the Livingstone Creek Formation (Frarey, 1967) which is equivalent to the Matinenda. Near Sudbury the volcanic rocks conformably underlie and are intercalated with the Matinenda. They conformably underlie the McKim and are considered to be Huronian strata (S. M. Roscoe). The rocks are basic to intermediate and acidic lavas with interbedded basic and acidic pyroclastics.

The Hough Lake Group of the Huronian Supergroup includes the Ramsay Lake, Pecors, and Mississagi Formations. The basal formation in the Elliot Lake area was originally named Whiskey by Roscoe (1957) but it is now correlated with the Ramsay Lake conglomerates of the Sudbury area. The Ramsay Lake Formation consists of a paraconglomerate with some interbedded greywacke and subarkose. The thickness varies from a few inches to possibly 400 feet. The thicker sections include a greater proportion of conglomeratic subarkose. The clasts in the paraconglomerate are subrounded, vary from pebbles to large boulders several feet in diameter, and consist of granitic rocks, quartz, chert, volcanic rocks, and rarely quartzite. Relative amounts of each lithology vary from place to place as does relative proportion of clast to matrix. Generally there is an excess of matrix relative to clasts so that the clasts are loosely dispersed in a disrupted framework, the paraconglomerates commonly grading to conglomeratic greywacke. The matrix is greywacke or subarkose and siltstone composed of subangular grains of quartz, plagioclase, and lithic fragments set in fine white mica and chlorite. The Pecors Formation comprises interbedded argillite, siltstone, greywacke, and subarkose and varies in thickness from about 100 to 2,000 feet. The formation appears to be thickest in the area between Blind River and Sudbury. At Sudbury, rocks of this formation have been included with either the Ramsay Lake or Mississagi Formation. The lower part of the formation is composed of argillite and siltstone and the upper part is thinly to thickly bedded greywacke or feldspathic quartzites; the latter is more abundant in the Espanola region. The ratio of sand grains to matrix varies but the regional pattern is not established. Crossbedding, graded bedding, ripple-marks, and mudcracks have been observed, and the argillites commonly are finely laminated. The Mississagi Formation is a relatively thick sequence of coarse- to medium-grained subarkose and subgreywacke. Thin interbeds of siltstone or shale are common. The rocks range in thickness from about 600 feet to possibly 10,000 feet, the thickest sections occurring in the south and east. The quartzites are composed of fairly well sorted subrounded to subangular grains of quartz, microcline, and plagioclase in a sericite matrix. Rocks in the lower part of the sequence are generally more feldspathic than those in the upper. The ratio of clasts to matrix varies throughout the formation.

The Quirke Lake Group is composed of three formations: the Bruce, Espanola, and Serpent. Its thickness ranges from 200 feet in the Elliot Lake area to about 5,500 feet south of Espanola. The lowest formation, the Bruce, is a paraconglomerate that varies in thickness from 20 feet to more than a thousand. It is composed of a mixed assemblage of subrounded to subangular boulders, cobbles, and pebbles that range from less than an inch to 3 or 4 feet in diameter. The clasts tend to be more abundant and larger in the lower part of the unit. They are composed of grey and pink granitic rocks, quartz, metavolcanic rocks, and rarely metasediments. The proportion of clasts in the matrix varies but is usually less than 20 per cent and they are dispersed in a greywacke or subarkose matrix composed of sand and silt-sized grains of quartz, microcline, plagioclase, and lithic fragments in a chlorite–sericite matrix. Some parts of the unit are conglomeratic greywacke or subarkoses. This heterogeneous assemblage rarely shows evidence of bedding or stratification. However, in a few localities the exposed long axes of clasts show a preferred orientation north-northwest to north-northeast (S.M. Casshyap). Contacts with underlying and overlying formations are generally sharp or gradational over a few feet. The Espanola Formation is composed of limestone, siltstone, argillite, and calcareous sandy units with some breccias and intraformational conglomerates. It is generally divisible into three members comprising a lower thinly bedded limestone with thin silt interbeds, a calcareous greywacke or siltstone and argillite sequence with silty limestone interbeds and sedimentary breccias, and an upper unit of interbedded silty limestone, calcareous siltstone, subarkose and thin dolomite. The three members are not everywhere present. The thickness of the unit varies from about 400 to 2,000 feet. The relative amounts of limestone or limy beds vary throughout the sequence as do the relative thicknesses of the three members. The Serpent Formation, the youngest unit of the Quirke Lake Group is composed of subarkose with some argillite and conglomerate. It has a characteristic pure white porcellaneous appearance and is well bedded, crossbedded, fine grained, and fairly well sorted. The subarkose is composed of subrounded grains of quartz, plagioclase, and microcline in a matrix of white mica.

The Cobalt Group, the uppermost group of the Huronian Supergroup, is composed of four formations of which the lowest two are most extensively exposed and best known. The basal Gowganda Formation consists of paraconglomerate, argillite, siltstone, subarkose, and greywacke. In Southern Province it generally overlies the Serpent Formation conformably or locally disconformably but, in Superior Province, it overlaps the other parts of the Huronian and lies unconformably on Archean rocks. Exposed thicknesses range from less than 1,000 to 10,000 feet but the true thickness of the formation is generally unknown as it is often the uppermost unit exposed. In the Lake Timagami area it is divided into series of units comprising a basal fluvial granite conglomerate overlain by the other members of alternating units of paraconglomerate or conglomeratic greywacke and thinly bedded

argillite and quartzite (Schenk, 1965). In the Cobalt area the Gowganda is divided into a lower paraconglomerate and greywacke and upper bedded greywacke or siltstone–argillite (Thomson, 1957). South of Espanola, S.M. Casshyap has established four members that in ascending order comprise a paraconglomerate or conglomeratic greywacke, a laminated or massive argillite and siltstone unit, conglomeratic greywacke, and a mixed unit containing subarkose, conglomerate, and argillaceous siltstone. There seems to be a decreasing amount of conglomerate from the bottom to top of the sequence. The paraconglomerates are composed of very poorly sorted subangular to subrounded clasts of granitoid rocks with less abundant amphibolite, siltstone, quartzite, and metavolcanic rocks dispersed in a generally massive matrix of subarkose or greywacke. Striated pebbles are reported from a number of localities. The clasts rarely exceed 20 per cent by volume of the rock. They range in size from very small pebbles to large boulders. The matrix is composed of poorly sorted angular grains of quartz, feldspar, and rock fragments in a sericite–chlorite matrix. In a few localities exposed long axes of clasts have a preferred orientation north-northwest to north-northeast (Pettijohn, 1962), which indicates the line of movement but not its sense. The interbedded argillites, siltstones, or subarkoses are commonly conglomeratic, containing scattered clasts of lithologies similar to those in the conglomerates that range in size from fractions of inches to several feet. Sedimentary structures such as graded bedding, current ripples, load casts, convolute bedding, and clastic dykes (Pl. IV-17) are common in these rocks. Locally the argillite–siltstones are thinly bedded or laminated and have the appearance of varved sediments (Pl. IV-18). These units contain pebbles or boulders that puncture and disrupt bedding but are overlain by undisturbed beds. Such clasts are thought to result from ice-rafting.

The Lorrain Formation overlies the Gowganda conformably and gradationally. It ranges in thickness from 6,500 feet in the Bruce Mines area to 20,000 feet in the Lake Panache area. In Superior Province, the Lorrain



PLATE IV-17
Sandstone dyke with central zone of pebbles in Espanola siltstone, Aphebian Quirke Lake Group, Southern Province, Espanola, Ontario.

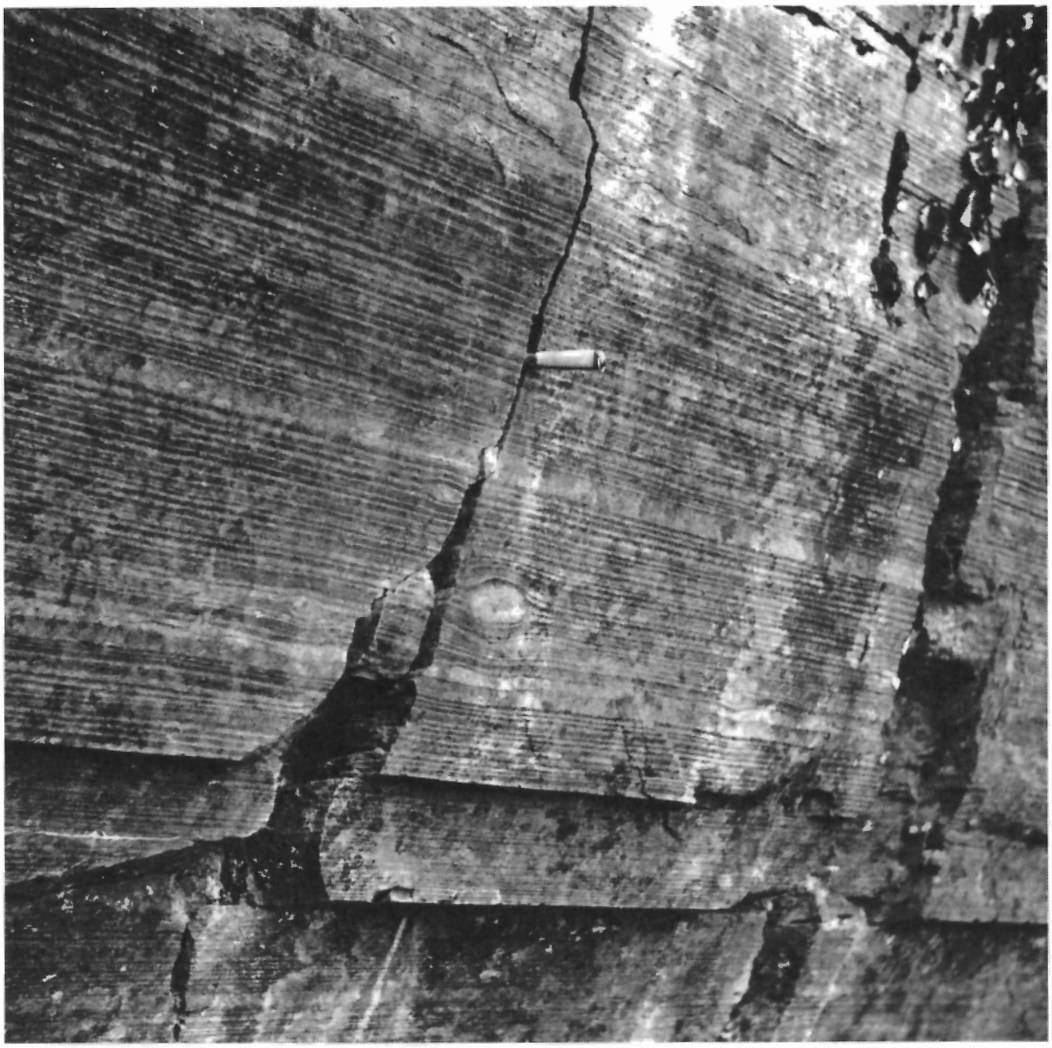


PLATE IV-18

Rafted pebble in Gowganda varved sediments, Aphebian Cobalt Group, Espanola, Ontario.

is the uppermost exposed formation of this group and only a maximum thickness of 1,000 feet is known. The formation is composed of a number of lithologies, some of which appear to be local whereas others persist throughout the basin. In the Bruce Mines area the Lorrain is divided into five units that from the base upwards include pink arkose or subarkose with minor siltstone and argillite, purple siltstone and quartzite, red to pink well-crossbedded quartzite with numerous thin beds or lenses of quartz-pebble conglomerate, jasper conglomerate and pale green feldspathic quartzite, and white quartzite with interbeds of pebble-conglomerate. The red quartzite and jasper conglomerate are apparently restricted to the western part of the geosyncline, to the Bruce Mines and Sault-Ste-Marie areas. Farther east in the area south and southwest of Sudbury where thick sections are preserved the Lorrain Formation comprises a basal pink, red or grey subarkose succeeded by coarse-grained greenish subarkose whose colour is caused by a greenish sericite that forms the matrix to the quartz and feldspar sand grains. The youngest unit is pink to white feldspathic quartzite and quartzite. The quartzite is much more mature than the older arenites in the sequence. Thin lenses of quartz-pebble conglomerate occur throughout the units of subarkose and crossbedding is common in all lithologies.

The uppermost two formations referred to by Collins (1925) as the "banded cherty quartzite" and "upper white quartzite" have been named by Frarey (1967) the Gordon Lake and Bar River Formations, respectively. The Gordon Lake is composed of thinly bedded siliceous and argillaceous siltstones with some quartzite beds and very fine grained quartzite beds that are chert-like. The sequence varies in thickness from about 1,000 feet at Bruce Mines to 3,500 feet at Lake Panache. Its lower contact is conformable and gradational with the Lorrain. The conformably overlying Bar River Formation comprises white to cream-coloured, medium- or fine-grained quartzite with ripple-marks and crossbedding. In the Bruce Mines area it is about 1,000 feet thick but in the Lake Panache area it exceeds 4,000 feet, and includes laminated quartzite, grey argillite, multicoloured siltstones, and several beds of hematite-rich quartzite.

In a general way the thickness of most Huronian formations increases from north to south and from west to east. Paleocurrent determinations based on crossbedding orientations in the various clastic formations indicate that the prevailing direction of sediment transport was from the northwest, north-northwest or north, and, in the easternmost part of the geosyncline, also partly from the north-northeast. The paleocurrent trends as presently known appear to be consistent throughout the time of deposition of the Huronian rocks.

Certain sedimentary features are common to the three paraconglomerate formations, the Ramsay Lake, Bruce, and Gowganda, and suggest that their depositional and tectonic environments were similar. Most studies have been concentrated on the Gowganda, and the more recent, such as those by Pettijohn (1962) and Schenk (1965), have favoured a glacial origin. Pienaar (1963) has suggested that they were deposited by coalescing mudflows triggered by tectonic activity, but the weight of evidence favours a glacial origin. Their great thickness and remarkable lateral extent of individual paraconglomerate horizons, the variety of lithologies in the boulders, the size of boulders and the great distance that the boulders must have been transported, and orientation of the boulders all argue against a mudflow mechanism of transport. In mudflow deposits individual horizons generally have a more limited lateral extent; the clasts are usually angular and unorientated, of local derivation, and lack lithological variety. Structures in underlying sediments are commonly disrupted. Similarly, the large dimensions of boulders, the distances of transport from nearest possible source areas, the lack of stratification and vertical or horizontal facies changes suggest that these rocks are not turbidites. Furthermore, Cooke (1923) and Schenk (1965) describe striated glacial pavements under the Gowganda where it is in contact with Archean rocks. In the thinly bedded siltstones and argillites or the varve-like laminated argillites of the Gowganda, pebbles or boulders puncture and disrupt the bedding and are best interpreted as examples of ice-rafted clasts. The thickness and apparent great lateral extent of the paraconglomerates, the numerous lithologies of distant source represented in boulders, the presence of varved sequences, the absence of phenoclasts of the underlying sediments, the preferred orientation of the clasts in the conglomerates, and the rare occurrence of striated boulders are all lines of evidence that support a glacial origin for the paraconglomerates. If these rocks are in fact tillites, then during Huronian time there were four or more glacial advances. The associated fine-grained sediments within the conglomeratic units may represent interglacial or glacial-fluvial sediments deposited in shallow water as indicated by sedimentary structures such as crossbedding and current ripples. Also structures such as graded bedding and convolute bedding suggest that these sediments were locally reworked by slumping and related turbidity currents. The generally conformable and often gradational contacts with underlying sediments suggest that the tillites may have been deposited subaqueously.

The Huronian sediments record deposition of sediment, largely immature, from a northerly source area that was generally tectonically active and had a rugged topography. The source area was underlain largely by granitic rocks. Most of the sediments were deposited in a fluvial environment but some reflect a deeper water domain just below wave base. There is evidence for periodic transgressions and regressions of the shoreline. The geosyncline into which the sediments were deposited was oriented in an easterly direction and probably extended much farther to the east and west than present exposures indicate.

Deformation of the Huronian Supergroup

Huronian rocks were deformed by a number of separate tectonic events. The earliest deformation in Southern Province is expressed by a series of folds whose axes trend west-northwesterly to easterly to east-northeasterly (Fig. IV-21). This deformation is generally moderate but the intensity seems to increase south of the Murray fault. At Elliot Lake and to the west the folds plunge gently westward but in the east they plunge eastward. In that part of the Cobalt Group north of Sudbury that is in Superior Province, the rocks are undeformed or very gently folded about northerly striking axes. These folds plunge to the north at low angles. The temporal and tectonic relationship of these structures to those of Southern Province are unknown. In Southern Province small folds and cleavage suggest that the earlier folds were redeformed and locally cross-folding is evident. The age of the major folding is not precisely known but in the Blind River area, diabase sheets appear to be either folded with the sediments or intruded along folded strata. In the Espanola area, the gabbro sheets seem to be later than folding (Ginn, 1965) but synchronous with faulting. J. A. Robertson (1967) and Pienaar (1963) conclude that south of Blind River the sheets are later than major folding. The Sudbury gabbros appear to be post-folding. The evidence is somewhat ambiguous but generally supports a post-folding age of intrusion which means that major folding of Huronian rocks took place before 2,150 m.y., the age of Nipissing diabase as determined by Van Schmus (1965).

Huronian rocks are cut by at least three sets of faults. The most obvious set dips steeply south and strikes easterly, roughly parallel with the trend of fold axes. On many the movement is north side down and to the east. Some faults occur along or near the axial planes of the longer folds. Initial movement may have taken place during folding of the Huronian strata but movement continued intermittently over a long period of time as late diabase dykes were displaced by some of these faults. The longest fault of this set, the Murray, can be traced from south of Sudbury almost to the Bruce Mines area, and it has a vertical and horizontal displacement of about a mile (Thomson, 1962). The northeast-trending faults are vertical; they have right-

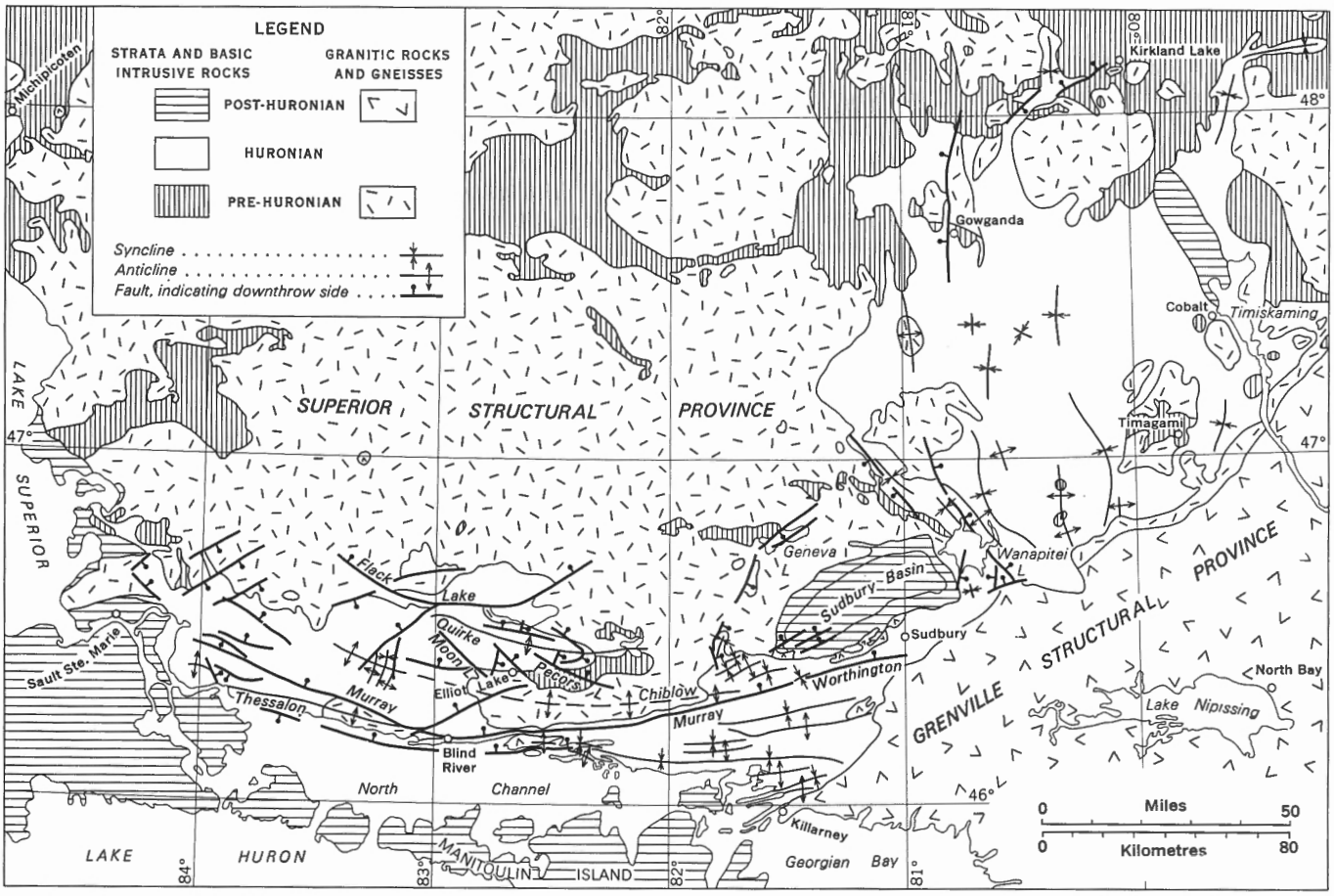


FIGURE IV-21. Major structures in the Huronian Supergroup, Southern Province, Ontario.

or left-hand displacement and either the northern or southern sides are downthrown. These faults cut the easterly striking faults and so are later. They displace the Sudbury Irruptive and some diabase intrusions. A third set of faults strike northwest and dip nearly vertically. Many with right-hand displacements are later than the easterly trending faults and displace the Nipissing diabase as well as younger diabase dykes. Locally, northerly striking vertical faults occur. Fault zones are distinguished by shearing and intense brecciation or hematite-stained quartz stockworks.

Metamorphism and Granitic Intrusions

For the most part Huronian rocks are only slightly metamorphosed to very low greenschist facies. However in certain relatively restricted areas assemblages of the epidote–amphibolite facies are found that include such minerals as staurolite or cordierite and garnet (Card, 1965a; J. A. Robertson, 1967). This higher grade metamorphism is later than the major folding in the region and is not entirely related spatially to intrusions of granitic rocks. Near Sudbury, Huronian rocks are metamorphosed near the Creighton, Cutler, and Murray granite masses. Along the boundary with Grenville Province Huronian rocks are metamorphosed to the amphibolite facies during the Grenvillian Orogeny.

The Cutler granite, along the north shore of Lake Huron, is a grey, medium-grained, equigranular aggregate of plagioclase quartz, potash feldspar, biotite, and muscovite. It may be massive or gneissic and commonly zones of cataclastic structures are developed by post-crystallization shearing. These structures are parallel with those related to early deformations. The Creighton granite is a massive or gneissic, commonly porphyritic, pink to reddish rock and is generally considered to be older than the Sudbury Irruptive. The Murray granite is composed of quartz, feldspar, and biotite and generally considered later than the irruptive but the evidence is contradictory. Remobilization of the granite during intrusion of the irruptive may have taken place. The Croker Island Complex north of Manitoulin Island cuts Huronian sediments. The pluton is circular and about 6 miles in diameter. It is composed of an outer zone of quartz monzonite, an interior ring of monzonite and syenodiorite, and a diorite–gabbro core (Card, 1965b). The order of intrusion of the various lithologies is from mafic to silicic types. Card suggests that the pluton was emplaced by forceful intrusion.

The age of the Croker Island Complex as determined by the Rb–Sr method is about 1,450 m.y. (Van Schmus,

1965). The Cutler granite has yielded a maximum radiometric age of 1,750 m.y. (Wetherell, *et al.*, 1960). The granites near Sudbury yield a range of ages. Early work using the Rb–Sr whole rock method indicated ages of about 2,000 m.y., but more recent work (Fairbairn, *et al.*, 1965) indicates ages of about 2,140 m.y. It seems reasonable to conclude that the Creighton and Murray granites are more than 2,000 m.y. old.

Gabbroic Rocks

The Nipissing diabase intrudes the Huronian sediments in numerous and extensive tabular masses as sills, slightly discordant sheets, or more steeply dipping dykes, many of which trend in an easterly direction. Thickness of individual intrusions varies from a few feet to 1,000 feet. In the thicker bodies differentiation has produced granophyre and diorite phases. The dark green rock is generally composed of labradorite, hornblende partly altered to chlorite, and lesser amounts of quartz, sphene, and iron oxides altered to leucoxene. In the Sudbury area, similar rocks are known as the Sudbury gabbro and they are probably time equivalents of the Nipissing diabase. These intrusions result from a major magmatic event, the Rb–Sr age of which is about 2,150 m.y. (Van Schmus, 1965). Some easterly trending diabase dykes in Huronian rocks yield K–Ar dates that average about 1,677 m.y. (Fahrig, *et al.*, 1965). Northwesterly striking diabase dykes occur throughout the region but are abundant in the Sudbury area and yield an age of about 1,220 m.y.

Sudbury Irruptive

The Sudbury Irruptive, an intrusion of great economic importance, is younger than the Huronian rocks and the overlying sediments of the Whitewater Group. The irruptive is composed of three zones: a lower norite and an upper micropegmatite or granophyre separated by a relatively narrow transition zone. The norite is a gabbroic rock composed of variable amounts of basic plagioclase commonly zoned and saussuritized, augite, hypersthene, and hornblende that may have formed from the original pyroxenes. Biotite and sulphides occur in small amounts along with variable amounts of interstitial micropegmatite or quartz and potash feldspar. The upper micropegmatite consists of a medium- to coarse-grained quartz, potash feldspar, plagioclase, and hornblende partly altered to chlorite.

The irruptive occurs as an elliptical ring enclosing an asymmetric basin, the long axis of which trends northeast. It is 37 miles long and 17 miles wide. The thickness of the intrusion is one to two miles and the micropegmatite phase is generally thicker than the norite. Breccia of fragments of irruptive and country rock commonly occurs as a narrow zone along the basal contact. The irruptive rocks intrude the Whitewater sediments, Archean granitic rocks and volcanic rocks of the Huronian Supergroup. Huronian rocks south of the irruptive face south, are folded about fold axes that trend easterly or slightly north of east, and are cut by granitic rocks and metamorphosed. The Whitewater sediments are gently folded about northeasterly trending axes. They are not metamorphosed or cut by granitic rocks. Regional structural relationships suggest that before intrusion of the irruptive, which has a Rb–Sr whole rock base of 1,720 m.y., the Whitewater rocks must have been separated from Archean and Huronian rocks by a profound unconformity.

The shape or form of the irruptive and its origin are still a matter of controversy. The weight of opinion probably favours the concept that the intrusion was emplaced horizontally along an unconformity between Whitewater sediments and older rocks, and after differentiation into its several phases it was warped or folded, together with the overlying sediments, to form the present basin structure, and then faulted. Recent paleomagnetic work suggests that the intrusion may have had an initial dip and been emplaced therefore along the deformed unconformity (Sopher, 1963). The nature of the irruptive body is not entirely known. Thomson (1956) however argues that the irruptive is a composite intrusion in a ring complex. H.D.B. Wilson (1956) considers that the norite and micropegmatite are the uppermost phases of a differentiated funnel-shaped intrusion, the lowest members of which are unexposed ultrabasic rocks. Dietz (1964) has proposed that the structure is a deformed astroblem and suggests that magma generated by force of impact of a giant meteor differentiated into the various phases of the irruptive.

Whitewater Group

Rocks of the Whitewater Group are restricted to the Sudbury Basin and are in contact only with the Sudbury Irruptive which intrudes them. The Whitewater Group is divided into three formations that are conformable and grade into each other. The lowest unit, the Onaping Formation, is composed of andesite tuff, tuff–breccia, lapilli tuff, minor bedded tuffs, rhyolite breccia, and, very locally, amygdaloidal and spherulitic andesite flows. The sequence has a maximum thickness between 4,000 and 5,000 feet. Thomson (1956) and Williams (1956) conclude that these rocks were formed from glowing avalanche debris. The Onaping grades upward into the Onwatin carbonaceous slates with minor limestone, cherty carbonate, and greywacke about 4,000 feet thick. The carbon content of the shales is characteristic. The Chelmsford Formation is composed of arkose with very thin interbeds of argillite. The arkose has angular to subangular grains of quartz and plagioclase. Crossbedding and ripple-marks are common and the rocks are generally thin bedded. Carbon also occurs in these sediments and limy, iron-rich ellipsoidal-shaped concretions are abundant in the sandstone. The Whitewater rocks, which are older than the Sudbury Irruptive, are Aphebian in age and probably younger than the Huronian rocks. They are tentatively

correlated with rocks of the Animikie Group of the Port Arthur area. The Whitewater strata are gently folded about axes that trend approximately northeast roughly parallel with the basin. The folding is stronger towards the southeast. Plunges of folds are gentle and are in both directions. The Whitewater and older rocks are cut by many faults, most of which strike northeasterly to east-northeasterly. The faults dip to the southeast at high angles and have considerable displacement. Steep-dipping faults that strike northeast, diagonally to the set above, may be related. The north rim of the basin is cut by a number of steeply dipping faults that strike northerly to north-northwesterly.

Port Arthur and Lake Superior Subprovinces

In the Port Arthur region the Archean rocks were eroded and a relatively flat featureless surface was produced preceding deposition of Aphebian sediments, the northernmost part of the extensive Animikie Group of Minnesota. In Ontario the group is composed of three conformable formations, the Kakabeka, Gunflint, and Rove. The Kakabeka Formation is a thin discontinuous conglomerate unit as much as 5 feet thick. It is generally included as the basal member of the succeeding Gunflint Formation. The conglomerate is composed of well-rounded pebbles of quartz, chert, jasper, and granite in a matrix of quartz sand. Pebbles are generally less than 2 inches in diameter. The Gunflint Formation includes a large number of lithologies that exhibit very rapid facies changes. Few members can be traced throughout the outcrop area. The formation has an average thickness of about 400 feet, and generally thickens from north to south. Many of the sedimentary units are iron-rich, and the formation is often referred to as the Gunflint iron formation.

In the Whitefish Lake area west of Port Arthur the Gunflint has been divided into two divisions (Goodwin, 1960). The lower division comprises a unit of algal chert, bedded chert and taconite, a soft black fissile pyritic argillite, and a unit of greenalite, magnetite and cherty taconite. The upper division is composed of algal chert, jasper, tuffaceous shale, taconite, and limestone. In Loon Lake area east of Port Arthur, a threefold division is evident comprising a lower unit of ferruginous carbonate, cherty taconite, and carbonate beds, followed by argillite with interbeds of carbonate and tuffs, and an upper unit of thin-bedded carbonate and chert and taconite. The uppermost limestone is only present locally. Individual units thicken or thin sharply and grade laterally into one another. The intermediate argillite east of Port Arthur is thought to be equivalent to the upper shale in the western section and appears to be the only unit that can be traced with confidence throughout the whole region. The taconites are composed of granules, about a millimetre in diameter, of chert, iron oxide, carbonate, and iron silicate in a matrix of chert and carbonate. The cherty carbonates are composed of thin, interbanded layers of chert and carbonate that contain minute spheroidal structures. The carbonates are commonly ferruginous dolomite or siderite.

The Rove Formation is composed of black, thinly bedded argillite with a few layers of red and green argillites, limestone, and greywacke. Certain zones contain large calcareous concretions as much as 8 feet in diameter. The formation attains a maximum thickness of 2,000 feet in Ontario and 3,000 feet in Minnesota. The formation thins as its northern margin is approached and this may be partly a characteristic of the original basis of deposition but is also partly due to erosion that preceded deposition of the overlying Keweenawan sedimentary rocks.

The Animikie Group is disconformably or unconformably overlain by Keweenawan rocks of the Sibley Formation that extends along the north shore and nearby islands of Lake Superior and inland to Lake Nipigon. The Sibley rocks rest on Rove shales and on several units of the Gunflint, and overlap the Archean rocks. The Sibley Formation is as much as 500 feet thick and composed of a discontinuous thin basal conglomerate that contains pebbles of underlying rocks in a quartz or carbonate sand matrix, succeeded by white to yellowish sandstone with thin interbeds of limestone and chert, and by carbonate and shale that in the south are deep red or maroon and iron-rich, but in the north are more typically white or light grey. The carbonates apparently are composed of carbonate grains with some quartz or feldspar grains in a calcite cement. Shale interbeds are common in the carbonate strata. The Sibley is disconformably overlain by about 50 feet of conglomerate, sandstone and mudstone of the basal Osler Formation. The upper part consists of thick basaltic lavas with interbeds of conglomerate or sandstone. Individual flows are as much as 200 feet thick. The thickness of the Osler is 6,000 to 10,000 feet. Similar rocks on the east shore of Lake Superior at Mamainse Point are estimated to be 15,000 to 17,000 feet thick (Thomson, 1953).

Intrusions into the Keweenawan and older rocks are gabbro sills or sheets and diabase dykes. In the Port Arthur area, the Logan sills intrude the Animikie strata at several horizons. They vary in thickness from a few feet to 200 feet, but in the Lake Nipigon area are as thick as 1,000 feet. Some sheets that cut the strata at low angles to the bedding probably were intruded after regional warping. The thicker sills are differentiated and contain a number of lithological phases, the uppermost being a granophyre. Diabase dykes occur in a northeasterly striking swarm in the Port Arthur area. Rocks of the Animikie and Keweenawan Groups are very gently folded. They dip to the south at a low angle of 5 to 10 degrees. Locally, tight folds are related to movements along faults or may be due to volcanic activity. The slight regional angular discordance beneath the Sibley Formation suggests a period of erosion and moderate

warping. All the rocks in the region have been cut by steep-dipping faults apparently with normal displacement that strike east to northeast and north to northeast. They are marked by narrow breccia zones and locally contain quartz veins.

In Minnesota the Animikie rocks have been folded and metamorphosed during the Penokean Orogeny. Minerals from metasediments yield K–Ar ages in the 1,650 to 1,800 m.y. range. The Animikie rocks are therefore Aphebian but are isolated from the Huronian rocks in the eastern part of Southern Province, with which they have little stratigraphic or lithological similarity. The Animikie rocks have many features in common with parts of the Kaniapiskau Supergroup or the Belcher Group. The Keweenawan rocks are similar to those in other parts of the Lake Superior Basin. K–Ar age determinations on whole rock samples of basic volcanic rocks, gabbro intrusions, and diabase dykes fall in the range 900 to 1,100 m.y. They are, therefore, Neohelikian or possibly Hadrynian.

NAIN PROVINCE

The Nain Province is defined chiefly on the basis of K–Ar age determinations. Known structural trends are generally north to northwest, subparallel with the structural grain of the adjacent Churchill Province to the west. Most of western Nain Province is characterized by K–Ar dates ranging from 1,300 to 1,500 m.y. and by granitic gneisses that are intruded by large masses of Elsonian anorthosite and mangerite. Eastern Nain Province is represented in the coastal region by gneisses giving K–Ar ages that have been referred to the Kenoran Orogeny and one or two dates possibly referable to the Hudsonian Orogeny. The boundary between the eastern and western parts has been drawn arbitrarily to include all the anorthosite–adamellite rocks in the western part. Knowledge of the geology of Nain Province is so limited that a tectonic synthesis is considered premature.

The most widespread and the largest intrusive masses in Nain Province are members of the anorthosite suite. K–Ar determinations on biotite have yielded dates averaging 1,400 m.y. In the Nain area (Wheeler, 1960) the anorthosite–adamellite association is sufficiently similar to the anorthosite–mangerite complexes of Grenville Province that it is difficult to doubt a common origin. The Kiglapait layered gabbroic intrusion is probably related to the same igneous episode as the Nain anorthosite and it may also be inferred that basaltic magma played a role in the origin of the anorthosite–adamellite series. The Michikamau intrusion (Emslie, 1965) has a fine- to medium-grained gabbro margin which is regarded as evidence that basaltic magma was an important component in the development of the intrusion. Late differentiates are not volumetrically abundant at Michikamau but some are similar to the Nain adamellites. The anorthosite–adamellite plutons of western Nain Province are believed to be essentially non-orogenic. This is indicated by the presence of well-preserved layered structures and unzoned



PLATE IV-19

Layered leuco-troctolite of the Michikamau anorthosite intrusion, Michikamau Lake, Labrador. Gravity stratified rhythmic layering of olivine-rich lower parts (dark, left side) grading into plagioclase-rich upper parts (light, right).

minerals of coarse grain-size, and the lack of metamorphic effects. Accordingly, the major orogenic activity in the region preceded their emplacement, and in all probability it was the Hudsonian. The widespread post-Hudsonian K–Ar ages found in western Nain Province may be due to extensive heating of the crust at the time of emplacement of the anorthosite–adamellite plutons.

Lake Michikamau area. Much of the area is underlain by foliated to gneissic quartzo-feldspathic rocks with large or small remnants of paragneiss. Mineral assemblages typical of the amphibolite facies are common west of Lake Michikamau, but, bordering a large north-south elongated complex of pyroxene-bearing syenitic and granitic rocks, the gneisses are in granulitic facies. The mineralogy of the syenitic rocks is similar to that of the Nain anorthosite–adamellite complex described later. Granulite facies rocks, although sporadic in distribution, are present over large areas east and northeast of Lake Michikamau. The gneissosity and foliation strike northwesterly except where deflected about the Michikamau intrusion.

The Michikamau anorthositic intrusion (Emslie, 1965) is about 800 square miles in area. Biotite gave an age of 1,400 m.y. Large volumes of leuco-troctolite, commonly layered (Pl. IV-19), together with gabbro, leuco-gabbro, and anorthosite make up the bulk of the intrusion, which has a chilled margin of fine-grained gabbro. Late differentiates are extremely rich in ferrous iron. An older group of paragneisses can be recognized at many places forming the wall-rocks of the Michikamau intrusion. These comprise mainly pelitic, semi-pelitic, and impure quartzitic gneisses; the biotites yield a metamorphic age of 1,520 m.y. Several large, medium- to coarse-grained, biotite- and pyroxene-bearing granitic masses occur throughout the region. Some may be younger than the Michikamau intrusion. Granite that cuts the anorthosite has an age of 1,360 m.y. Biotite from the large mass southeast of Lake Michikamau gave a similar age of 1,395 m.y. The Neohelikian is represented by remnants of flat-lying arkose, sandstone, and conglomerate, believed to be correlative with the Seal Lake Group to the east. The conglomerates commonly contain numerous pebbles of intermediate to silicic volcanic rocks, some with abundant feldspar phenocrysts. The only known source of these pebbles is the Croteau Lake Group unconformably underlying the Seal Lake Group.

Nain area. The oldest recognized rocks are a gneiss complex consisting partly of paragneiss. Granulite facies mineral assemblages are typical of much of the area north of Davis Inlet and amphibolite facies assemblages occur to the south. Marble bands are few and thin; quartzites are absent. Ultramafic layers and lenses occur in the gneisses. An Archean age is indicated by several K–Ar determinations on biotite and hornblende and ranges from 2,390 to 2,665 m.y. In the vicinity of Hopedale the migmatitic gneisses include several sill-like altered peridotite bodies. Large, northeast-striking, vertically dipping, uralitized diabase dykes are common. The Nain anorthosite and related bodies are intrusive into the gneiss complex. Composition varies from meta-gabbro to anorthosite and leucodiorite. Chatoyant plagioclase is widely distributed. The chief mafic minerals are monoclinic and orthorhombic pyroxenes, and olivine is locally common in the mafic facies. Medium to very coarse grained textures are typical. Cataclastic zones are spatially related to the margins of the intrusion (Wheeler, 1960). The adamellite intrusions vary in composition from granodiorite to granite and show both intrusive and transitional

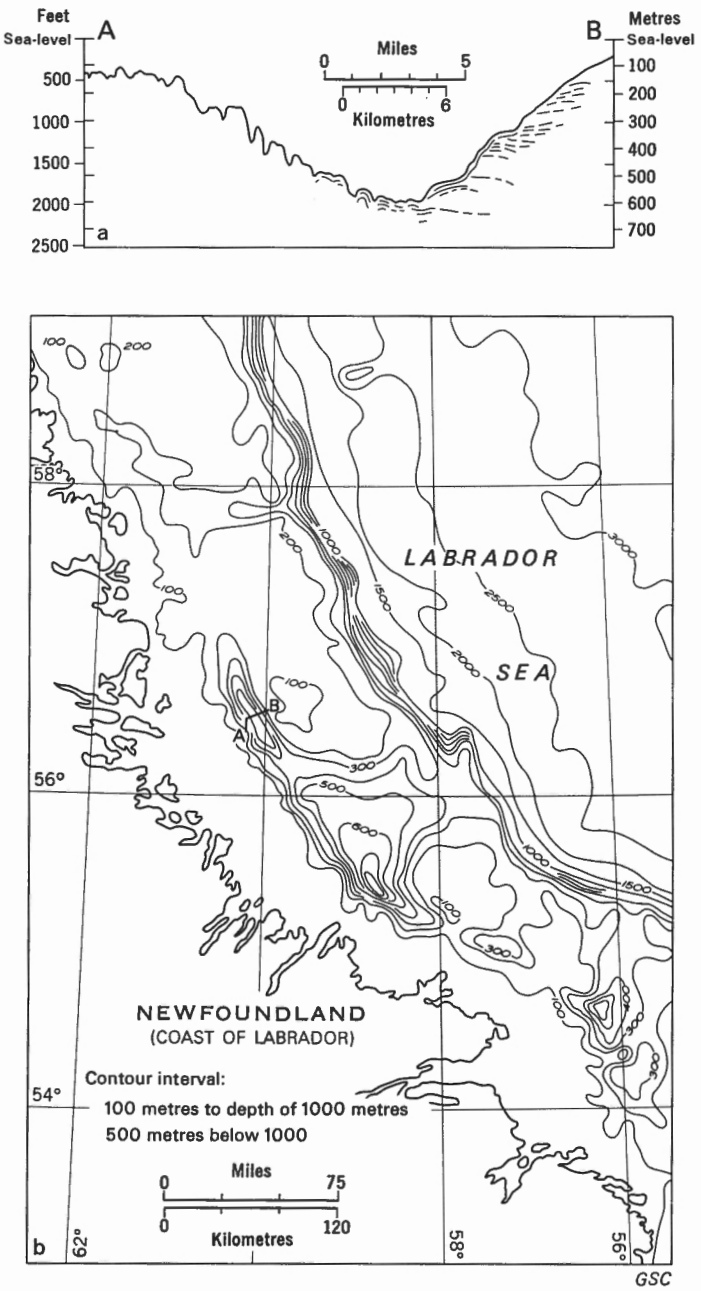


FIGURE IV-22. Labrador Shelf of the Atlantic Continental Shelf (Grant, 1966): (a) Seismic cross-section of the "marginal channel" (vertical exaggeration 15X); (b) Bathymetry.

relationships to the anorthositic rocks. Hornblende is the chief mafic mineral but biotite, orthopyroxene, clinopyroxene, and fayalitic olivine also occur. These rocks are similar in composition and associations to the mangerite intrusions in Grenville Province. The Kiglapait intrusion is a layered basaltic pluton that has extremely iron-enriched differentiates and is comparable in many respects to the Skaergaard Intrusion of East Greenland. Several K–Ar biotite dates from granitic rocks in the district (Beall, *et al.*, 1963) support an age of about 1,400 m.y. for the Nain anorthosite and the Kiglapait intrusion.

Northwest of Nain, the Siamarnekh Formation unconformably overlies adamellite. It consists of about 380 feet of flat-lying red subarkose similar to the Neohelikian Seal Lake Group.

Northern Labrador Coast. North of the Nain district granitic gneisses and granite outcrop along the coast. At the head of Hebron Fiord and northward to Saglek Fiord pyroxene-bearing gneisses are reported. Near the head of Nachvak Fiord, a northwest-striking zone of ultramylonite and pseudotachylite separates the hypersthene-bearing gneisses on the east from fine-grained gneiss carrying some marble to the west. Rocks of the Ramah Group are exposed in fiords from Saglek Fiord north to Nachvak Fiord. These consist of an estimated thickness of 4,000 to 6,000 feet of quartzite, argillite, slate, ferruginous quartzite, dolomite, and chert. As much as 240 feet of volcanic rocks are present near the base of the sequence. The rocks are gently dipping and rest unconformably on older gneisses; locally they are folded and faulted.

The Kaumajet Mountains are underlain chiefly by the Mugford Group of volcanic rocks comprising basalt, andesite, trachyte, and tuff. The volcanic rocks at some places unconformably overlie as much as 150 feet of quartzite, ferruginous quartzite, argillite, and dolomite that may correlate with the Ramah Group. Elsewhere the Mugford Group lies unconformably on the older gneisses that have yielded dates of 2,045 and 2,255 m.y. Parts of the group are slightly folded and block faulted but otherwise little disturbed. The Ramah Group could correlate with the Aphebian rocks of the Labrador Fold Belt, although there is a closer resemblance to the succession of the Cape Smith Fold Belt as the thick volcanic Mugford Group unconformably overlies a sedimentary succession with lean iron-formation. A single K–Ar date of 948 ± 90 m.y. suggests that the Mugford Group is much younger.

The bathymetric configuration of the Labrador Shelf (Fig. IV-22) indicates a narrow marginal channel, paralleling the coastline of northeastern Labrador for hundreds of miles, which could be a fault zone formed during Cenozoic uplift of the continent. From a seismic profile across the channel southeast of Nain, A. C. Grant (1966) concludes that the eastern flank of the channel is underlain by flat-lying layered rocks. These could be post-Precambrian in age or the relatively undisturbed Ramah and Mugford Groups.

GRENVILLE PROVINCE

Tectonic Summary

Highly metamorphosed Archean and Aphebian rocks have been recognized within Grenville Province because the Kenoran and Hudsonian folded belts are truncated by the Grenville Front and it is possible that rocks of the older orogenic belts may extend well into Grenville Province. Southwest of Chibougamau, Archean mafic, intermediate, and silicic lavas, pyroclastics, quartzo-feldspathic and slaty metasedimentary rocks in Grenville Province are metamorphosed into mica schist, biotite and hornblende paragneiss, and amphibolite. In the Wabush – Mount Wright district granitoid gneiss with granulite facies mineralogy similar to those in Superior Province are probably also Archean. Aphebian marble, quartzite, and iron-formation of the Labrador Fold Belt extend into central Grenville Province.

The Grenvillian orogenic cycle begins with the deposition of sediments of the Paleohelikian Grenville Group, accompanied at least locally by volcanism. These rocks are now recognized only in the southwest part of Grenville Province. Temporal equivalents of the Grenville Group may once have been widespread but subsequently removed by uplift and deep erosion. During or after Paleohelikian deposition and prior to significant deformation, large quantities of basic magma were intruded into the lower part of the sequence and underlying basement, resulting in formation of widespread anorthosite–mangerite plutons. Smaller intrusions of gabbro, diorite, and granitic and syenitic rocks may also have been emplaced at about this time. The culmination of regional metamorphism and the major deformation of the Grenvillian Orogeny followed. In some regions of high-grade metamorphism gneiss domes rose from the underlying granitic basement into the sedimentary sequence, perhaps accompanied by the emplacement of magmatic granites at higher levels. Structural trends in the Grenville Group are northeasterly as are those close to the Grenville Front. Elsewhere, complex fold patterns record deformation that predates the Grenvillian Orogeny. The present moderate to high grade of regional metamorphism throughout much of Grenville Province together with the characteristic complex structural style is evidence that strong uplift and deep erosion have exposed rocks whose environment of formation was the catazone. Following the period of intense orogenic activity some small granitic and syenitic plutons were emplaced.

Zones of cataclasis and mylonitization such as those along the Grenville Front are believed to have developed during the later stages of deformation and probably after the main metamorphic episode.

After the close of the Grenvillian Orogeny small-scale intrusive activity is represented by the Chatham–Grenville stock, alkaline complexes, and diabase dyke swarms. The Oka alkaline complex was intruded in the early Cretaceous.

Archean and Aphebian

Along the northwestern margin of Grenville Province highly metamorphosed Archean and Aphebian rocks have deformational histories that predate the Grenvillian Orogeny. Elsewhere in Grenville Province complex structural trends also suggest multiple deformation. Combination of these observations has led to speculation that the typically complex fold patterns are the result of superpositions of two or more orogenies and that rocks formed during orogenies older than the Grenvillian might be widespread. In the Surprise Lake area (Déland, 1956; Déland and Grenier, 1959) progressive metamorphism of Archean metasedimentary and metavolcanic rocks occurs at the Grenville Front btween Superior and Grenville Provinces (Fig. IV-23). The northwestern part, in Superior Province, is underlain by Archean rocks comprising pillowed mafic and intermediate lavas intruded by sills of meta-gabbro, silicic lavas, quartzo-feldspathic and slaty metasedimentary rocks, and small amounts of pyroclastics. Low greenschist facies mineral assemblages are present in these rocks. The sequence has been tightly folded about easterly trending axes. The southeastern part, in Grenville Province, is underlain by gneissic granite, biotite and hornblende paragneiss, hornblende gneiss, and amphibolite. Foliation and gneissosity strike northeasterly. The metasedimentary rocks are progressively metamorphosed into mica schists, biotite, and hornblende paragneiss. The mafic volcanic rocks are transformed into amphibolite (Fig. IV-24). The width of the zone of transformation averages two to three miles and mineral assem-

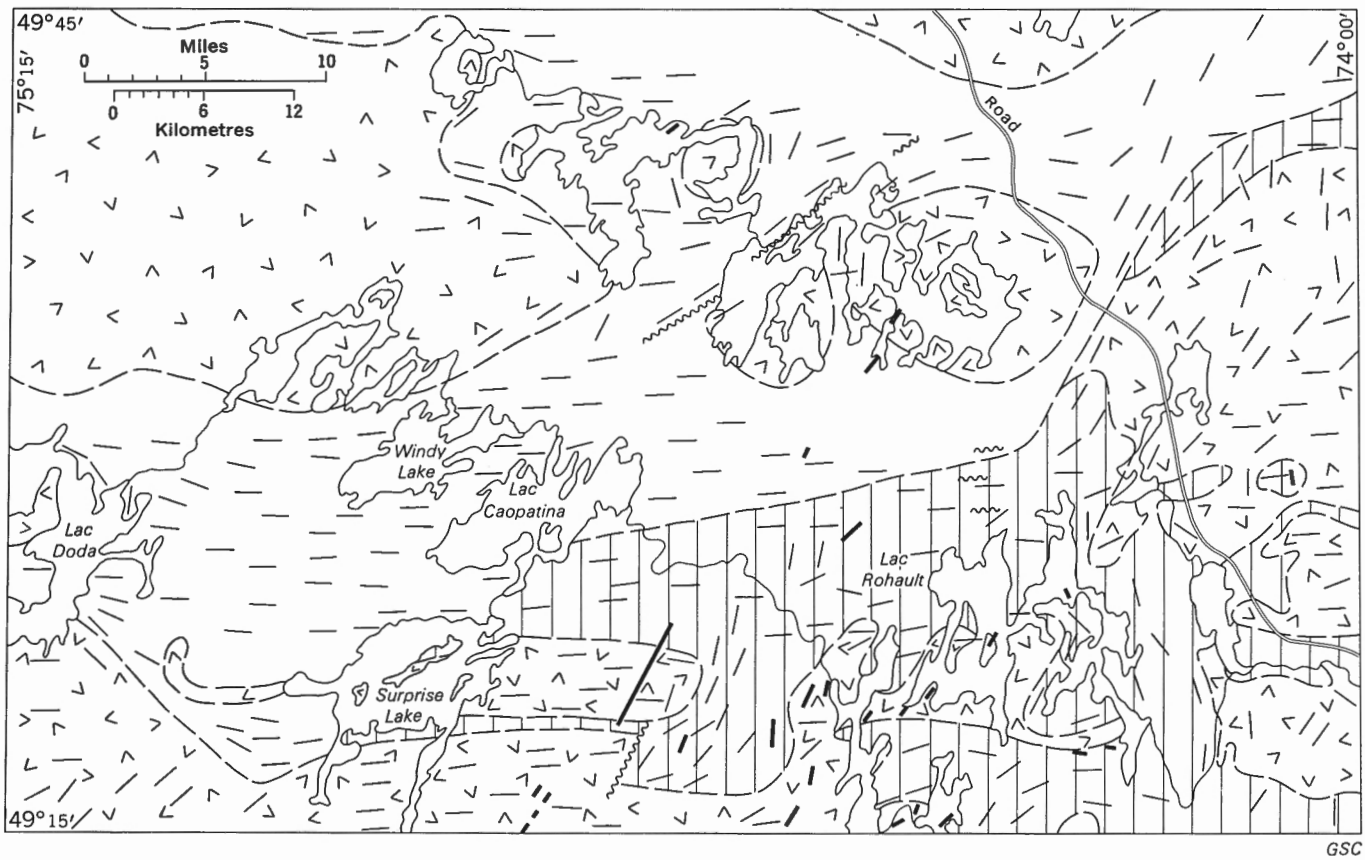


FIGURE IV-23. Geology of Suprise Lake region, Quebec (Déland and Grenier, 1959).

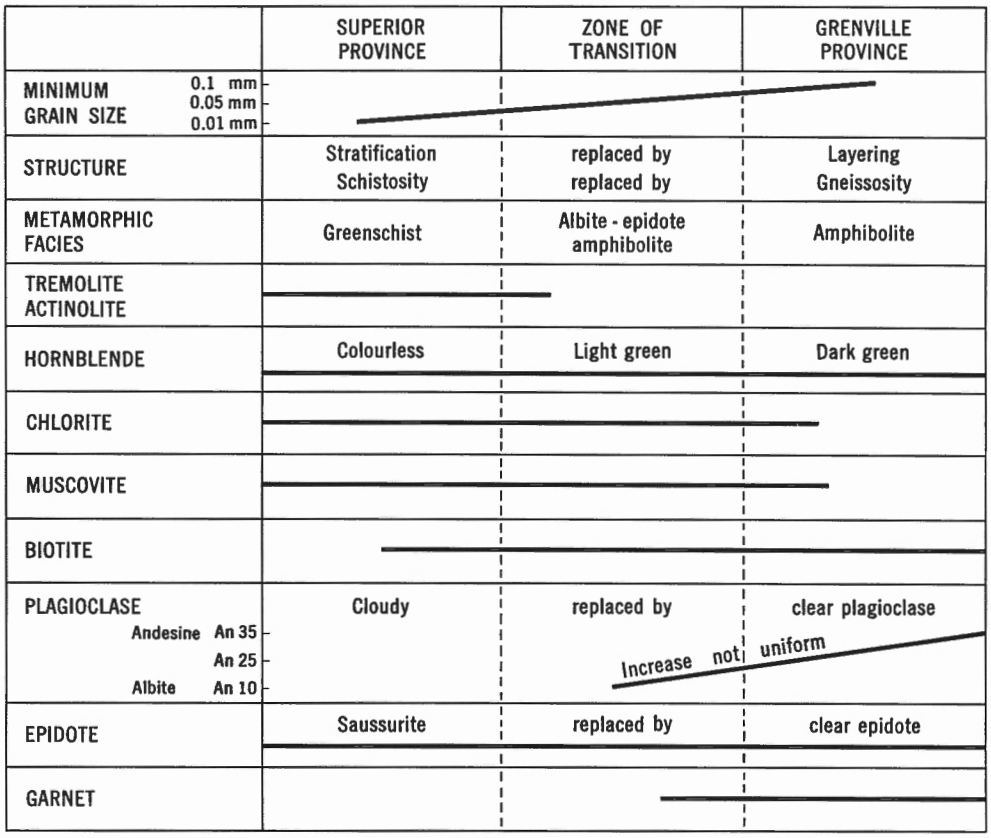


FIGURE IV-24
Mineralogical and textural changes in Archean rocks across Grenville Front, Suprise Lake area, Quebec (Déland, 1956).

blages of the upper greenschist facies are characteristic. Accompanying the increase in metamorphic grade, primary structures such as pillows and amygdules are almost completely obliterated in the metavolcanics and stratiform foliation replaces bedding in the metasediments. Passing beyond this zone to the east and south, biotite and hornblende paragneiss and amphibolite become coarser grained. The presence of deep green hornblende and oligoclase-andesine in the mafic rocks is characteristic of the almandine–amphibolite facies. There is some faulting but no major dislocations are apparent. The boundary between Superior and Grenville Provinces is chosen by Déland (1956) as the low side of the transition zone between the greenschist and almandine–amphibolite facies.

In the Wabush–Mount Wright district, the distinctive lithologies of iron-formation quartzite and carbonate of the Aphebian Kaniapiskau Supergroup have been recognized in Grenville Province (Duffell and Roach, 1959; Fahrig, 1960; Clarke, 1961). North-northwest striking unmetamorphosed rocks of the Labrador Fold Belt can be traced into high-grade schists and gneisses. There is also a rapid change in metamorphic grade and structural style at the Grenville Front. North of the Grenville Front (Fahrig, 1960; Wynne-Edwards, 1961) granulite facies gneiss forms a basement to the Kaniapiskau rocks, and the Sims Formation, largely quartzite, unconformably overlies it. The biotite isograd approximately indicates the southern limit of essentially unmetamorphosed rocks. Abundant evidence of faulting and cataclasis is found in the vicinity of the biotite isograd. Northwest and northeast directions of folding are recognized south of the Grenville Front whereas only northwest fold axes occur to the north. South of the front there are high-grade gneisses with mylonite zones and evidence of thrusting from the south. In the Archean granitoid gneisses of the granulite facies retrograde effects become increasingly apparent and almandine–amphibolite facies mineral assemblages predominate. Accompanying this metamorphic change is a structural change from gently dipping foliations and gently plunging lineations in the north to steep foliations and lineations in the south. Folding south of the Grenville Front is considerably more complex than to the north. At least two fold directions are recognized in the Aphebian rocks, northwest and northeast; the latter is generally considered to be younger. Gastil and Knowles (1960) interpret numerous structural discontinuities in the Wabush area as low angle thrusts. Folds, asymmetrical and overturned to the northwest also suggest thrusting from the southeast. In the Wapussakatoo Mountains (Gill, *et al.*, 1937) tight folds overturned to the northwest are attributed to thrust faulting. Quirke and Collins (1930) recognized remnants of Aphebian sediments of the Huronian Supergroup southeast of the Grenville Front in the Lake Huron–Lake Panache district of Ontario. They considered that vertical displacement of the Grenville Province side by as much as 6 miles was necessary to account for the abrupt disappearance of the thick Huronian formations.

Apart from the localities described above there are no other places in Grenville Province where Archean and Aphebian rocks have been demonstrated to exist. However, in recent years evidence bearing on the recognition of basement and cover rocks in Grenville Province has been discussed by numerous authors. Hewitt (1962) interprets granite gneiss domes in the Hastings–Haliburton Highlands of Ontario as autochthonous. He suggests that under catazonal conditions granitic rocks became mobile and rose as domes into the overlying paragneisses. North of Ottawa, Pollock (1962) has concluded that rocks beneath the Grenville Group have been remobilized and intruded into the Grenville Group. Pollock includes within the basement complex, rocks of the anorthosite–mangerite association which are now believed to have been intruded during the Elsonian Orogeny. In the Adirondack Highlands of New York State, Walton and DeWaard (1963) suggest that most of the igneous rocks including those of the anorthosite–mangerite form a basement to Grenville sediments. Their hypothesis of evolution of the region involves at least two major orogenic cycles.

Wynne-Edwards (1964) suggests that supracrustal rocks within Grenville Province are largely limited to the Grenville Group in the southwestern part of the province and that rocks elsewhere are interpreted as having formed a basement complex to the once widespread, thick, Grenville Group sedimentary sequence. Wynne-Edwards, *et al.* (1966), on the basis of work in western Quebec, have modified the previous interpretation to include the anorthosite–mangerite series as a post-Grenville Group igneous event, and in addition they propose that structural effects of orogenies predating the Grenvillian Orogeny can be recognized (Fig. IV-25).

Helikian

Grenville Group. The term Grenville series was first applied to the sequence of marble, quartzite, and paragneiss outcropping near the village of Grenville, Quebec, by Logan in 1863. Although positive correlation is not possible, most workers have accepted correlation of the marble-rich sequences in eastern Ontario, western Quebec, and the northwest Adirondack region with the type area. Osborne (1936) suggested that the thickness of the Grenville Group in the type area was between 5,000 and 10,000 feet and in the Haliburton–Bancroft area, somewhat less than 20,000 feet. In the Shawinigan Falls district of Quebec, he estimated that 3,000 to 5,000 feet of quartzite, garnet–sillimanite gneiss and marble overlies about 2,000 feet of amphibolite. Engel and Engel (1953) estimated about 16,000 feet of Grenville Group rocks in the northwest Adirondacks. Due to folding and flowage

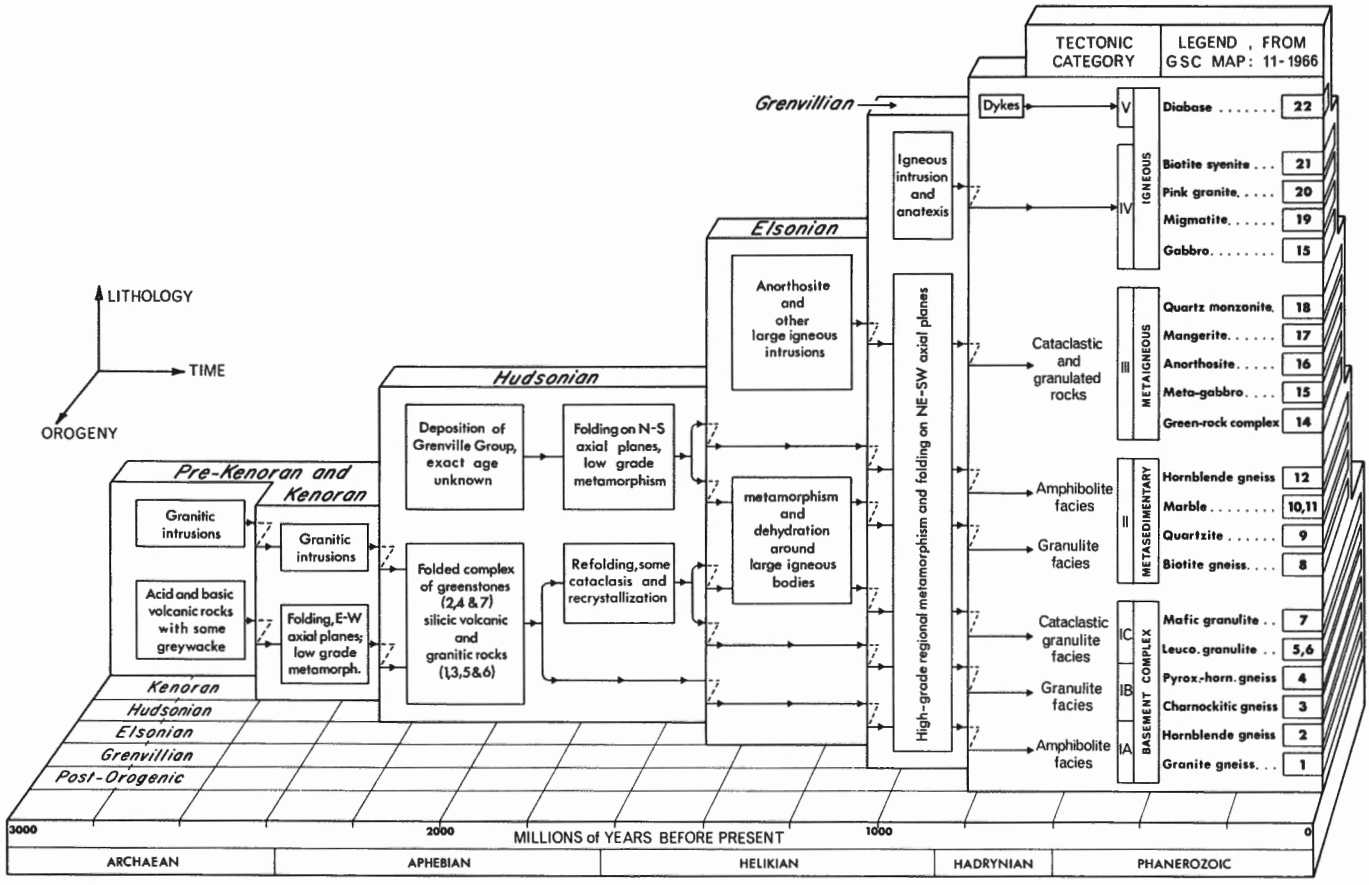


FIGURE IV-25. Tectonic steps in the evolution of Grenville Province, Mont-Laurier–Kempt Lake map-areas, Quebec (by H. R. Wynne-Edwards).

deformation these thicknesses are rather uncertain. Neither the top nor base of the Grenville Group has been defined. In eastern Ontario, basic and intermediate metavolcanic rocks occur mainly in the lower part of the exposed Grenville section, and metavolcanics are also found intercalated with metasedimentary rocks (Lumbers, 1964).

The thick metasedimentary sequences of marble, quartzite, and paragneiss typical of the Grenville Group occur only in the southwestern part of Grenville Province. Elsewhere, as for example, east of Georgian Bay, some marble occurs in amphibolite and feldspathic paragneiss but the sequence seems sufficiently dissimilar from the Grenville Group that correlation is doubtful. In the French River area clean quartzite is present but may be equivalent to the Aphebian Huronian formations northwest of the Grenville Front (Hewitt, 1960). Northeast of the type region of the group quartzitic rocks occur but marbles are rare and amphibolite and paragneiss predominate. In the Gagnon area of Quebec the appearance of marble, quartzite, and iron-formation marks the southwestern limit of rocks recognizable as metamorphic equivalents of the Aphebian Kaniapiskau Supergroup.

The K–Ar dates of about 950 m.y. associated with the Grenvillian Orogeny are a minimum age for the Grenville Group. Unequivocal evidence of the maximum age of deposition is, however, not available but a post-Archean age can be strongly supported. Hewitt (1957) concluded that the available evidence at that time favoured a late Precambrian age for the Grenville Group. Wynne-Edwards (1964) proposed a Helikian age as being most likely. Stockwell (1964) indicated that the distribution of marble and quartzite remnants across Grenville Province suggested a correlation with Huronian rocks on the southwest and with the Kaniapiskau Supergroup of the Labrador Trough to the northeast, and he adopted an Aphebian age for the Grenville Group. Wynne-Edwards, *et al.* (1966) also conclude that the Grenville Group is Aphebian. The lithological assemblage of the Grenville Group in the type region is sufficiently distinctive that arguments supporting correlation with known Aphebian rocks on lithological grounds alone are not compelling. Aphebian rocks in both the Penokean and Labrador Fold Belts have been affected by the Hudsonian Orogeny. Wynne-Edwards, *et al.* (1966) suggest that tight north-south folding of Grenville Group rocks in southwestern Quebec formed during the Hudsonian Orogeny. Northerly trending folds and accompanying intense cataclasis in part of the Morin anorthositic intrusion and the adjacent area to the east must have post-dated consolidation of the anorthosite, and therefore at least part of the north-south folding in this region is probably Neohelikian. If all the north-trending folds are of this age then the necessity for the existence of the Grenville Group at the time of Hudsonian folding is removed. Assignment of the Grenville Group to the Aphebian leaves no sedimentary sequence clearly related to the Grenvillian orogenic cycle. A Paleohelikian age is believed to be more likely and this does not seriously conflict with Rb–Sr whole rock isochron (Krogh and Hurley, 1965) and U–Pb zircon (Silver and Lumbers, 1965) ages from eastern Ontario.

Hastings Group. The Hastings Group is recognized only in southeastern Ontario. It overlies the Grenville Group but the relationship between the two groups is not certain everywhere. Lithologically, the Grenville and Hastings sequences are similar in that they contain limestone, quartzite, volcanics, and amphibolite. Conglomerate at the base of the Hastings sequence contains pebbles of quartz, quartzite, limestone, argillite, and granite and has been cited as evidence of an unconformity between the two groups. Ambrose and Burns (1956) later demonstrated the presence of two and possibly three conglomerate horizons thus casting doubt on the existence of a regional unconformity. In Barrie township, Moore (in press) reports evidence for an angular unconformity evident in the outcrop patterns, structural trends, and the presence of a basal conglomerate between Hastings Group metasedimentary rocks and Grenville Group metavolcanic and metasedimentary rocks.

Wakeham Group. Rocks of the Wakeham Group outcrop in an area of about 1,500 square miles on the north shore of the St. Lawrence River in Quebec. Estimates of about 20,000 feet for the maximum thickness of the sequence have been suggested by Claveau (1949) and Grenier (1957). Injection of numerous thick gabbro sills has added to the total thickness of the section. The group consists of pure and impure (including calcareous) varieties of quartzite with some pelitic schists and gneisses and a little conglomerate and marble. The base and top of the sequence are undefined. Grenier divided the Wakeham Group into an upper unit of calcareous quartzite with interbeds of phyllite and white quartzite, a middle unit of fine-grained white quartzite, and a lower unit of impure quartzite with minor quartz–mica schist, biotite gneiss and hematite, and rutile-bearing quartzite. Cross-bedding and ripple-marks have been found. Although there is no direct evidence of an unconformity at the base of the Wakeham Group, the granitoid gneisses east of the Caron Lake fault may represent the basement on which the Wakeham Group was deposited, because the fold pattern in the gneiss is discordant with the folds in the Wakeham Group.

The major structures in the Wakeham are upright, north-northwest trending folds that plunge southerly at moderate angles. The grade of metamorphism is upper greenschist facies over much of the area (Grenier, 1957), but may reach amphibolite facies in garnet–staurolite schist adjacent to a faulted contact with altered gabbro. In addition to the gabbro sills, younger biotite granite also cuts the Wakeham Group. K–Ar biotite dates of 845 m.y. from a staurolite schist and of 870 m.y. from the presumed basement gneiss are considered to be minimum ages for the Wakeham Group. Claveau and Grenier consider possible correlation with the Aphebian rocks of

Labrador Fold Belt. Stockwell (1964) suggests a Neohelikian age for the Wakeham Group chiefly because of Grenier's proposal that the hematite–rutile-bearing quartzites were derived from erosion of the titaniferous Allard Lake anorthosite complex. However, east of Allard Lake, Wakeham strata are intruded by syenites similar to those associated with the Allard Lake anorthosite (Retty, 1944) which suggests a Paleohelikian age or older. However, until better evidence is available the group is considered Neohelikian and correlated with the Seal Lake Group.

Southeastern Labrador. The oldest rocks in the Seal Lake area of Labrador (Fig. IV-26) are granitoids containing remnants of schist and paragneiss. These are intruded by the huge Harp Lake anorthositic pluton. On this basement, at different localities, lie the Croteau Lake, Letitia Lake, and Seal Lake Groups. The Croteau Lake Group outcrops east of the Seal Lake Group and is separated from it by a fault. The lower strata consist of black shale, minor quartzite, greywacke, dolomite, and metabasalt overlain by boulder conglomerate and arkose. The thicker upper part of the group is composed mainly of intermediate and silicic volcanic rocks that contain feldspar and quartz phenocrysts. Amygdaloidal and pyroclastic layers are common. The rocks strike northeast and dip gently southeast.

The Letitia Lake Group (Brummer and Mann, 1961) appears to correspond to the upper volcanic part of the Croteau Lake Group as described by Fahrig (1961). The lower part of the Letitia Lake Group consists of porphyries with phenocrysts of quartz and feldspar. The upper part of the group contains rhyolites, fine to coarse pyroclastics, argillites, and quartz–sericite schists. The total thickness of the group is estimated to be 8,500 feet.

The Seal Lake Group unconformably overlies the Letitia Lake Group. It is at least 34,200 feet thick. The basal, Bessie Lake Formation, 4,200 feet thick, is composed of blue, white, and pink quartzite, interbedded basalt, and basal conglomerate. It is overlain by the Wuchusk Lake Formation, 20,000 feet thick, of interbedded chert, variegated shale, phyllite, argillite, quartzite, and calcareous beds with thick diabase sills. The Whiskey Lake Formation comprises 3,000 feet of varicoloured argillaceous rocks with minor red quartzite. The overlying Salmon Lake Formation is also 3,000 feet thick. It is composed of varicoloured shale with amygdaloidal basalt and diabase sills succeeded by the Adeline Island Formation, 1,400 feet of varicoloured slate, shale, and pink and white quartzite. The Upper Red Quartzite Formation is 2,600 feet of red quartzite with pebble bands.

The Seal Lake Group has been folded into an arcuate synclinorium 75 miles long and 28 miles wide, with the folds overturned to the north. A number of subsidiary folds are superimposed on the major structure. To the northwest, relatively flat lying strata of the Seal Lake Group underlie Shipiskan Plateau. Most of the Seal Lake rocks show little effects of regional metamorphism, but shearing and schistosity become evident near the thrust zone bounding the granite in the south. The zone of major thrusting is marked by mylonite 100 yards wide.

K–Ar dates on biotite and hornblende from granitic gneiss that is unconformably overlain by the Seal Lake Group gives ages of 1,430 m.y. and 1,350 m.y., respectively. Muscovite from quartzite collected 3 miles south of the thrust zone and considered to be a metamorphic equivalent of the Seal Lake Group gave a K–Ar age of 975 m.y. A 960 m.y. whole rock K–Ar date was obtained from amygdaloidal basalt, and an age of 865 m.y. from a diabase

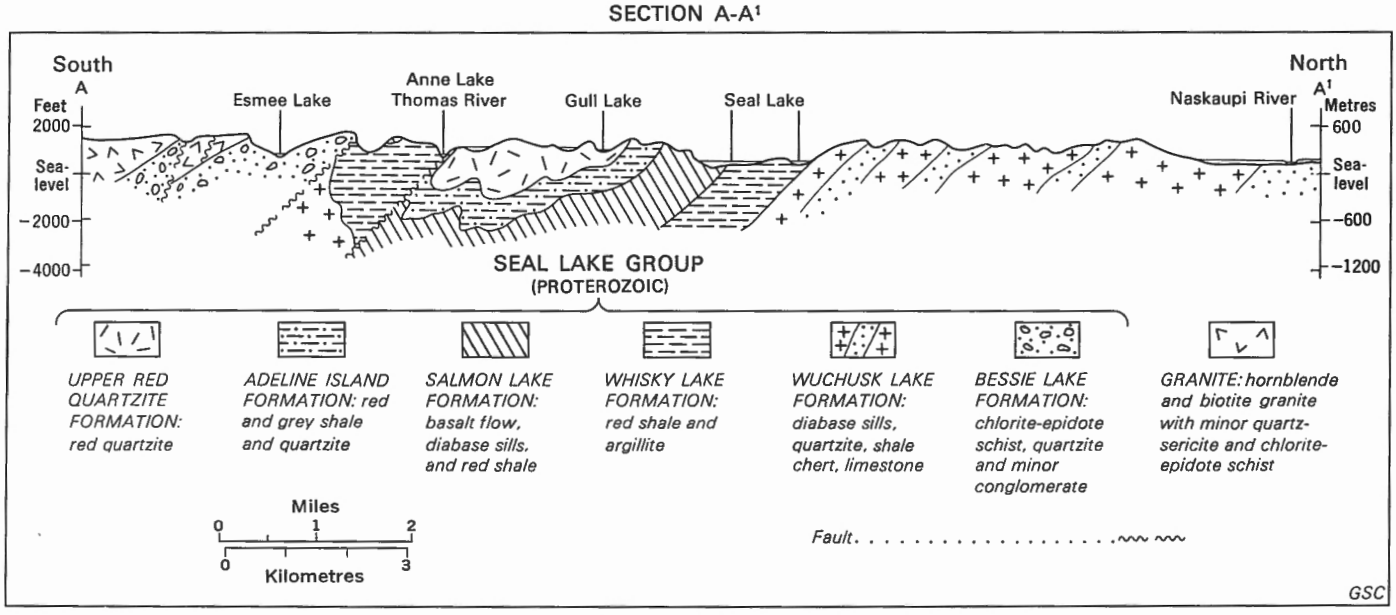


FIGURE IV-26. Structure-section through Naskaupi (Seal Lake) Fold Belt, Grenville Province, Labrador (Brummer and Mann, 1961).

sill intruding the Seal Lake Group. These dates provide an approximate minimum age for the group.

Near Aillik, Labrador, rocks of the Aillik Group consist of quartzite with interbedded carbonates, slate, and conglomerate. They may correlate with the Seal Lake Group. The rocks have been invaded by granite and partly granitized. Biotite samples from granite gneiss in southeastern Labrador yield K–Ar ages of 960 m.y. and 1,095 m.y. A biotite from a granite that invades the Aillik Group gives an age of 1,645 m.y. This latter age is from a sample very close to the Grenville Front and may be anomalous. If, however, it is valid it provides a minimum age for the Aillik Group and would rule out correlation with the Seal Lake Group.

Hadrynian

The Double Mer arkosic sandstone outcrops north of Lake Melville, Labrador. It is relatively undisturbed and is believed to be post-Grenvillian Orogeny. Kindle (1924) estimates a minimum thickness of 500 feet for the unit and describes it as well-cemented, dull red, coarse arkosic sandstone with pebble bands and local crossbedding. The contact with underlying rocks has not been described. The beds are relatively flat lying except where tilted by faulting.

Intrusive Igneous Rocks

By far the most abundant and distinctive intrusive rocks in Grenville Province are members of the Elsonian anorthosite–mangerite series. The origin of these rocks is controversial, mostly because of structural complexities and metamorphism during the Grenvillian Orogeny. They are believed to be genetically related but the nature of the relationship is not clear (Philpotts, 1966). Evidence of at least some differentiation *in situ* is often present but it is not known whether this is a major factor or if differentiation at depth or assimilation of crustal rocks have played a role in development of the diverse rock types. There is usually geological evidence to indicate that the anorthositic rocks are early members of the series, and textures and structures are commonly present that suggest that accumulation of plagioclase crystals has taken place. Evidence from Nain Province strongly suggests that basic magma has played a prominent role. Heath (1966) found that initial Sr^{87}/Sr^{86} ratios for fifteen anorthosite bodies in North America and Norway fell in the narrow range between 0.703 and 0.706 and this supports the theory of an origin from basic magma derived from the mantle or lower crust. The presence of layered structures and coarse unzoned crystals in the anorthositic rocks suggests very slow cooling in a relatively quiescent, non-tectonic environment. In southwest Grenville Province rocks of the anorthosite–mangerite series intrude the Grenville Group. Buddington (1962) has shown that the anorthositic complex in the Adirondacks was emplaced before strong deformation of the Grenville Group and that it behaved as a rigid mass during the deformation. In the Morin anorthosite–mangerite complex of southwestern Quebec the form of the rock units and the relationships among them suggest a similar conclusion. Intense cataclasis of a broad zone in the eastern part of the intrusion is consistent with failure of a rigid body.

In northern New York and southwestern Quebec the large Adirondack and Morin anorthositic plutons are marginal to the thickest sections of the Grenville Group. Although small intrusive masses of anorthositic gabbro are found within the group, large bodies are absent. For both massifs the roofs are considered to be more or less domal. A gravity survey over the Adirondack anorthosite suggests layer-like or lopolithic form. Structural data on the Michikamau anorthositic intrusion in Labrador (Emslie, 1965) also indicate a lopolithic or funnel-like form. These forms are not uncommon for large basic intrusions and imply that the magma spread along a pre-existing permissive horizon. They also suggest that the basic control on the structural position of large anorthositic plutons may have been the unconformity between the basement and the overlying, thick, relatively undisturbed Grenville sediments. Other intrusive rocks with a wide variety of petrographic characteristics, size, and form occur in Grenville Province. Some are truly post-orogenic granitic intrusions but these are probably not common (Pl. IV-20). Much of Grenville Province is believed to

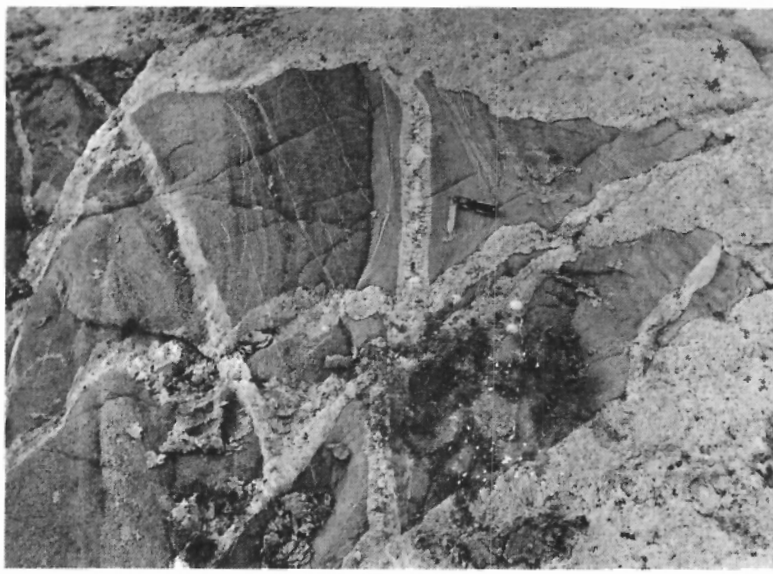


PLATE IV-20
Leucocratic coarse-grained Grenvillian monzonite of Wolfe Lake pluton intrusive into pyroxene gneiss, Grenville Province, Westport, Ontario.

represent the catazone environment and, therefore, it is difficult to distinguish synorogenic intrusions from those that predate the Grenvillian Orogeny.

Alkaline rocks occur in southwestern Grenville Province. The best-studied are the nepheline syenites of eastern Ontario at Blue Mountain, Haliburton–Bancroft, and French River (Hewitt, 1960). All have similar lithological and textural characteristics and field relations. In general, the nepheline syenites are gneissic and for the most part conformable with the paragneisses in which they occur. Pegmatitic varieties commonly show cross-cutting relations. In the Haliburton–Bancroft district nepheline-bearing rocks occur in a discontinuous band more than 80 miles long. Individual bodies rarely exceed a few hundred feet in width and two to three miles in length. Evidence for nephelinization of wall-rocks is abundant. Alkalic mafic rocks are associated with the nepheline syenite in Glamorgan and Monmouth townships. In western Quebec, a stock of coarse-grained alkaline syenite includes some nepheline syenite, and several small subcircular alkali syenite and alkali gabbro plutons also occur.

Small ultramafic bodies occur in a few areas. Most are partly altered to serpentine, tremolite, and talc but considerable amounts of fresh olivine are present. In the Gagnon–Mount Wright district several small ultramafic bodies occur with peridotitic rocks and a large gabbro complex. In the Bristol–Masham area a small peridotite body contains as much as 60 per cent olivine and enstatite with some serpentine and a little magnetite and tremolite. In the Gouin Reservoir area two ultrabasic masses with olivine, augite, tremolite, and carbonate are altered to some degree.

After the close of the Grenvillian Orogeny a number of localized igneous events took place in Grenville Province at different times. In the Chatham–Grenville stock near Grenville, Quebec (Osborne, 1936), a coarse syenite facies occurs in the western part and a granite facies to the east.

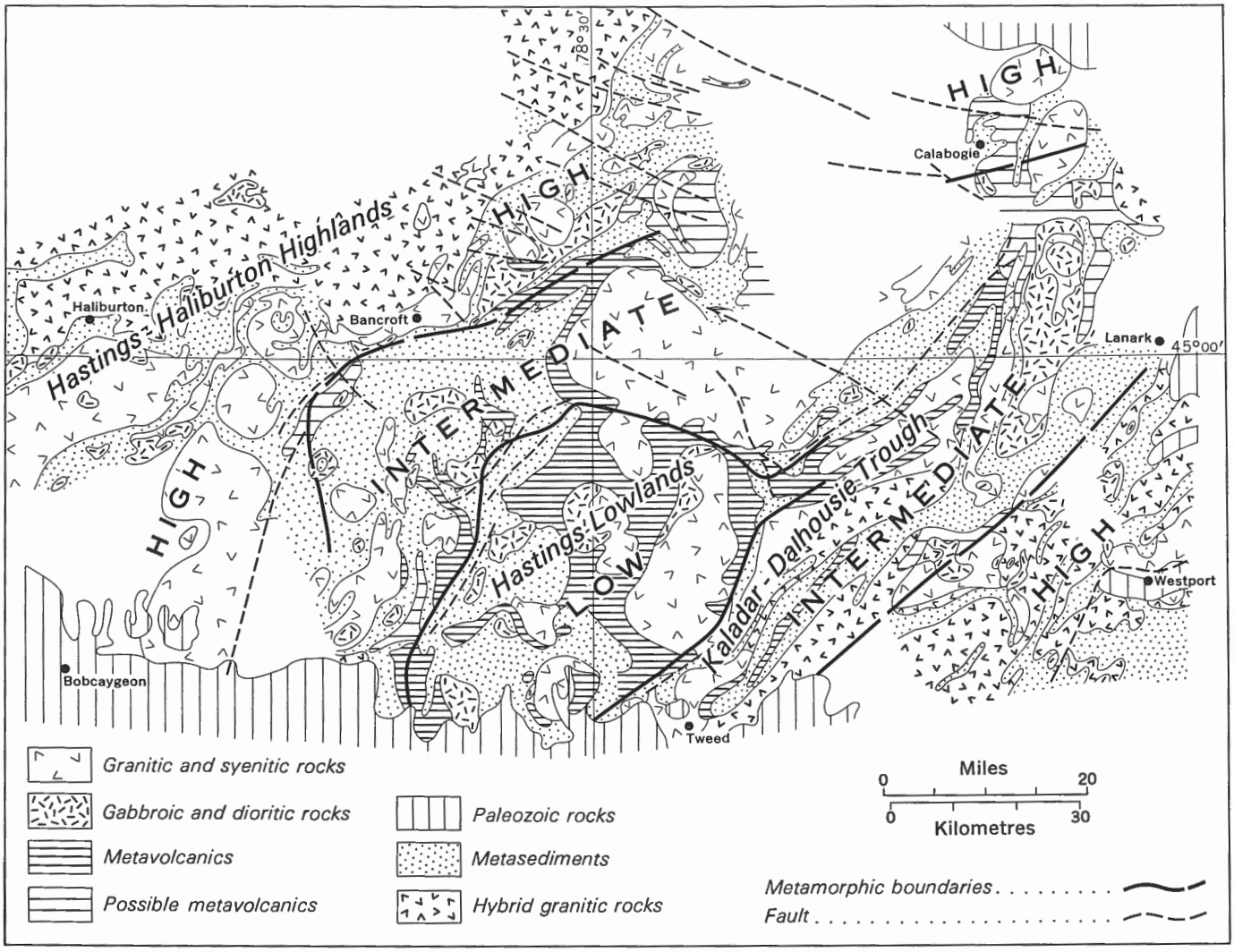


FIGURE IV-27. Geology and metamorphic zones of part of Grenville Province in southeastern Ontario (Lumbers, 1964).

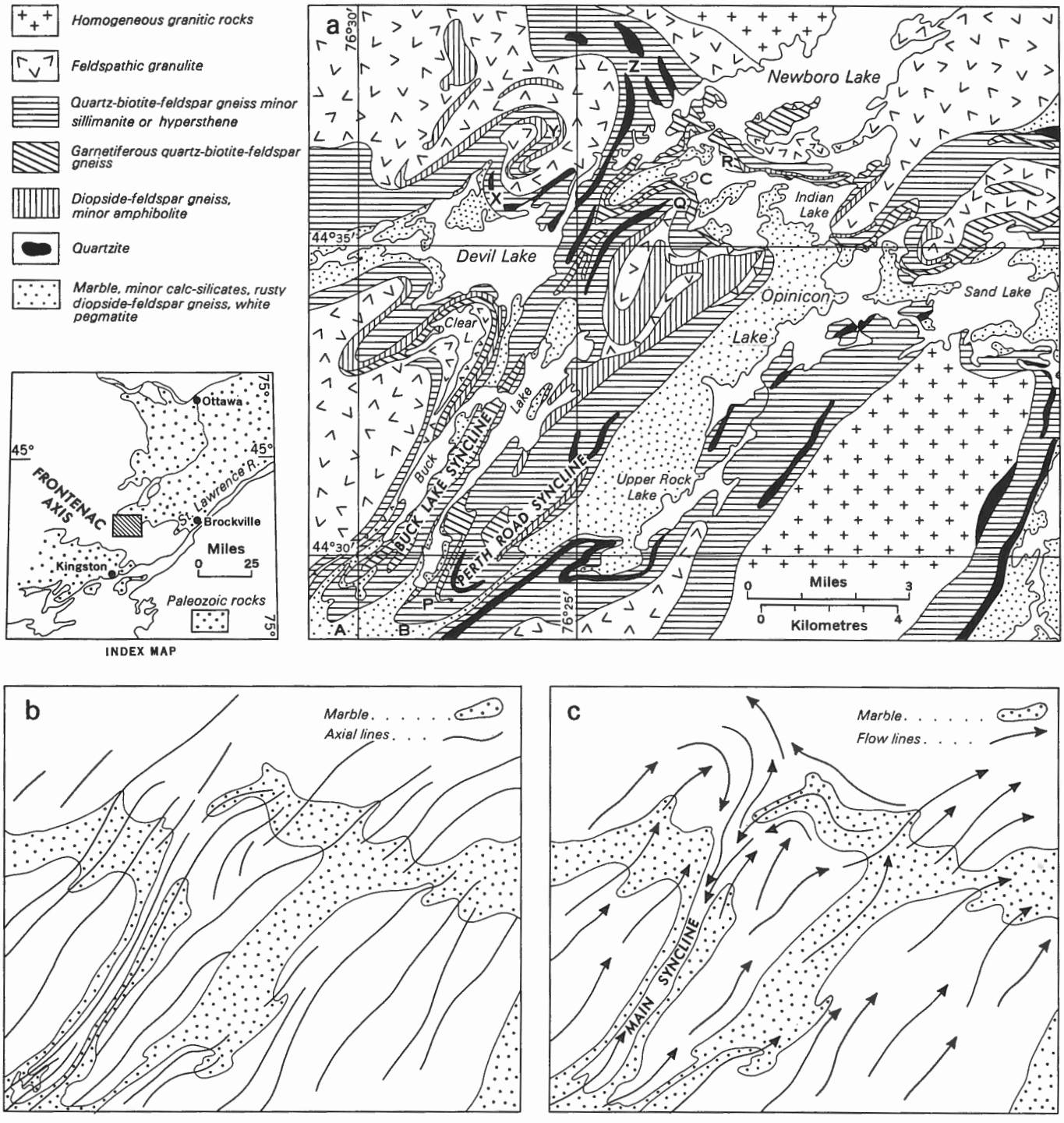


FIGURE IV-28. Geology of part of the Westport map-area, Ontario (Wynne-Edwards, 1963): (a) The folds plunge northeast at an average of 35°, so that the map is an oblique section through the structure. Stratigraphic units are repeated at successive structural levels B, C (marble); X, Y, Z (gneiss); and P, Q, R (gneiss) by refolding about the same axis. (b) Axial lines. (c) Flow pattern.

Both facies contain sodic pyroxene and amphibole. The margins are syenite porphyry. Cataclasis of the wall-rock has taken place and dykes and zones of microbreccia are considered to have accompanied intrusion. The hornblende has an average K–Ar age of 642 m.y. that is Hadrynian. An intrusion at nearby Rigaud Mountain is similar. Four small alkalic intrusive complexes occur in and around Lake Nipissing. A potassium–argon biotite date from the one on Newman Island gave an age of 560 m.y. A large diabase dyke swarm strikes east-west across eastern Ontario and western Quebec, being concentrated north of the Ottawa River. A whole rock K–Ar date from

this swarm gave an age 790 m.y. This age agrees with the date from the Chatham–Grenville stock which is known from geological evidence to be younger. Biotite ages from three samples of a lamprophyre dyke in the vicinity of Aillik in Labrador gave a mean age of 570 m.y.

Metamorphism

In northwestern Grenville Province the grade of regional metamorphism is the almandine–amphibolite facies. Kyanite occurs in rocks of appropriate composition in areas near the Grenville Front. Rocks of granulite facies are irregularly distributed, but occur mainly in the southwest, the approximate isograd between the amphibolite and granulite facies extending from near Lake Albanel southwesterly to the Ottawa River west of Ottawa (Osborne and Morin, 1962). In southeastern Ontario, rocks of the Paleohelikian Grenville Group have been raised to granulite facies during the Grenvillian Orogeny. In the northeast in the Mount Wright district, Archean granulite facies rocks have been retrograded to amphibolite facies. It is apparent therefore that high-grade regional metamorphic rocks of more than one age are present in Grenville Province, but it is difficult to distinguish them because of the lack of generally applicable criteria in recognizing and resolving the effects of polymetamorphism.

In parts of the Hastings Lowlands (Fig. IV-27), greenschist facies mineral assemblages are preserved and other criteria such as the structural character and mode of granite emplacement suggest a shallow mesozone environment. Bordering this area are higher grade rocks indicative of progressive metamorphism accompanied by a change in structural style and the character of granitic plutons that suggest a gradual transition to a catazone environment (Hewitt, 1962). High-grade gneisses with mineral assemblages of the almandine–amphibolite and granulite facies are characteristic of the Hastings–Haliburton Highlands area. The granitic plutons tend to be concordant masses. There is evidence of metasomatism in the formation of hybrid gneisses, nephelinized gneisses, certain amphibolites, and metamorphic pyroxenites. Deformation by plastic flow is characteristic. Along the southern edge of the highlands, a series of granite and hybrid granite gneiss bodies occurs which Hewitt considers is mobilized granitic material that rose by plastic flow to form domes under conditions of catazone metamorphism.

The boundary between the highlands and lowlands is not clearly defined (Best, 1966), but there is some evidence of faulting that has produced southeast-dipping zones of mylonite. Much of Hastings Lowlands is underlain by rocks of low and intermediate metamorphic grade. These rocks include schist, argillite, fine-grained "blue" limestone, marble, amphibolite, and basic to silicic metavolcanics and pyroclastics. Primary structures such as crossbedding, graded bedding, and pillows are commonly preserved. The folded stratified rocks are intruded by crosscutting syenite, gabbro, and granite plutons. Migmatites are rare. Flowage features are seen only in marble and are not common. Hewitt (1962) considers that the Hastings Lowlands area represents a mesozonal environment.

Along the Frontenac Arch, the grade of metamorphism increases to upper amphibolite and granulite facies. In the Westport area, the complex structure is interpreted by Wynne-Edwards (1963) as due to continuous non-uniform flow (Fig. IV-28). He postulates synchronous refolding as the more rapidly advancing anticlinal parts flow into the troughs of the adjacent, retarded, synclinal zones. The pattern of apparently refolded folds he interprets as having formed during a single period of deformation, as the flowage had taken place parallel with the axial planes. The Westport concordant quartz monzonite pluton is believed to have been emplaced into a vertically plunging fold or vortex formed by flowage of marble (Wynne-Edwards, 1957). The body is considered to be late or post tectonic and is cylindrical or funnel-shaped.

Isotopic Dates

It has become clear in recent years, particularly in Grenville Province, that the dating methods in common use (K–Ar, Rb–Sr, and U–Pb) and the materials dated (various minerals and whole rocks) do not necessarily date the same event. The mean of 950 m.y. K–Ar dates is commonly considered to be the time of the Grenvillian Orogeny, the time of the last major regional metamorphism causing widespread recrystallization. It is probably more correctly referred to the time at which the minerals

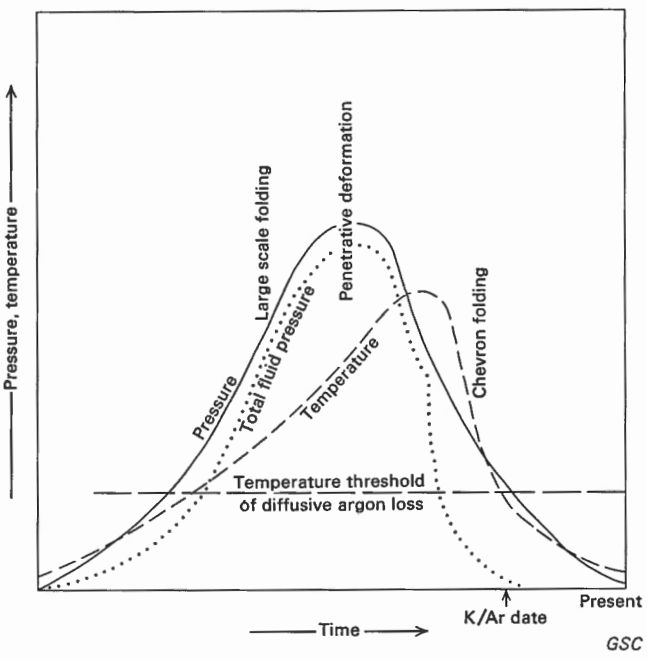


FIGURE IV-29. Inferred relationships between time, temperature, pressure, and deformation in the metamorphic interior of an orogenic belt (Armstrong, 1966). The point in time given by a K–Ar date is indicated.

or rocks were cooled below some critical temperature thus isolating the isotopic system. K–Ar dates in Grenville Province may therefore be more closely related to the rate of uplift and erosion than to the culmination of regional metamorphism (Armstrong, 1966). If other conditions are comparable, K–Ar biotite dates for example may be expected to approach more closely the time of culmination of metamorphism in the terrane of low-grade regional metamorphism than in a high-grade regional metamorphic terrane as typified by much of Grenville Province. The widespread 950 m.y. K–Ar biotite dates, although clearly referable to the Grenvillian Orogeny, probably do not reflect the culmination of regional metamorphism that likely occurred some 100 m.y. earlier as suggested by Rb–Sr and U–Pb zircon age determinations. Armstrong stresses that there is no exact correspondence between metamorphism and structural events and that metamorphism spans a period rather than having an exact age (Fig. IV-29).

Rb–Sr whole rock isochrons and U–Pb zircon ages are available from only a few areas in Grenville Province. Grant (1964) has obtained a Rb–Sr whole rock isochron from granitic rocks that straddle the Grenville Front southwest of Lake Timagami. Sampling extended 5 miles into Grenville Province and yielded an isochron age of 2,350 m.y. Rb–Sr whole rock isochrons have been determined by Krogh and Hurley (1965) for several rocks in southeastern Ontario. The Burleigh gneiss yields an isochron age of 1,103 ± 39 m.y., interpreted as a time of extensive metasomatism accompanying formation of the pluton. An isochron age of 1,285 ± 41 m.y. was found for the Blue Mountain nepheline syenite in Methuen township. The Kaladar gneissic granite adjacent to the Kaladar–Dalhousie synclinorium gave an isochron age of 1,035 ± 42 m.y. An isochron age of 1,016 ± 39 m.y. for the Westport pluton is in agreement with the geological interpretation of late tectonic emplacement. The Gananoque syenite gave an isochron age of 1,096 ± 50 m.y. Garnet–sillimanite–cordierite paragneiss from the Westport area gave an isochron age of 1,230 ± 72 m.y. Silver and Lumbers (1965) report U–Pb zircon ages from eastern Ontario. Low-grade metarhyolite from the Hastings Lowlands gave an age of 1,310 ± 15 m.y. Zircon ages from plutons throughout the region yielded two principal age groups: 1,125 ± 25 m.y. and 1,250 ± 25 m.y. Plutons of both groups are found in the Hastings–Haliburton Highlands and in the Hastings Lowlands. From northwestern Grenville Province, Davis, *et al.* (1967) obtained Rb–Sr whole rock isochron ages of 1,730 ± 75 m.y. for the French River granite, 1,500 ± 75 m.y. for granite in the Lake Muskoka area, and 1,330 ± 70 m.y. for granite south of North Bay, Ontario.

DIABASE DYKE SWARMS

The Precambrian diabase dykes and genetically related rocks are described and related to the regional geology in the preceding sections. The purpose of this section is to develop a more coherent view of these rocks in time and space. In the past any attempt to study the dyke rocks systematically was severely limited by lack of knowledge of their regional distribution and age. Recently, geological mapping and aeromagnetic surveys have established the regional distribution of virtually all the great dyke swarms, and it is now evident that many of the major dyke swarms transected the entire craton as it existed at the time of their intrusions. Several K–Ar age determinations provided the approximate ages of most of the swarms (Fig. IV-30). The average ages are based on approximately 100 determinations (Fahrig and Wanless, 1963; Burwash, *et al.*, 1963; Leech, 1966). Three types of material provide reasonably concordant K–Ar ages. These are chilled marginal dyke material, biotite concentrated from the dykes, and micas from country rocks occurring within a centimetre or so of the contact. Correlation of diabase dykes and related rocks has been aided by determining the character of their remnant magnetism (Fahrig, *et al.*, 1965). Similar paleomagnetic pole positions are evidence for correlation whereas different positions are evidence of different ages.

Diabase dykes are customarily considered to be the intrusive equivalents of tholeiitic flood basalts. In tholeiites, any olivine present generally shows reaction with orthopyroxene or pigeonite. The predominant pyroxene is Ca-poor augite and interstitial quartz may be present. They are hypersthene normative. The other great basalt group, the alkali basalts, contains olivine showing growth parallel with that of Ca-rich augite. Pigeonite and hypersthene are rare or absent. These basalts are nepheline normative. It has been shown that different dyke swarms of the District of Mackenzie have different K_2O contents that are characteristic of the swarms (Burwash, *et al.*, 1963). Leech (1966) indicated that the increase in K_2O content is related to the decreasing age of the swarms with the possibility of the existence of more complex relationships. For example, the increase in alkalinity may not be a simple time function but may be influenced by the tectonic history of the areas cut by the dykes, in which case each orogenic region may exhibit a sequence of basaltic swarms whose alkalinity increases with decreasing age. Both tholeiitic and alkalic basalts form diabase dyke swarms in the Canadian Shield (Fahrig, *et al.*, 1965). The average alkali–silica ratios of thirty-one swarms (Fig. IV-31) have been divided into two groups: Paleohelikian and older, and Neohelikian and younger.

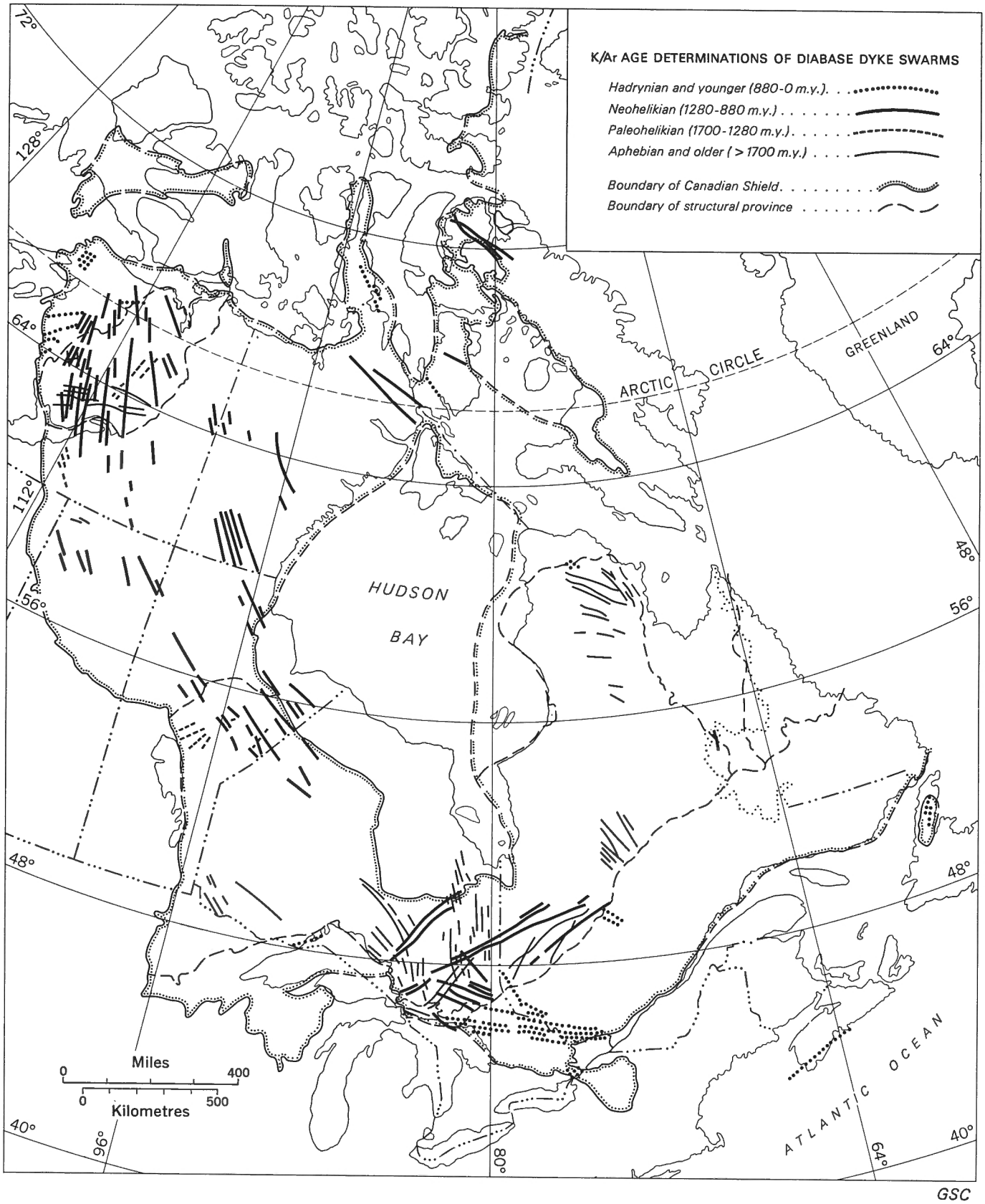


FIGURE IV-30. Diabase dyke swarms of the Canadian Shield (by W. F. Fahrig).

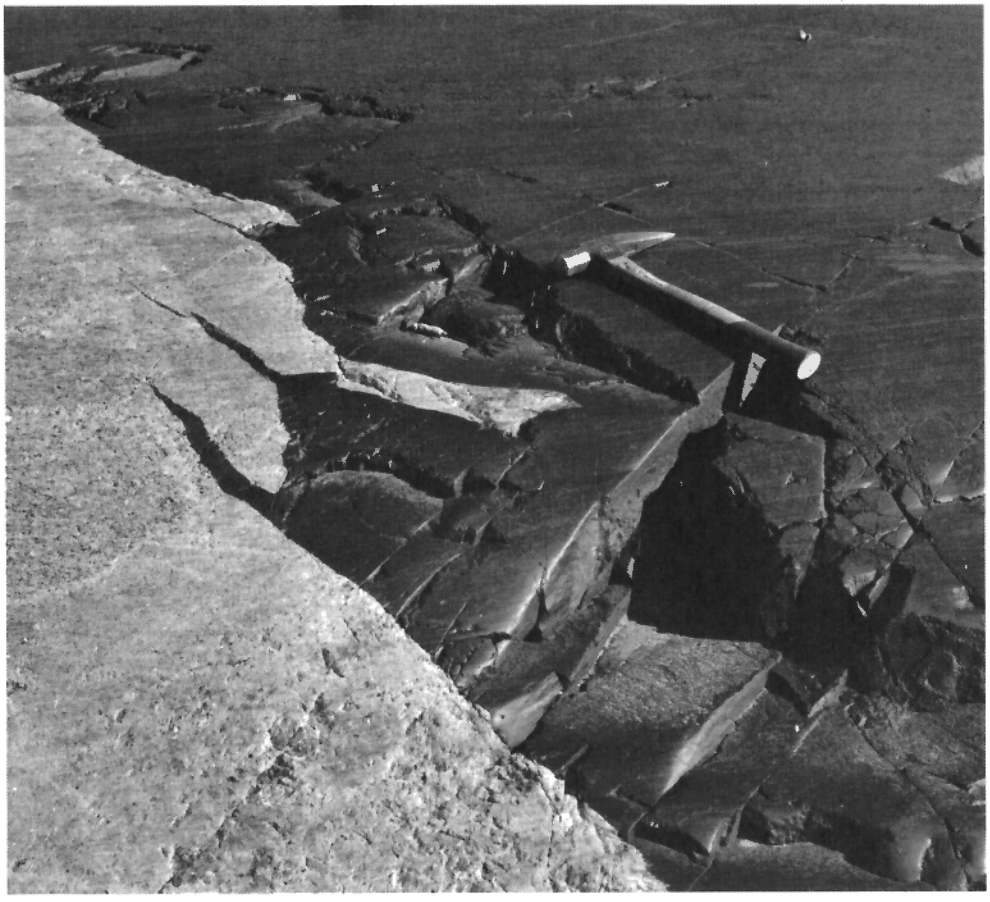


PLATE IV-21
Contact between diabase dyke and granite near Povungnituk, east coast of Hudson Bay, Superior Province, Quebec.

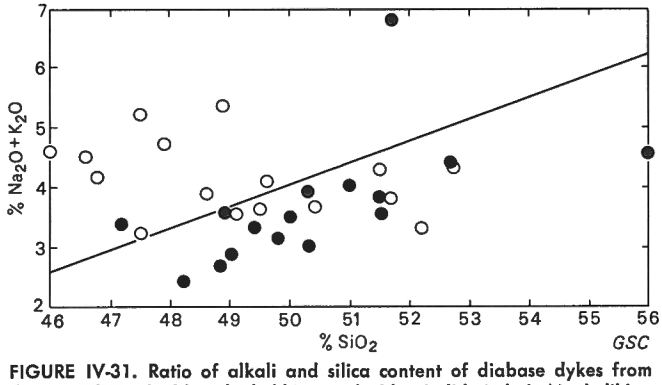


FIGURE IV-31. Ratio of alkali and silica content of diabase dykes from the Canadian Shield: Paleohelikian and older (solid circles); Neohelikian and younger (open circles). The diagonal line separates rocks with tholeiitic mineral composition (lower right field) from rocks with alkalic mineral compositions (according to G. A. MacDonald and T. Katsura).

There is a tendency towards increasing alkalinity with decreasing age. This is not a sharp division, as there is at least one known pre-Neohelikian alkali basalt swarm, and there are several Neohelikian and younger tholeiitic swarms. In addition, some of the swarms whose data are plotted on the figure may be duplicates as some samples although widely separated may represent the same swarm.

It is apparent that alkali basalt dykes are intrusive into the craton and are chiefly Neohelikian or younger in age. The change in chemistry of dyke-fed basaltic material may reflect fundamental differentiation taking place in the mantle. However, it has yet to be established whether the increase in alkalinity was gradual, or whether a rather abrupt change occurred at some stage, perhaps at the end of the Paleohelikian. Poldervaart's formula has also been used where suitable chemical data were available together with the alkali silica ratios to illustrate the relationship

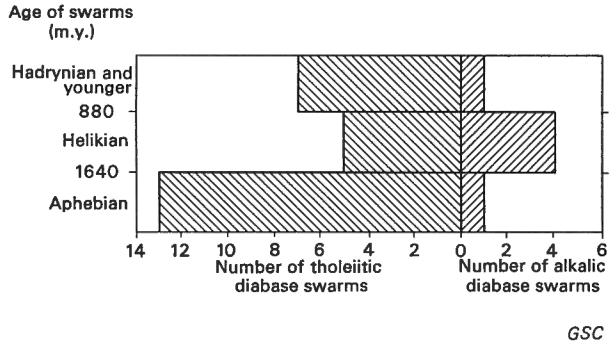


FIGURE IV-32. Classification of basaltic dyke swarms according to Poldervaart's formula.

between the number of dyke swarms, their ages, and their classification (Fig. IV-32). As basaltic rocks have contributed significant amounts of detritus to Precambrian sedimentary basins, the change in their chemistry should be reflected in a progressive change in the chemistry of the sediments and in the gneisses and other crystalline rocks that have resulted from the ultra-metamorphism of the sediments.

The presence of large numbers of diabase dykes of various ages indicates that large volumes of subcrustal material were added to the continental crust throughout the geological past. This material provides the best sample of upper mantle derived-material that is available from within continental areas and provides valuable data on the variation with geological time of the abundances of major and minor elements supplied from below the crust. The

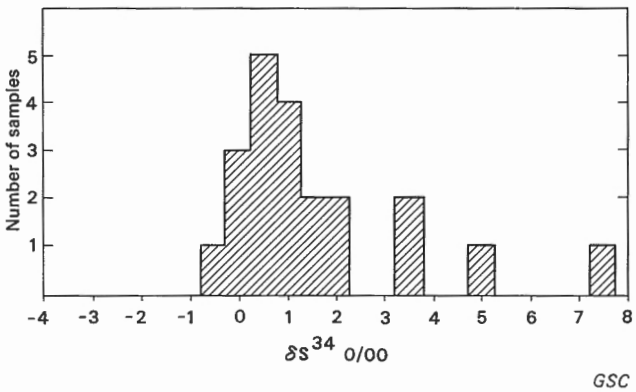


FIGURE IV-33. Sulphur isotope ratios of sulphides from diabase dykes of the Canadian Shield (by Akira Sasaki).

abundance of heat-producing elements, K, U, Th, and the isotope ratios of elements such as S and Pb are of special interest. On Figure IV-33 the sulphur isotope ratios of sulphides from twenty-one Canadian Shield diabase dykes are shown, whose ages range from Aphebian to early Paleozoic. The $\delta^{34}0/00$ is an expression that relates the S^{32}/S^{34} ratios of the samples to that of a standard sample, in this instance, to sulphur in troilite of the Canon Diablo meteorite. The grouping of the $\delta^{34}0/00$ values near the zero point indicates that the sulphides of these dykes have S^{32}/S^{34} ratios similar to the meteorite value. This value represents the ratio of sulphur in basalt generated in the upper mantle.

Diabase dykes are probably indicative of tensional conditions in the crust, some of continent-wide dimension which have resulted in continental stretching (Pl. IV-9, -21). The mechanism that caused this tension is likely to have acted in a more extreme fashion at various times and to have actually resulted in the breakup of continental masses. Diabase dyke swarms can be considered as direct evidence of continental drift. In addition, they are features of such continuity that the reassembling of continental masses into hypothetical proto-continents requires that these linear features pass from one segment to another. Payne, *et al.* (1965) have shown that, if Greenland and Scotland are moved along paths determined by physical outline and paleomagnetic studies, ancient dyke swarms more than 2,000 m.y. old in Greenland, Scotland, and Canada fall on a great circle (Fig. IV-34). They suggest that this continuous line of intrusion of more than 2,500 miles presents independent evidence of continental displacement.

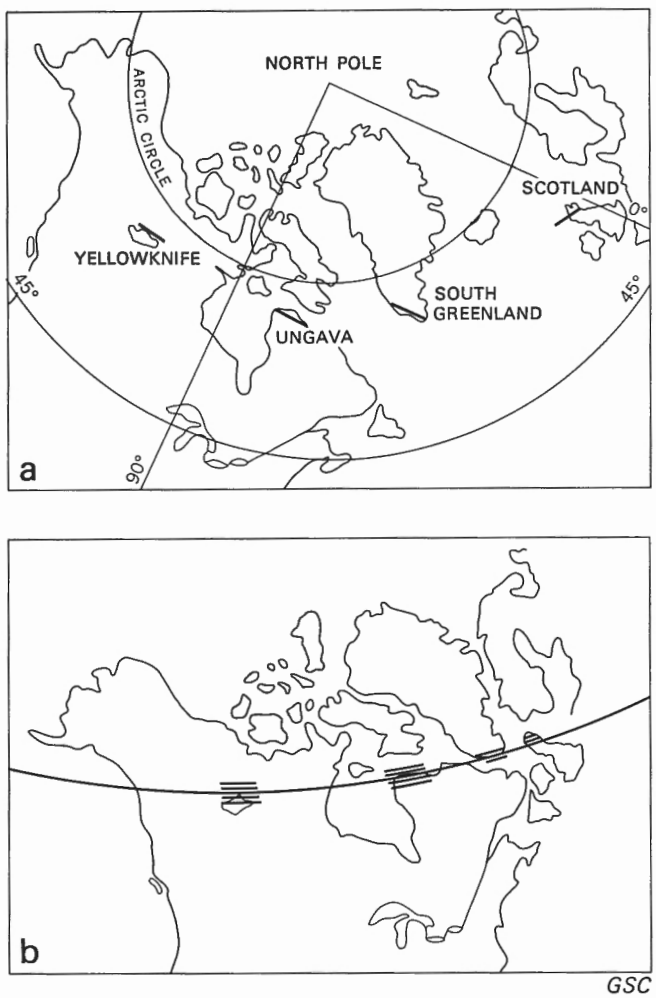


FIGURE IV-34. Diabase dyke swarms, 1,900 to 2,200 m.y. old, of Canada, Greenland, and Scotland: (a) present position; (b) hypothetical restoration as landmass of Laurasia (Payne, et al., 1968).

SHOCK METAMORPHIC STRUCTURES

Some subcircular structures in the Canadian Shield are characterized by the occurrence of shock metamorphic effects evident in breccias, rocks of volcanic appearance, and abnormal glassy or porcelain-like appearance of the country rocks. In shape they are either simple bowl-shaped depressions with a raised bedrock rim, or shallow depressions with little or no rim, but with a prominent mound in the centre that may be higher than the rim. The diameters range from 6,500 feet to 38 miles. In general the smaller structures are the simple depressions and the larger ones are the central peak depressions, but there are exceptions. About sixteen structures are known (Fig. IV-35). The ages of most structures, about half of which are associated with Paleozoic sediments, fall in the range 200 to 450 m.y. Some of the Canadian shock metamorphic structures may be the result of meteorite impact. Others are the result of geological processes related to the intrusion of alkaline igneous rocks.

The breccia is polymict with a vast range in size of fragments. Glass may be present as fine splinters in the matrix, or as irregular balls and flaps. Different types of glass and rapid variation in the properties of a single glass fragment occur. The breccia matrix in some cases may be partly welded and show flow structure. Fragments of non-glassy material often show shock metamorphism. The breccias are termed suevite. Several shock metamorphic features commonly occur together, either on breccia fragments or in the country rocks. Planar features are fine cleavage and slip lines, commonly seen in quartz and feldspar. Occasionally they may be seen in hand specimen of quartz as a very fine, regularly spaced cleavage. Under the microscope they appear as closely spaced, fine, dark lines. Orientations in quartz parallel the basal prism (0001) and the rhombohedron $(01\bar{1}3)$ are particularly diagnostic of shock metamorphism. Thetomorphic glass partly or entirely preserving the form of felsic minerals is common. Mafic minerals are little affected. Plagioclase glass (maskelynite) is more common, but quartz glass is also found. The refractive index and density are higher than that of the fused minerals. Melt glass with vesicles and streaky flow structure are less commonly found. Kink bands occur on biotite and occasionally hornblende. Shatter cones with parasitic radiating striae (horsetailing) on the surface are present. The cones occur as both positive and negative casts.

Fracturing of country rocks of unusual style is commonly found in and near shock metamorphic sites. The joints have rough surfaces and sometimes a white coating. The number of joint sets is abnormally high, and their regularity of attitude abnormally low. The regional joint sets in the vicinity are commonly deflected. Small scale normal faults occur in the rim of the structures.

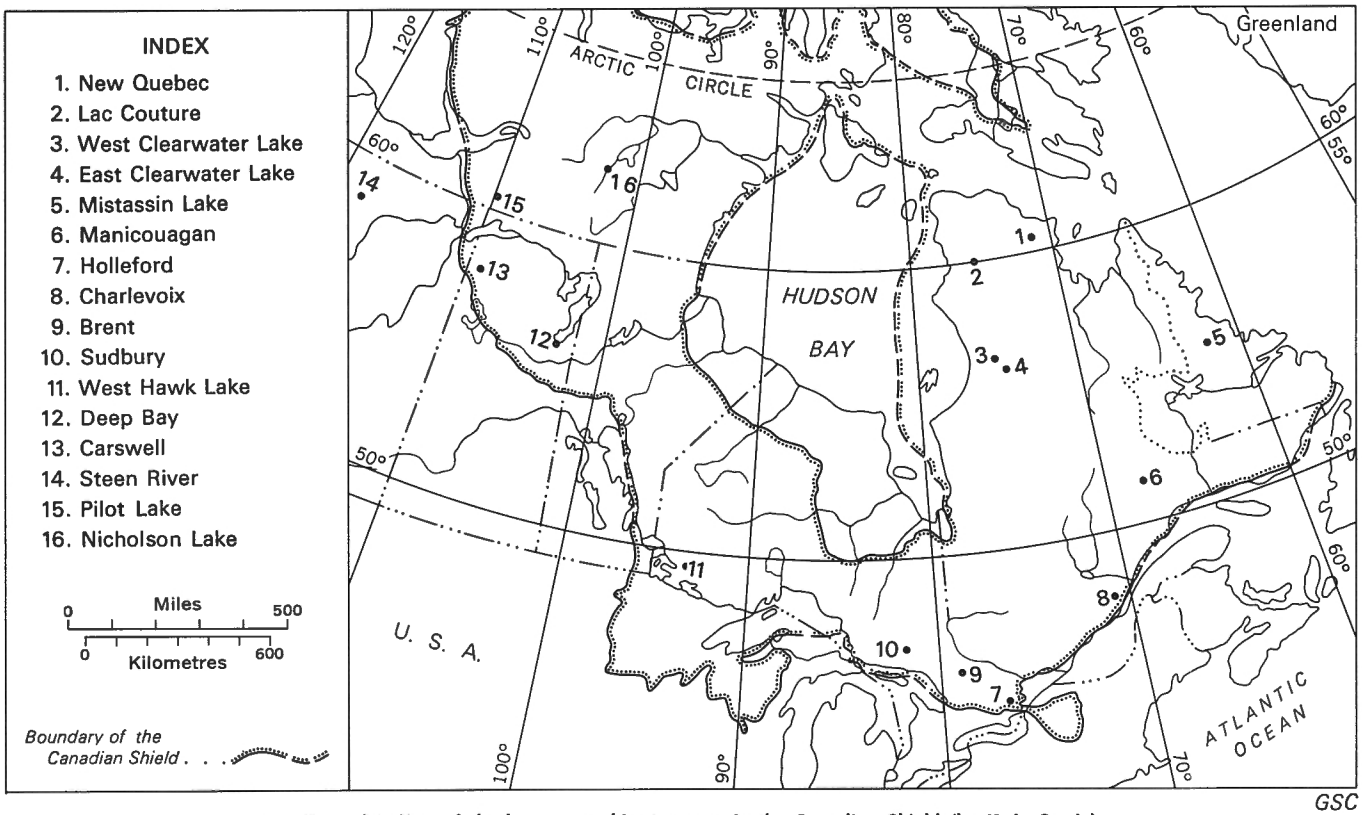


FIGURE IV-35. Sites of shock metamorphic structures in the Canadian Shield (by K. L. Currie).

Rocks of igneous appearance, that is with textures of normal igneous rocks, are aphanitic to fine grained with few or no inclusions. In some samples the contacts with the country rocks are sharp; in others they grade through an inclusion-rich zone into breccia. Dykes are common but small and short. Those analyzed are andesitic or dacitic in composition, moderately to strongly alkaline.

Some shock metamorphic structures seem to be funnel-like depressions underlain by a mass of breccia that narrows downward. Large scale depression or upheaval of the bedrock does not occur in their vicinity. The others, with central uplifts, are associated with large scale down-faulting at their margins that, together with the large uplift at the centre, has the effect of preserving some older sedimentary rocks, generally of Paleozoic age. Shock metamorphic effects are generally intense in the central uplift. The volcanic rocks appear to have been fed from fissures near the outside of the structure.

The New Quebec crater is a slightly glaciated, simple crater of late Pleistocene age, 10,400 feet in diameter, 825 feet deep (Currie, 1965). Granitic gneiss in the vein is hydrothermally altered to epidote and calcite. There is a marked bending of the regional joints near the crater. Two boulders bearing thetomorphic glass in a matrix of volcanic material were recovered southwest of the crater. A subcircular negative gravity anomaly of 6 milligals suggests the presence of breccia beneath the structure.

The West Clearwater Lake crater, Quebec (Bostock, 1968) is a nearly circular depression 20 miles in diameter, bounded by series of curving linears (Pl. IV-22). In the centre is a ring of islands about 8 miles in outside diameter and half a mile wide. The stratigraphic succession on these islands is fractured and altered country rocks with remnants of Ordovician limestone, glass-bearing breccia cut by slightly different breccias, inclusion-rich dacite, and inclusion-poor dacite. The dacite is markedly richer in potash and magnesia, and poorer in soda than the country rocks. Two K–Ar dates average 290 m.y. Drilling has shown dykes of dacite to persist at least to 1,300 feet. Complex hydrothermal mineralization of gypsum, tetrahedrite, and millerite is associated with unusual alkaline ultrabasic veinlets. An irregularly shaped negative Bouguer anomaly of 8 milligals occurs over the crater. Intensely shock-metamorphosed country rocks form shoals in the centre of the lake. The East Clearwater Lake crater is a quadrate basin about 14 miles on each side separated from West Clearwater Lake by a screen of islands (Dence, 1968). Much of the lake is more than 300 feet deep. The central uplift is composed of fractured but almost unmetamorphosed basement; its size is uncertain. It is surrounded by at least 300 feet of massive and brecciated to massive, dark rocks of igneous appearance. A simple, subcircular, negative Bouguer gravity anomaly of about 10 milligals is over the crater.

Manicouagan is the largest and most complex crater in Canada; it is octagonal and is 38 miles across. The central peak is rectangular, about 6 by 12 miles, and towers 1,700 feet above its surroundings. All types of shock metamorphic features noted above are present in the region of the central uplift. Three distinct varieties of igneous rocks are present. Pipes and small flows of meimechite and alkali basalt are the older, while vast volumes of trachyandesite and larvikite are the younger. The stratigraphic sequence is similar to that at Clearwater

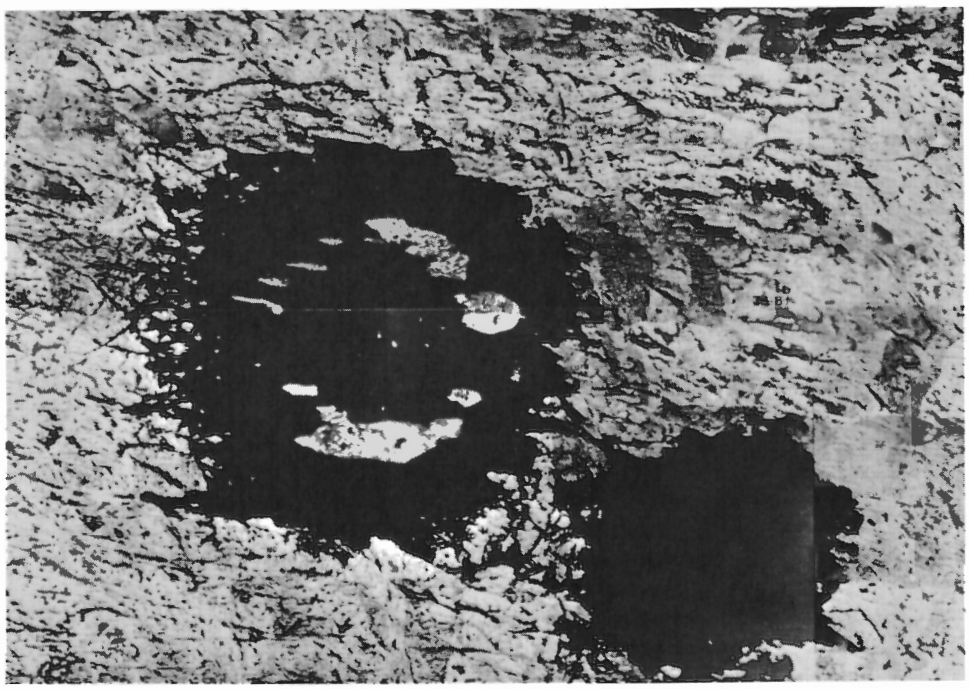


PLATE IV-22

Vertical air photograph mosaic of Clearwater Lake shock metamorphic structure, Clearwater Lake, Quebec.

but breccia is almost absent. A large number of K–Ar ages average 210 m.y.

The Brent crater, Ontario, is a simple structure without a rim, about 11,000 feet in diameter, filled with flat-lying Ordovician sedimentary rocks 800 feet thick (Dence, 1968). Beneath them is 2,000 feet of breccia displaying shock metamorphic effects cut by alnoite dykelets. Carbonate nodules in the dykelets have oxygen isotope ratios characterisitic of carbonatite. A mass of hyper-potassic trachyte is at a depth of 2,900 feet, below which brecciation rapidly disappears. An age determination on the alnoite gave 590 m.y. and three dates from the trachyte average 330 m.y. Lamprophyre dykes radial to the crater are present in the country rocks.

The Carswell structure, Saskatchewan, is 24 miles in diameter. It comprises an outer ring of complexly folded Carswell dolomite downfaulted into the Paleohelikian Athabasca Formation, and an inner core of Archean or Aphebian crystalline rocks. Relative displacements on the faults are thought to be of the order of 4,000 feet. Trachyandesite is present as dykes. They have a K–Ar age of 475 m.y. Poorly developed shatter cones occur in the basement rocks. Shock metamorphism is not found in the sedimentary rocks. Clear-cut evidence of the action of ultrahigh pressures has not been found in Canadian occurrences of shock metamorphism (with the dubious exception of one coesite deposit). The impact theory does not explain the presence of coherent blocks of rock raised in the centre of eight of the structures. The presence of alkaline igneous rocks in at least four of the craters, in particular alkaline rocks usually associated with explosive activity, strongly suggests an endogenetic origin for at least some of the craters.

HUDSON PLATFORM

The Hudson Platform lies in the central part of the Canadian Shield and comprises the remnants of the Moose River and Hudson Bay sedimentary basins. The smaller Moose River Basin borders and extends beneath James Bay. The larger Hudson Bay Basin to the north is mainly covered by Hudson Bay. Only parts are present on land in northern Hudson Bay Lowland and on the islands of Southampton, Coats, and Mansel. Separating the two basins is the northeast-trending Cape Henrietta Maria Arch (Fig. IV-36). Two structural provinces of the Canadian Shield, the Superior and the Churchill, border and extend beneath Hudson Platform. Their boundary extends from the vicinity of Nelson River, Manitoba, eastward and southeastward beneath the Paleozoic cover of Hudson Bay Basin to reappear in the Sutton Inlier on Cape Henrietta Maria Arch; from there it extends beneath the Paleozoic and Mesozoic rocks of Moose River Basin beneath James Bay to the east side of Hudson Bay. Within Superior Province, plutonic rocks of Archean age form most of the Precambrian basement. Rocks of Proterozoic age dominate Churchill Province; some are plutonic crystalline rocks and others, as in the Sutton Inlier, are slightly altered and little deformed Aphebian sedimentary rocks.

Phanerozoic strata of Hudson Platform consist of nearly flat lying to gently dipping sedimentary rocks (Fig. IV-37). In Moose River Basin, rocks of Ordovician, Silurian, and Devonian ages are represented by a variety of shallow marine facies composed of sandstone, shale, limestone, dolomite, and evaporite deposits that are succeeded by non-marine sediments of Lower Cretaceous age.

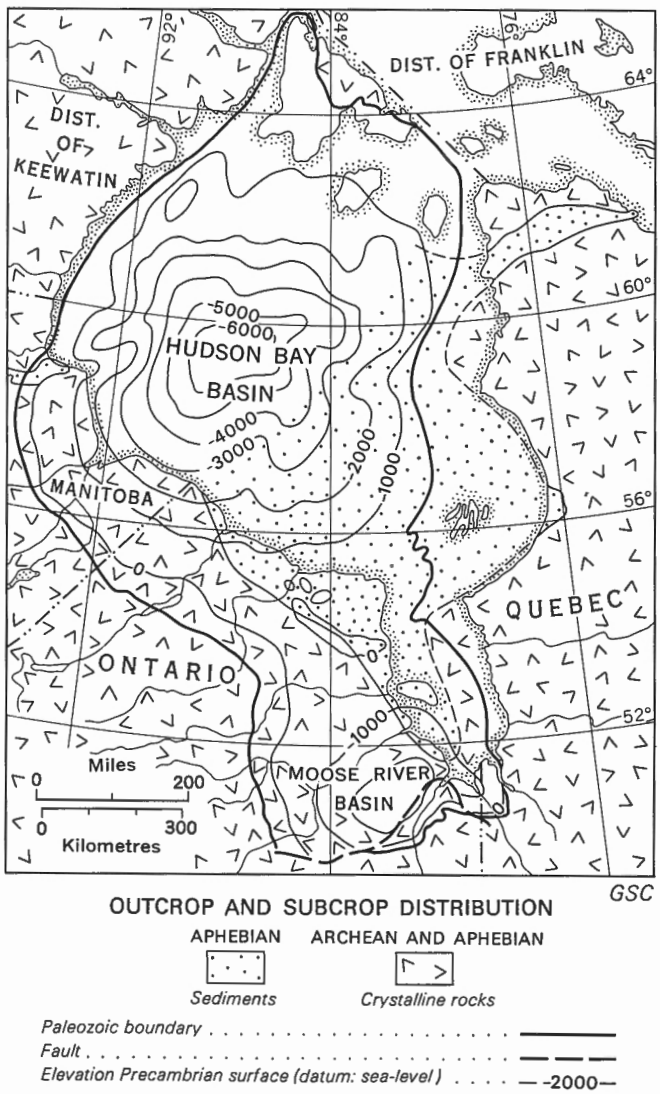


FIGURE IV-36. Configuration of Precambrian rocks beneath Paleozoic cover of Hudson Platform (modified from Hobson, 1967).

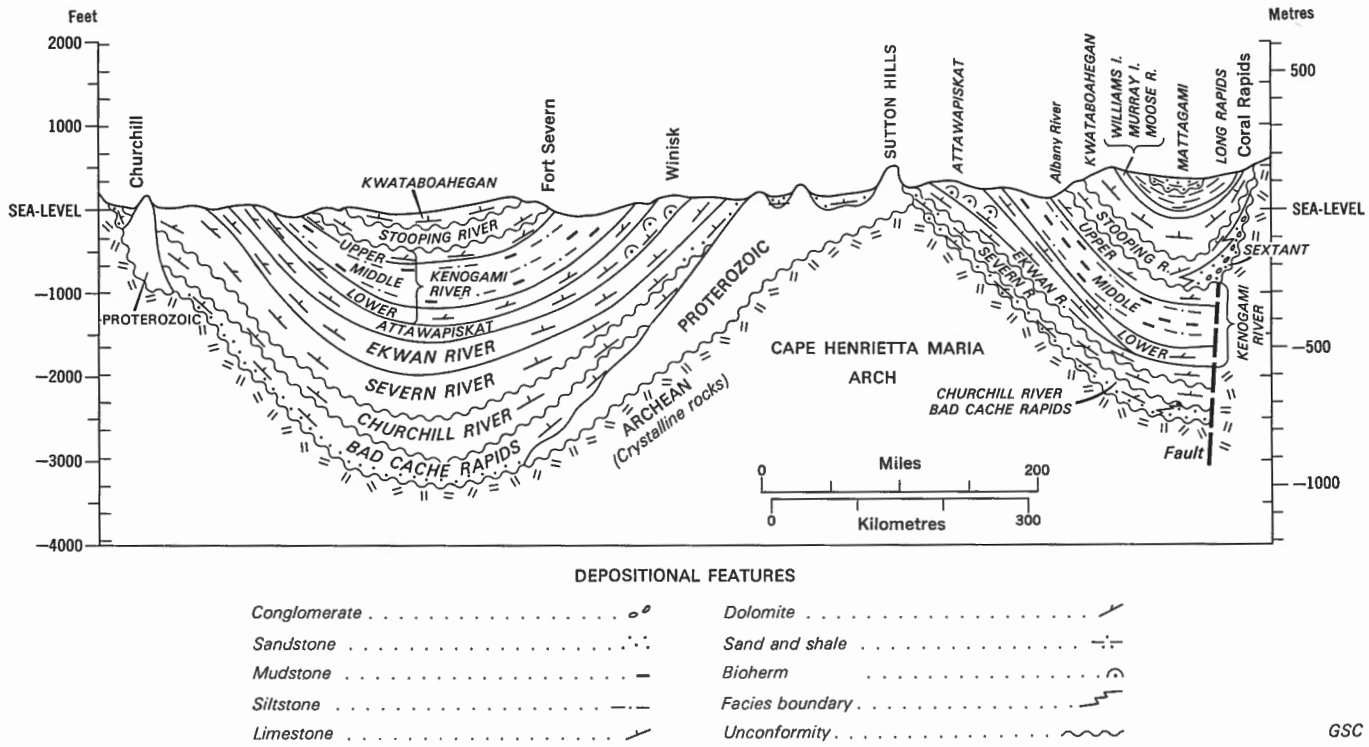


FIGURE IV-37
Cross-section of Phanerozoic rocks of southern Hudson Platform (by B. V. Sanford and A. W. Norris).

Their combined thickness is about 2,500 feet in the southern part of the basin. In Hudson Bay Basin, similar rocks of Ordovician, Silurian, and Lower and early Middle Devonian ages are represented on the mainland where they have a combined thickness of about 3,000 feet in Manitoba. In the northern part of the basin, only Ordovician and Silurian rocks are exposed on the islands of Southampton, Coats, and Mansel. A thicker Phanerozoic sequence, 6,000 feet as determined by geophysical surveys (Hobson, 1967), is presumably present offshore in the central part of the basin beneath Hudson Bay. Paleozoic outliers are present on the Canadian Shield at Clearwater Lake, Waswanipi Lake, Lake St. John, Lake Nipissing, Lake Timiskaming, Manicouagan Lake, and Mattawa. The Hudson Bay and Moose River Basins are now separated by parts of the Canadian Shield from Williston Basin in the southwest and from the Michigan and Quebec Basins to the south. The northern boundary of Hudson Bay Basin is Bell Arch, trending northwestward through Southampton Island and bordering Foxe Basin of the southeastern Arctic Platform.

PALEOZOIC

Ordovician

Tectonic Summary

Although the first widespread Paleozoic submergence of the southern Canadian Shield was during the Middle Ordovician (Wilderness), it was possibly not until the Late Ordovician (Edenian) that the seas finally transgressed the central part of the Canadian Shield to initiate the Hudson platformal sequence (Fig. IV-38). In the Edenian, arkose, orthoquartzitic sandstone, and shale were deposited as a thin veneer over the irregular Precambrian surface, succeeded by marine fossiliferous limestones and minor dolomites (Pl. IV-23). The Maysvillian was presumably a time of uplift of parts of the Canadian Shield, particularly Severn and Fraserdale Arches, during which the Edenian on the southern margin of Hudson Bay Basin was bevelled by erosion and, if formerly present, was removed entirely from Moose River Basin south of Cape Henrietta Maria Arch. During the Richmondian, a widespread transgression again connected Hudson, St. Lawrence, Interior, and Arctic Platforms. Fossiliferous limestones were deposited along the present margins of Hudson Platform during this stage, giving place to dolomitic limestones and dolomites with interbeds of gypsum or anhydrite in central Moose River Basin and possibly also in the central Hudson Bay Basin. The Fraserdale Arch south of Moose River Basin formed a highland from which coarse detritus was derived, represented by sandstone along the southern margin of the basin.

Upper Ordovician

Rocks of early Late Ordovician (Edenian) age in Hudson Bay Basin form the Bad Cache Rapids Group, and may be present also in Moose River Basin. The group is divisible into two formations (Nelson, 1964). The basal Portage Chute Formation consists of arkose, orthoquart-



PLATE IV-23. Upper Ordovician Bad Cache Rapids Group unconformably on Precambrian granitic gneiss, Churchill River, Manitoba.

zitic sandstone, and argillaceous or calcareous sandstone and shale, 4 to 13 feet thick, succeeded by 71 feet of light grey, microcrystalline, dolomitic, locally bioclastic, and nodular limestone. It is overlain by the Surprise Creek Formation composed of 64 feet of light yellowish tan and grey, mottled, finely crystalline, cherty, dolomitic limestone. A similar lithological and faunal assemblage overlies Precambrian crystalline rocks on Southampton Island (Nelson and Johnson, 1966), which suggests a wide distribution of the Bad Cache Rapids Group beneath Hudson Bay. In addition, outliers of rocks in part equivalent to the Bad Cache Rapids are present at Clearwater Lake, east of Hudson Bay, and in two isolated localities west of the bay. The Bad Cache Rapids Group is faunally and lithologically similar to the Red River Formation of the Lake Winnipeg area of southern Manitoba (Nelson, 1964).

The Churchill River Group of late Late Ordovician (Richmondian) age disconformably succeeds the Bad Cache Rapids Group in Hudson Bay Basin. Churchill River strata rise to surface in the western Hudson Bay Lowland, but are in fault contact with Precambrian rocks along the southwestern margin of Moose River Basin. They again rise to surface immediately south of James Bay, and on Southampton Island in northern Hudson Bay. The Churchill River Group in the southern Hudson Bay Basin is divisible into the Caution Creek Formation that consists of bioclastic, grey to yellowish grey and tan, mottled, cryptocrystalline dolomitic limestone, 42 feet thick, and into the overlying Chasm Creek Formation of massive to rubbly-bedded, light yellowish grey to buff, locally fucoidal, microcrystalline limestone, 190 feet thick (Nelson, 1964). A similar lithological and faunal succession occurs on Southampton Island except that the uppermost light brownish grey, finely crystalline limestones locally contain dark brown oil shale interbeds (Nelson and Johnson, 1966). In central Moose River Basin rocks possibly equivalent to the Churchill River Group consist of about 300 feet of grey-tan, microsucrosic dolomite containing interbeds of gypsum and anhydrite. Towards the southern margin of the basin, the dolomite becomes increasingly arenaceous, and eventually grades into 70 feet of arkose, granite pebble-conglomerate, minor shale, yellowish dolomitic sandstone, and sandy dolomite bordering the Canadian Shield (Remick, Gillain, and Durden, 1963).

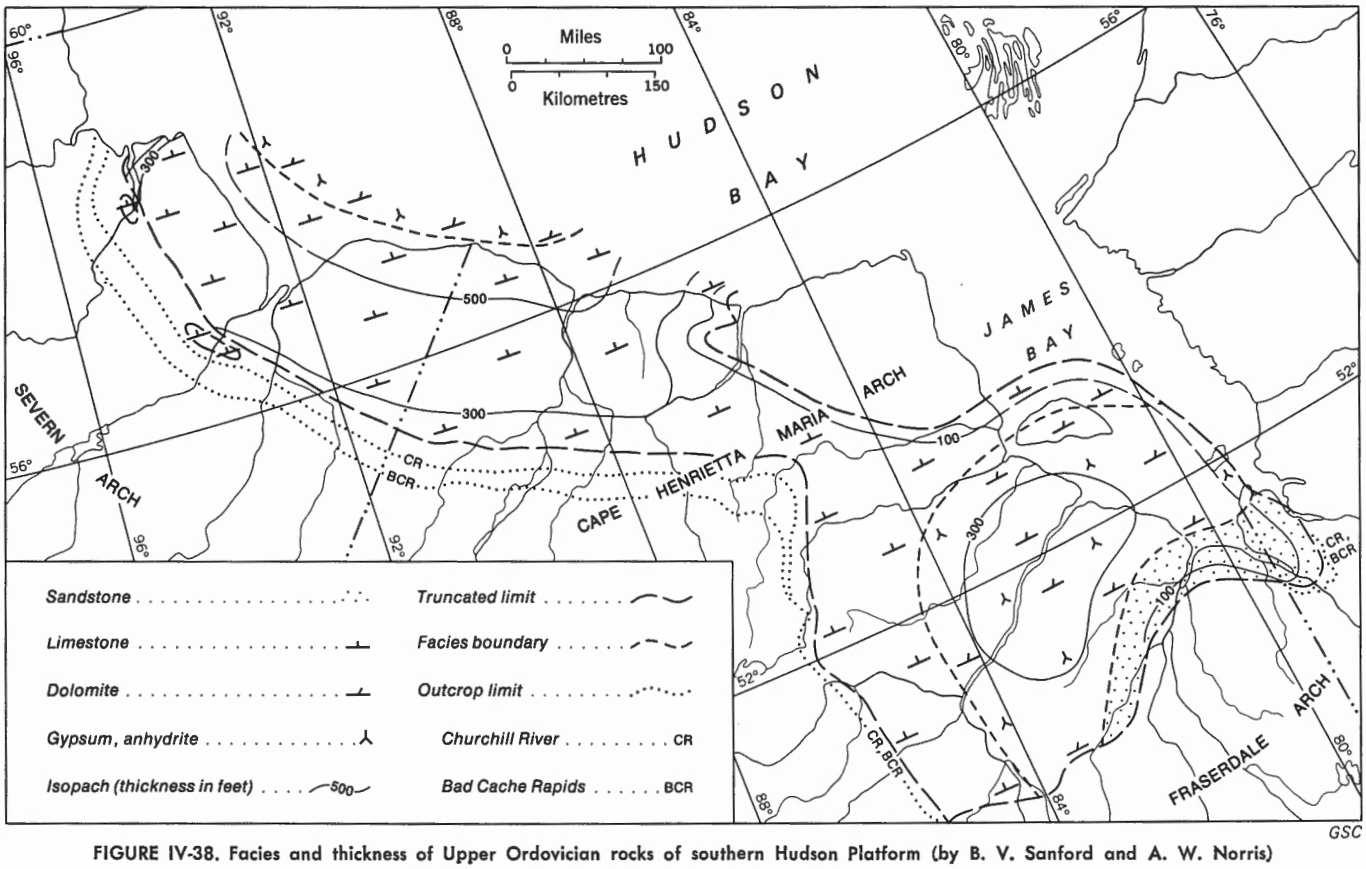


FIGURE IV-38. Facies and thickness of Upper Ordovician rocks of southern Hudson Platform (by B. V. Sanford and A. W. Norris)

The Churchill River Group is correlated with the Stony Mountain Formation of southern Manitoba (Nelson, 1964), and is coeval with the Queenston–Kagawong–Meaford Formations of southern Ontario. Equivalent strata are present at Lake St. John outlier, but are not represented in the Lake Timiskaming outlier. The latter region, accordingly, may have been land.

Silurian

Tectonic Summary

Although much of the southern Canadian Shield was inundated during the Alexandrian, Hudson Platform was presumably emergent. In the early Niagaran the sea transgressed into Williston Basin and onto Arctic and Hudson Platforms leaving thick limestone deposits (Fig. IV-39). Late Niagaran sedimentation on Hudson Platform was characterized by major reef development comparable to that of the Great Lakes region of Canada and the United States (Fig. IV-40). A fringing reef facies developed on the flanks of Hudson Bay and Moose River Basins, giving place to inter-reef facies in the more rapidly subsiding central parts of the basins. During the Late Silurian, Cayugan, Severn and Fraserdale Arches again became emergent, providing clastic detritus for the Moose River and Hudson Bay Basins to the north (Fig. IV-41). The normal marine environments that prevailed during the preceding Niagaran presumably became restricted, resulting in the deposition of primary dolomites, gypsum, and anhydrite in association with sandstone, mudstone, and siltstone, the latter forming the dominant facies.

Niagaran

Of the Niagaran formations the Severn River is a relatively uniform rock unit throughout Hudson Bay and Moose River Basins, whereas the succeeding Ekwan River and Attawapiskat Formations are closely related lithologically and form separate map units only on the margins of the basins. The Severn River Formation disconformably overlies the Upper Ordovician Churchill River Group, except in the vicinity of Churchill, Manitoba, and on Cape Henrietta Maria Arch, where it overlaps the Ordovician to rest directly upon the Precambrian. The formation rises to the surface along the western margin of the Hudson Bay Lowland and forms the youngest Paleozoic rocks over the Cape Henrietta Maria Arch (Fig. IV-38). It is in fault contact with the Precambrian on the southern margin of the Moose River Basin, and outcrops on Coats and Southampton Islands in the Hudson Bay. Maximum thickness of the formation in Moose River Basin is 150 to 200 feet, and 700 feet along the southern margin of Hudson Bay Basin. The Severn River is composed of

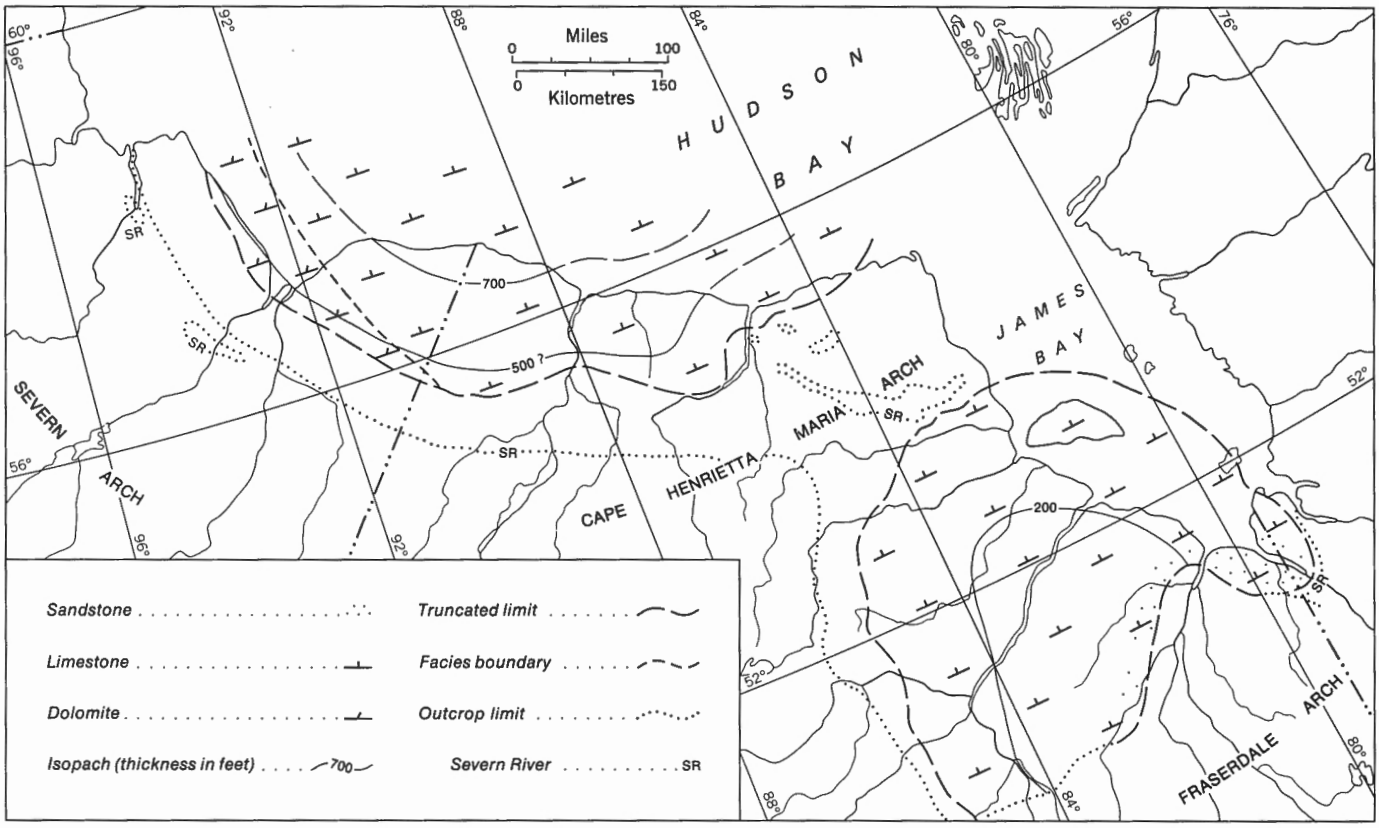


FIGURE IV-39. Facies and thickness of Silurian (early Niagaran) rocks of southern Hudson Platform (by B. V. Sanford and A. W. Norris).

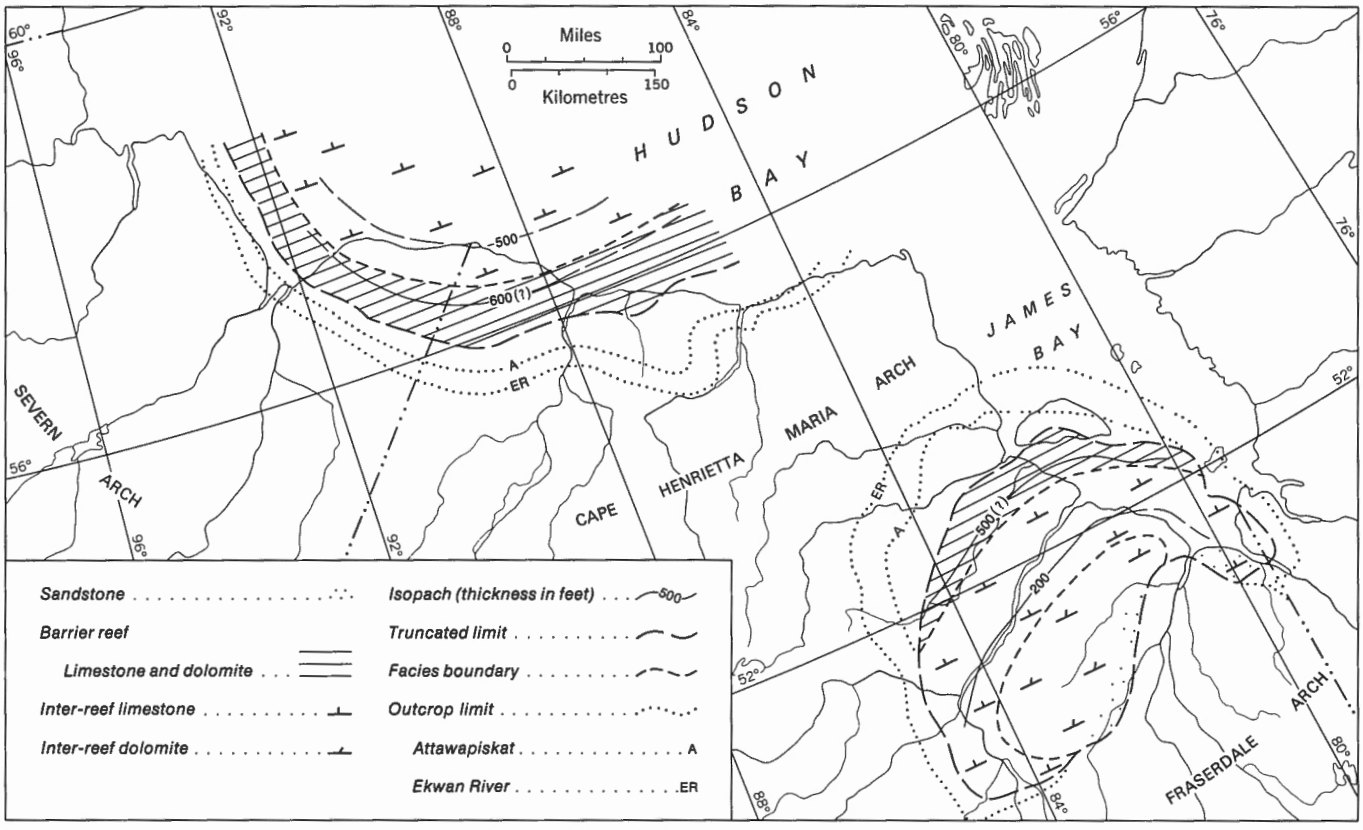


FIGURE IV-40. Facies and thickness of Silurian (late Niagaran (?)) rocks of southern Hudson Platform (by B. V. Sanford and A. W. Norris).

brown and tan, finely crystalline to aphanitic limestone and dolomite and mottled limestone with coarse, bioclastic, and fragmental limestone, and flat pebble-conglomerate locally present. Although generally thin bedded, the formation may be massive, swelling into thick biostromal lenses. The Severn River Formation correlates approximately with the upper part of the Wabi and Cataract Groups of Lake Timiskaming outlier and southwestern Ontario, respectively, and the lower part of the Interlake Group of southern Manitoba.

The Ekwan River Formation overlies the Severn probably disconformably, and, where combined with the succeeding and, in part, laterally equivalent Attawapiskat fringing reef facies, probably reaches thicknesses of 500 to 700 feet along the margins of the Hudson Bay and Moose River Basins (Fig. IV-40). Because of the rapid change to inter-reef facies the Ekwan River–Attawapiskat interval thins to 200 feet or less in central Moose River Basin, and perhaps to a comparable thickness in Hudson Bay Basin. The Ekwan River Formation is composed of light grey, tan, and brown, locally bituminous, fine- to medium-crystalline limestone and dolomite. It is thin to thick bedded and locally biostromal. The formation contains varying amounts of bioclastic, mainly crinoidal detritus and fragmental limestone, and in southern Moose River Basin, particularly in Quebec, sandstone interbeds and lenses. The Ekwan River Formation is tentatively correlated with the Thornloe Formation of Lake Timiskaming outlier, the Fossil Hill and Amabel Formations of southwestern Ontario, and part of the Interlake Group of southern Manitoba.

The Attawapiskat Formation conformably succeeds the Ekwan River, reaches its maximum development bordering the Cape Henrietta Maria Arch, and changes basinwards to inter-reefal carbonate rocks. The Attawapiskat is a reef complex composed of small bioherms that consist of thick structureless masses of yellowish tan and brown, vuggy, cavernous limestone, weathering ash grey. Flanking the bioherms are thick beds of carbonate and bioclastic detritus that dip at steep angles away from the reef cores. These beds become finer and thinly bedded where they grade into the inter-reef facies. Succeeding the biohermal facies and forming a dominant part of the Attawapiskat are thick biostromal beds of light grey or tan and brown, massive, vuggy limestones and dolomites. At numerous localities these strata form domes, apparently draped over buried bioherm reefs lower in the Attawapiskat. The bioherms and associated carbonate rocks of the Attawapiskat are closely similar in lithological character to the barrier reef facies of the late Middle Silurian Guelph

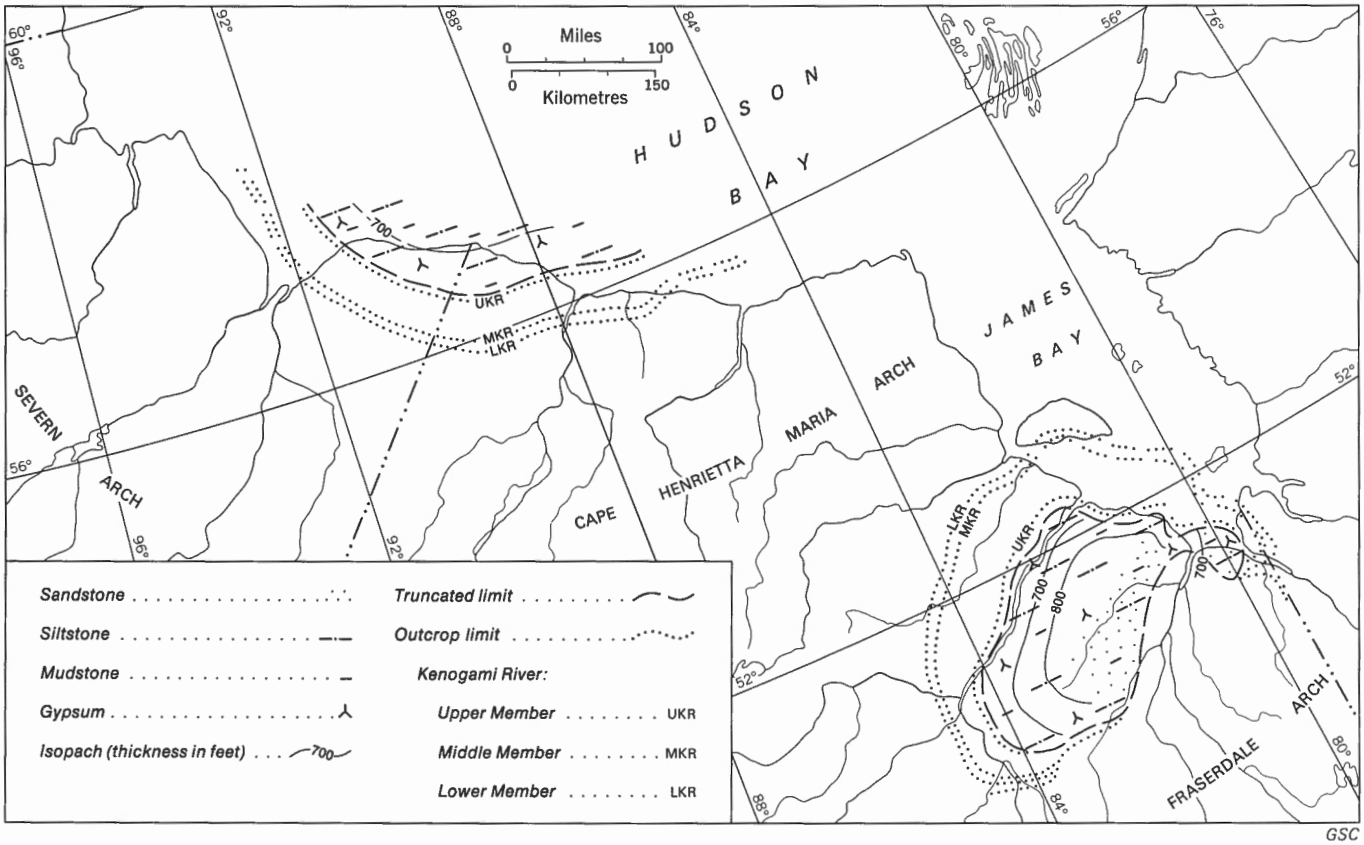


FIGURE IV-41. Facies and thickness of Upper Silurian rocks of southern Hudson Platform (by B. V. Sanford and A. W. Norris).

Formation of southwestern Ontario, and are tentatively correlated with that formation. The Attawapiskat is also probably equivalent to the Chemahawin Member of the Cedar Lake Formation of southern Manitoba.

Cayugan

The Kenogami River Formation conformably succeeds the Attawapiskat Formation, or the partly equivalent Ekwan River. The formation has a known thickness of 835 feet in eastern Moose River Basin, and its thickness in Hudson Bay Basin is presumably comparable (Fig. IV-41). It is divisible into three members. The lower member is a uniform sequence, 74 to 174 feet thick, of thin- to thick-bedded, brown, finely crystalline dolomite, containing thin interbeds of white gypsum or anhydrite. The middle member consists of gypsiferous, red and green siltstone, mudstone, sandstone, minor argillaceous dolomite, and honeycomb limestone, 475 to 552 feet thick. The upper member is a uniform succession of brown, tan, and cream coloured, microsucrosic oölitic dolomite, and dolomite breccia, 37 to 109 feet thick. The Kenogami River Formation is considered coeval with the lithologically similar Salina and Bass Islands Formations of southwestern Ontario, and the Ashern Formation of southern Manitoba.

Devonian

Tectonic Summary

During the early Early Devonian (Gedinnian–Siegenian), Fraserdale Arch was epeirogenically uplifted (Fig. IV-42). Uplift of the Precambrian basement took place on a fault along the southern margin of the Moose River Basin to bring Archean plutonic rocks into juxtaposition with Ordovician and Silurian strata. Thus a highland area extended along the southern margin of the basin as the initial late Early Devonian sea trangressed onto Hudson Platform. Rocks of Emsian age deposited at the foot of the fault scarp consist of non-marine sandstone, siltstone, shale, and conglomerate. These give place laterally to normal marine limestones and dolomites in the central part of Moose River Basin, and marine carbonate rocks form the entire sequence in the Hudson Bay Basin. The local terrigenous sequence was succeeded and overlapped by marine limestones, the latter presumably forming a continuous deposit across the Canadian Shield extending to Michigan and Allegheny Basins in the south. At the close of Early Devonian time, the sea retreated from Hudson Platform, initiating a brief interval of erosion.

The Middle Devonian is represented by normal marine limestone, evaporites, and minor shale. The faunas are

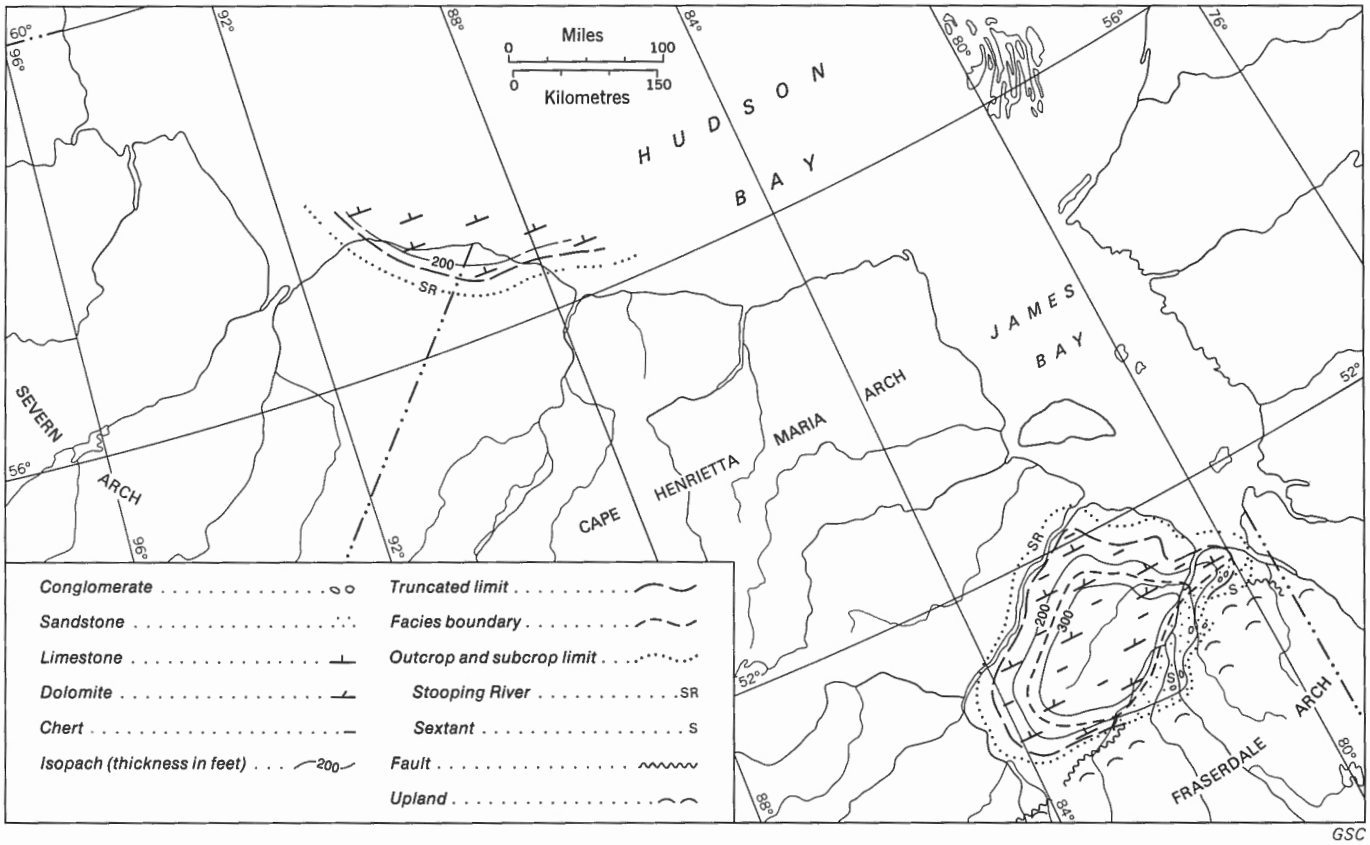


FIGURE IV-42. Facies and thickness of Lower Devonian rocks of southern Hudson Platform (by B. V. Sanford and A. W. Norris).

similar to those of the Allegheny and Michigan Basins and indicate three intervals of submergence and interconnection: early Eifelian, late Eifelian, and mid-Givetian. The latter intervals show some faunal similarity to parts of the Middle Devonian of Williston Basin in southern Manitoba. During the early Givetian hiatus evaporite deposits were removed by solution along the southern margin of Moose River Basin resulting in collapse and brecciation of interbedded and overlying carbonates. This interval also corresponds with the development of the Mackinac Straits breccia of northern Michigan, which suggests that widespread regional epeirogenic uplift of the eastern craton took place.

Upper Devonian rocks of Hudson Platform are composed of black shales similar to those in the Allegheny and Michigan Basins. They were presumably derived from the same provenance to the southeast in the Appalachian Geosyncline uplifted during the Acadian Orogeny and transported far across the southern Canadian Shield.

Lower Devonian

Two formations of Early Devonian age are recognized in the Hudson Platform: the non-marine Sextant Formation, confined to the southern part of Moose River Basin, and its marine equivalent, the Stooping River Formation. Where it overlies Archean rocks of the basement, the Sextant consists of as much as 175 feet of reddish pink and grey, mottled, quartz feldspar conglomerate, coarse feldspathic sandstone, and red or grey, locally carbonaceous, siltstone and shale. In Quebec, two outliers of the Sextant rest unconformably upon undivided Upper Ordovician strata, and the Middle Silurian Ekwan River Formation. The Sextant Formation grades basinward to thin-bedded, nodular, cream, tan and grey, finely crystalline cherty limestone and dolomite of the Stooping River Formation, about 300 feet in thickness. In central Moose River Basin and in Hudson Bay Basin, the Stooping River disconformably succeeds the upper member of the Kenogami River Formation. In southern Moose River Basin the Stooping River Formation transgressively overlaps the Sextant to rest directly upon the Precambrian. Plant remains are locally abundant in the Sextant and indicate a continental environment of deposition. The spores have been dated as late Lower Devonian (Emsian). The marine shelly fauna in the Stooping River Formation indicates a correlation with the Bois Blanc Formation of southwestern Ontario and Michigan, and is also dated as Emsian.

Middle Devonian

The Middle Devonian Kwataboahegan, Moose River, Murray Island, and Williams Island Formations have a combined thickness of about 700 feet and are composed predominantly of marine limestones with minor shale and evaporites. Although these formations are well developed in Moose River Basin, only the Kwataboahegan is present on the mainland part of Hudson Bay Basin. A more complete Middle Devonian sequence is presumably present offshore, beneath Hudson Bay. The Kwataboahegan Formation disconformably succeeds the Stooping River. It consists of tan and light to dark chocolate brown, bituminous, coral limestone. The beds are commonly thin to medium bedded but became thick bedded to massive and locally biostromal along the extreme southern margin of Moose River Basin. On the basis of its rich coral fauna the Kwataboahegan is correlated with the Amherstburg Formation and the Edgecliff Member of the Onondaga Formation of southwestern Ontario.

Succeeding the Kwataboahegan with fairly abrupt but conformable contact are the carbonate and evaporite deposits of the Moose River Formation. In central Moose River Basin these are composed of light to medium brown and tan, aphanitic to microsucrosic limestones and dolomites containing thick beds of white gypsum, with a combined thickness of more than 100 feet. Along the southern margin of the basin, gypsum is sporadic, having been removed in large part by solution, which left the limestones and dolomites contorted and brecciated. Although relict bedding can be observed here and there in the breccia, the formation is largely a jumbled mass of carbonate fragments as much as several feet in diameter in a calcisiltite and micrite matrix.

The Murray Island Formation overlies the Moose River disconformably. It consists of about 20 feet of thick-bedded, light brown, crinoidal limestone grading upward to thin-bedded, finely crystalline to aphanitic limestone. The formation has a fracture pattern that apparently formed from slight subsidence as a result of leaching or compaction of the underlying evaporites. The Murray Island contains a brachiopod fauna that suggests correlation with the Elm Point Formation of southern Manitoba. It is also tentatively correlated with the lithologically similar Dundee and Rogers City Formations of Michigan Basin considered late Eifelian and early Givetian.

The Williams Island Formation, 150 to 200 feet in thickness, succeeds the Murray Island Formation, possibly with a disconformable contact. The lower strata are blue-grey and red shales locally containing thin interbeds of gypsum and coarse brown crinoidal limestone. These grade upward to tan or light brown and light grey, finely crystalline to granular, high calcium limestone, containing traces of gypsum and minor limestone breccia. The coral and brachiopod fauna of the Williams Island Formation in Moose River Basin indicates correlation with the Hamilton Formation of Givetian age of southwestern Ontario. Fossils characteristic of the Dawson Bay Formation of Manitoba, mixed with Williams Island forms, occur in rubble near the mouths of several rivers flowing into Hudson Bay near the Manitoba–Ontario boundary, and were apparently scoured by glacial action from rock outcrops located offshore.

Upper Devonian

The Long Rapids Formation, 285 feet thick, forms the youngest Paleozoic strata of Hudson Platform. They

rest with an abrupt and possibly disconformable contact upon the Williams Island Formation and are composed of sparsely fossiliferous, dark brown and black, carbonaceous, fissile shales with thin interbeds of soft greyish green shale and hard grey-green dolomite. It is tentatively correlated with the lithologically similar Upper Devonian Kettle Point Formation of southwestern Ontario.

MESOZOIC

Tectonic Summary

After the widespread submergence of Hudson Platform and the Canadian Shield during the early and middle Paleozoic, the Shield was uplifted and stripped of most of its sedimentary cover. This epeiorogeny may have occurred mainly during the late Paleozoic and early Mesozoic as non-marine clastic sequences of late Jurassic and early Cretaceous ages occur in the Williston, Michigan, and Moose River Basins, and on the Atlantic Continental Shelf. In addition, alkalic magmatic activity, the Monteregian Intrusions, occurred in western Quebec Basin. They are dated as late Jurassic to early Lower Cretaceous and may be associated with faulting. Lamprophyre and kimberlitic dykes and sills were intruded into Devonian rocks on Abitibi River at Sextant and Coral Rapids near the southern margin of Moose River Basin.

Lower Cretaceous

The Mattagami Formation is recognized only in the Moose River Basin where it unconformably overlies bevelled edges of various formations from Upper Silurian to Upper Devonian age and rests locally on the Precambrian. It is about 170 feet thick. The lower part consists of dark grey to black fire clays, and brownish to yellowish micaceous sands with fragments of carbonized plant stems and thick seams of lignite, succeeded by a sequence of varicoloured plastic clay, and white quartz sand and clay. Fossil plants and spores have been dated as Upper Jurassic or Lower Cretaceous, with a preference for the latter age. Drilling has delineated about 10 million tons of coal at or close to surface and substantial reserves beneath greater thicknesses of overburden.

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V. Economic Minerals of the Canadian Shield

Introduction 152

Structural Provinces and Metallogenic Associations 154

Metallogenic Evolution of the Canadian Shield 156

Radioactive Elements 163

Ferride Metals 169

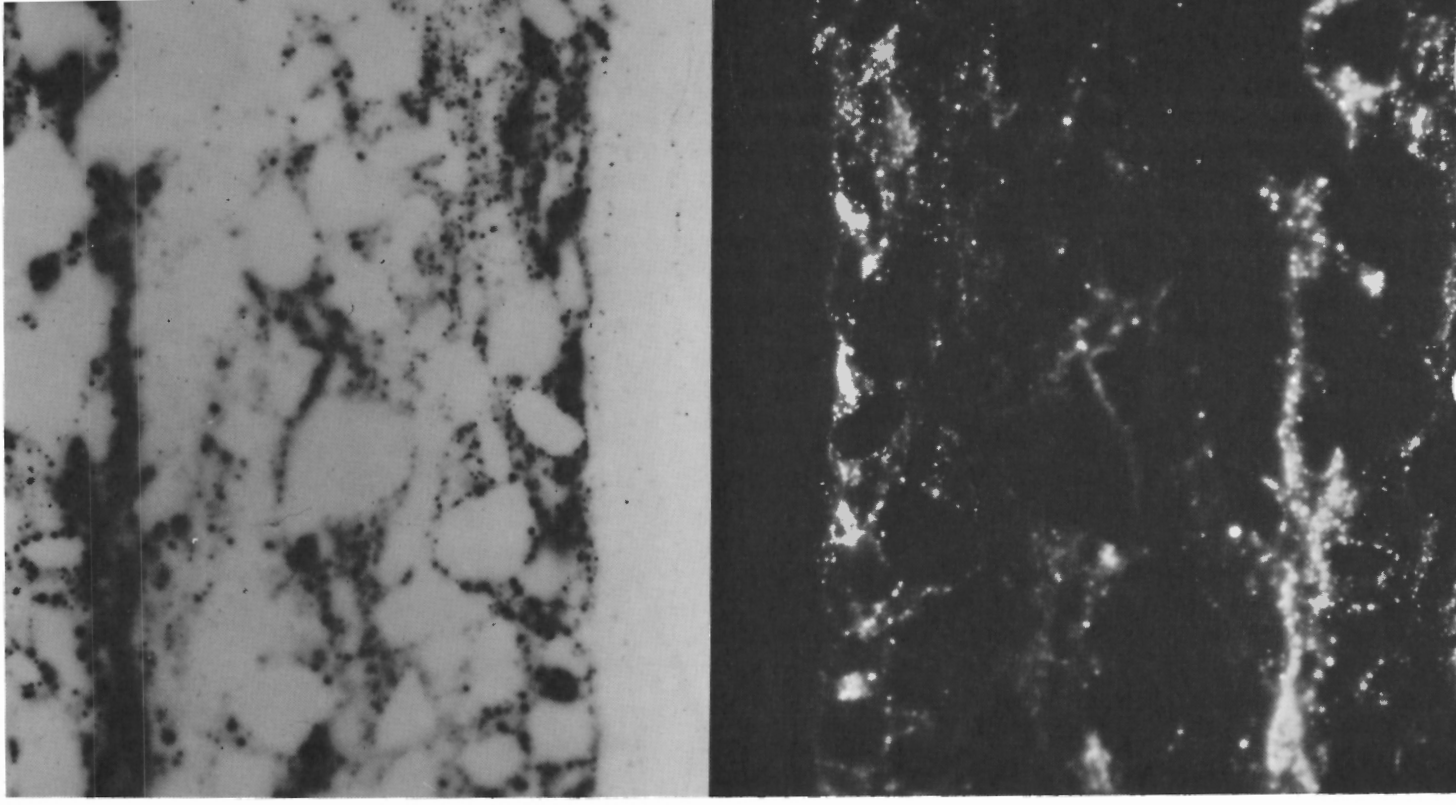
Base Metals 185

Precious Metals 200

Minor Metals 217

Industrial Minerals 220

Selected References 223



INTRODUCTION

A wide range of mineral commodities is produced from the Canadian Shield including precious metals, base metals, ferrous and ferro-alloy metals, nuclear metals, and some non-metals, notably insulants, and chemical and metallurgical minerals (Table V-1). The ten most valuable mineral commodities produced in the Canadian Shield in 1966 in order of decreasing value were: iron ore, nickel, copper, zinc, gold, uranium, platinum group metals, silver, magnesium, and lead. These ten commodities account for 42 per cent of the total mineral production including mineral fuels in Canada in 1966.

Iron ore is produced from banded iron-formations in Quebec and Labrador and also northern Ontario. Nickel production is from large nickel–copper deposits at Sudbury, Ontario, and the Thompson–Moak Lake and Lynn Lake areas of Manitoba. Formerly, other production came from Rankin Inlet on the west shore of Hudson Bay. Copper is derived from the nickel–copper deposits at Sudbury and Lynn Lake, from large stratiform copper–zinc deposits in northern Quebec, Ontario, and Manitoba. Gold is obtained both as a byproduct of the base metal operations and from lode deposits in Ontario, Quebec, Manitoba, and the Northwest Territories. Uranium production is from large uraniferous quartz–pebble conglomerates in the Elliot Lake area of Ontario. Formerly, other production came from vein and pegmatitic deposits

TABLE V-1 *Production of the ten main mineral commodities from the Canadian Shield in 1966*

	Quantity ($\times 10^6$)	Value ($\times 10^6$ Dollars)	Percentage of Total Canadian Production (by quantity)
Iron ore	38 (ton)	406	94.0
Nickel	445 (lb)	375	99.5
Copper	738.7 (lb)	331.4	73.0
Zinc	877.2 (lb)	132.4	45.5
Gold	3.3 (troy oz)	119.1	95.1
Uranium (U_3O_8)	7.9 (lb)	54.3	100.0
Platinum group metals	0.4 (troy oz)	32.3	100.0
Silver	21.4 (troy oz)	29.9	64.0
Magnesium	13.4 (lb)	4.2	100.0
Lead	9.6 (lb)	1.4	1.6
Total value: $1,486,000,000 (42% of total Canadian production)			

1 The Introduction and sections on Metallogenic Associations and Metallogenic Evolution were prepared by A. M. Goodwin; Copper–Zinc, Lead, and Uranium by A. H. Lang; Nickel, Platinum, and Chromium by J. A. Chamberlain; Iron by G. A. Gross; Gold by D. R. E. Whitmore; Silver and Cobalt by R. W. Boyle and A. G. Johnston; Titanium by E. R. Rose; Beryllium, Cesium, Lithium, Molybdenum, Niobium, Tantalum, Tin, and Tungsten by R. Mulligan. The remaining parts are from previous editions with revisions by the staff of the Mineral Deposits Section. When acknowledging, reference should be made to individual author and page number.



*See Page 793

V

Economic Minerals of the Canadian Shield

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Uraniferous conglomerate, Blind River, Ontario. Section on right under natural light shows rounded quartzite pebbles with grains of pyrite in matrix. Autoradiograph negative print forms centre section and a positive print the left section.

in southeastern Ontario, Saskatchewan, and the Northwest Territories. Silver is produced from silver–cobalt deposits in the Cobalt area, Ontario, as a byproduct of copper–zinc operations in northern Quebec, Ontario, and Manitoba, and from gold–silver deposits in Quebec, Ontario, Manitoba, and the Northwest Territories. Lead is a byproduct of copper–zinc production in Quebec, Ontario, and Manitoba. Titanium oxide comes from titaniferous magnetite deposits in Quebec.

Certain belts of rocks are particularly rich in mineral deposits. The most productive belt, which lies in Quebec and Ontario, contains the great copper–zinc, uranium, gold, nickel, and iron deposits of the Chibougamau, Matagami, Noranda, Timmins, Blind River, Michipicoten, Manitouwadge, and Steeprock Lake areas. Additional productive belts, the source of copper, zinc, nickel, uranium, gold, and silver, occur at Flin Flon, Lynn Lake, Lake Athabasca, Great Slave Lake, and Great Bear Lake. Major iron-bearing belts cross Ungava and Baffin Island.

The rates of production of mineral commodities from the Canadian Shield, and their individual value, fluctuate over the years with changing market demands and technology and the impact of new discoveries. Gold production has been in steady decline for 30 years in response to the fixed market price of gold bullion. Production of nickel and iron have increased steadily in response to strong demand. Uranium production, directly related to its use as a fuel, has fluctuated drastically in recent years. Silver production is destined to increase substantially due to the recent discovery of a single, unusually large copper–zinc–silver deposit at Kidd Creek, Ontario. Mineral production has come principally from the margins of the Shield (Map 1252A), particularly the southern parts, mainly as a matter of accessibility and transportation. The northern and central parts of the Shield, far removed from habitation, difficult to reach, and expensive to supply, have received correspondingly less exploration attention. Despite this, a substantial number of gold, nickel, and iron occurrences have been found, sufficient to suggest that many others await discovery. However, the paucity of significant copper–zinc and uranium discoveries to date in the northern parts of the Shield raises the possibility of a regional deficiency in these and other metals.

History of Exploration

Geological studies and mineral exploration have been inextricably linked in the Canadian Shield. In the early stages exploration travel across the Shield was by canoe and other shallow-draft boats on the navigable streams and lakes. Since the advent of air transportation, float-equipped planes have habitually used lakes and streams for landing and recently helicopters have been used to reach still less accessible terrain. Systematic searches for minerals may be said to have begun with the founder of the Geological Survey of Canada, Sir William Logan. Appointed "Provincial Geologist" in 1842, Logan's assignment was to make a geological survey of Canada and to organize ". . . a plan of investigation as may promise to

lead to the most speedy and economic development of the mineral resources of the country." Logan's early reports of progress and annual reports culminating in his classic "Geology of Canada, 1863" are storehouses of information on mineral occurrences in eastern Canada; included is reference to copper occurrences on the north shore of Lake Huron destined to become the Bruce mine, the first copper mine in Canada.

In 1852, Sterry Hunt identified ilmenite in the "titanic iron ore" at St. Urbain, Quebec. Alexander Murray, in 1846, first reported nickel at the Wallace mine near Whitefish Falls, Ontario, and, in 1856, described nickel in "magnetic trap" on Salter's meridian near the present site of the International Nickel Company Creighton mine. In 1866, H. G. Vennor reported the first occurrence of gold in Ontario at the Richardson mine north of Madoc. Between 1877 and 1895 Robert Bell and A. P. Low reported large iron deposits on the Belcher Islands of Hudson Bay and along the Koksoak River in northern Quebec and Labrador. G. C. Hoffman first reported niobium in Canada in 1882 in Maissoneuve township, Quebec. In 1900, J. M. Bell reported the copper–cobalt occurrences later found by Gilbert Labine to contain pitchblende at Great Bear Lake, Northwest Territories. In 1903, W. G. Miller identified silver in ore discovered by Fred LaRose at Cobalt, Ontario. In 1904, W. A. Parks reported gold in quartz veins at Kirkland Lake and Larder Lake, Ontario. E. H. Horne, in 1917, located mineralized rock at the Horne orebody, Noranda, Quebec. In 1934, H. V. Ellsworth identified pitchblende in vein material from Beaverlodge, Saskatchewan. J. E. Thomson, in 1931, reported the occurrence of sulphide minerals near Lake Manitouwadge, Ontario. Franc Joubin located large uranium deposits by drilling at Elliot Lake, Ontario in 1953. In 1962–65 high-grade hematite iron ore was explored in Baffin Island by Murray Watts and associates.

Exploration Techniques

The traditional methods of mineral exploration involving geological mapping and conventional prospecting with or without the aid of simple instruments such as dip-needle and gold pan have gradually evolved. Modern, airborne, geophysical surveys employ magnetic, electromagnetic, radioactive, and remote sensing devices. Ground geophysical surveys produce electrical, magnetic, radioactive, and seismic data. Equally sophisticated geochemical programs comprising stream, soil, and bedrock sampling, and wall-rock alteration studies are based on complex instruments and analytical procedures.

Present exploration practices include intensive searches for mineral deposits in large areas selected on the basis of known geological relationships. Increasingly, mineral deposits are viewed as integral parts of complex geological environments and geological theory is, therefore, an essential guide to successful exploration. Dramatic evidence of the impact of technology upon exploration is provided by a comparative summary of discovery methods in recent decades (Lang, 1967). In the decade 1946–55, of the seventy-seven mines brought into production, fifty-three (69 per cent) are attributed to conventional prospecting; seventeen (22 per cent) to geological methods; and seven (9 per cent) to geophysical prospecting. In the succeeding decade, 1955–65, of the 175 new mines brought into production, eighty-seven (50 per cent) are attributed to conventional prospecting; forty-nine (28 per cent) to geological methods; twenty-eight (16 per cent) to geophysical methods; and eleven (6 per cent) to other methods. Geochemical prospecting was a supporting method in some of the discoveries in the 1955–65 decade. Lang concludes that conventional prospecting is still important, that geological and geophysical methods are about equally important, that geochemical prospecting is increasing in value, and that combined programs of mineral exploration are increasingly effective.

STRUCTURAL PROVINCES AND METALLOGENIC ASSOCIATIONS

The imprint of the main Precambrian orogenies that divide the Shield into the seven structural provinces also gives rise to distinctive metallogenic associations. The oldest orogeny, the Kenoran, dated at about 2,500 m.y., is the last important interval of folding, metamorphism, and intrusion in Superior and Slave Provinces. The Hudsonian Orogeny, dated at about 1,700 m.y. occurred throughout Churchill, Bear, and Southern Provinces. The Nain Province was affected by the Elsonian Orogeny. The Grenvillian Orogeny, dated at about 950 m.y., affected Grenville Province. The Canadian Shield contains many distinctive lithic communities of which the mineral occurrences themselves are a very important part. Most mineral deposits are considered to express the geological environments prevailing during the periods of host rock accumulation, consolidation, and deformation. Accordingly, an understanding of the original geological environment commonly provides metallogenic insight and contributes substantially to mineral exploration.

Archean lithic assemblages contain great thicknesses of volcanic rocks, predominantly basalt–andesite–rhyolite associations typical of orogenically active continental margins including island arcs and of flysch-type greywacke–conglomerate–breccia–shale assemblages also produced in orogenically unstable environments. Archean granitic rocks include a high proportion of layered migmatite and gneiss attributed in large part to metasomatic alteration of pre-existing sedimentary rocks. Archean mineral deposits are likewise products of orogenic environments. They include stratiform polymetallic sulphide,

gold–silver, and nickel sulphide deposits together with sedimentary iron-formations. The mineral deposits are associated preferentially with igneous rocks, especially volcanic rocks.

Proterozoic lithic assemblages of the Shield feature platformal or miogeosynclinal deposits including great thicknesses of mature sandstone–quartzite, dolomite and limestone, shale, granular iron-formation, and tholeiitic flood basalts. In some belts, geosynclinal-type basalt and associated greywacke and shale occur together with large mafic and ultramafic intrusions, the latter commonly flat lying and occasionally layered. Alkalic intrusions, including carbonatites, are widely though sparsely distributed. Proterozoic mineral deposits are, accordingly, mainly products of comparatively stable, platformal crustal environments. Included are extensive placer uranium deposits, granular iron-formations, carbonate-enclosed lead–zinc deposits, complex silver–cobalt and nickel–copper sulphide and sulpharsenide deposits in the mafic intrusions, titaniferous magnetite deposits, and niobium-bearing carbonatites.

Superior Province

Superior Province is characterized by easterly trending belts of Archean supracrustal rocks separated by large areas of granitic rocks. The supracrustal rocks in many of the belts are of low metamorphic grade. Mafic intrusions and rare post-orogenic alkalic intrusions are present locally.

Numerous stratiform polymetallic sulphide deposits such as the copper–zinc–silver–gold affiliation are associated with the volcanic rocks. The deposits occur either in the volcanic rocks and associated mafic intrusions or in nearby sediments. Notable examples are the deposits at Chibougamau, Matagami, and Noranda in Quebec, and at Timmins, Timagami, and Manitouwadge in Ontario.

Banded iron-formation of the Algoma type is common in the volcanic-sedimentary belts. The larger, richer parts may constitute orebodies. Large reserves of low-grade concentrating-type, magnetite–hematite ores are present at Kirkland Lake, Timagami, Bruce Lake, North Spirit Lake, and Lake St. Joseph, all in northern Ontario. Siderite and pyrite deposits are present locally as are secondary hematite–limonite deposits.

Gold lode deposits are common in the volcanic-rich belts, many at the contact with small granitic stocks or in shear zones. The typical quartz–carbonate veins contain sulphide minerals, tellurides, and free gold. Native silver is also generally present. Some gold deposits occur in banded iron-formation. At Kirkland Lake the gold deposits are associated with syenite stocks. In addition, much gold is found in the large polymetallic sulphide deposits. Large native silver orebodies are associated with mafic sills of Proterozoic age in Cobalt Subprovince and with the large volcanic-associated polymetallic sulphide deposits in Ontario and Quebec. Molybdenum-, lithium- and beryllium-bearing deposits occur in pegmatitic and granitic rocks in Quebec and Ontario.

Slave Province

Slave Province contains belts of Archean volcanic and sedimentary rocks that generally trend north and are separated by extensive areas of granitic rocks. The most common metal in Slave Province is gold. It is widely distributed in a variety of rocks; more than a thousand occurrences have been reported (Lang, 1962). The larger, producing deposits are associated with volcanic rocks and intrusions in the southern part of the province. A few occurrences of copper, lead, cobalt, nickel, and native silver have been found also in the south. Pegmatites with small amounts of lithium, tungsten, beryllium, niobium, tantalum, and tin are known in the eastern part of the province. As known, significant deposits of the base metals, uranium and iron, are rare to absent.

Bear Province

This province is underlain by Proterozoic sedimentary and volcanic rocks, in part deformed by the Hudsonian Orogeny, and in part later and flat lying. The characteristic metal of Bear Province is uranium. Hundreds of occurrences of pitchblende, mainly in veins, are known over a large part of the province. East of Great Bear Lake small deposits contain native silver, cobalt, nickel, copper, and gold. In the northern part of the province basaltic lava flows are mineralized with copper.

Churchill Province

Churchill Province, the largest in the Canadian Shield, contains a great variety of rocks of both Archean and Proterozoic ages. Archean volcanic-rich belts contain mineral associations similar to those of Superior Province. Polymetallic sulphide deposits and occurrences of the copper–zinc–gold–silver affiliation occur in the Flin Flon belt of Manitoba and Saskatchewan and the Ennadai belt of the Northwest Territories. Nickel–copper sulphide ores are associated with mafic to ultramafic intrusions in the Lynn Lake and Thompson areas of Manitoba and at Rankin Inlet and Ferguson Lake in the Northwest Territories. Numerous gold lode occurrences are present in the Flin Flon belt in Manitoba, in the area north of LaRonge, Saskatchewan, and in the eastern part of the Ennadai belt. Most occurrences consist of gold-bearing quartz veins in sheared volcanic and sedimentary rocks in the vicinity of porphyry granite intrusions.

Major iron deposits are present in the iron-formations of the Labrador Geosyncline. The iron ores take the form of direct-shipping hematite ores in the central part of the belt and of concentrating-type magnetite–hematite ores in the southern part. Low-grade iron deposits are present in the sedimentary sequences on Belcher Islands and those adjacent to the mainland of Hudson Bay. Large high-grade hematite iron deposits lie in metasedi-

mentary strata on Baffin Island. Magnetite-bearing iron-formations of possible commercial grade and size are present in the Ennadai belt. Smaller bands of iron-formation, including iron sulphide concentrations, are present in some narrow belts of metamorphosed volcanic and sedimentary rocks in northern District of Keewatin and on southern Baffin Island.

Uranium deposits occur in sedimentary and granitic rocks in the Beaverlodge area bordering Lake Athabasca, Saskatchewan. Pitchblende is present in brecciated granitized material and as calcitic vein-filling. At the Gunnar mine carbonatized albite syenite contains disseminated pitchblende and uranophane. Uranium mineralization also occurs in Aphebian sediments near Lake Mistassini in northern Quebec.

Asbestos occurrences are associated with mafic to ultramafic bodies in the Cape Smith belt in northern Quebec. Crocidolite showings have been found in Aphebian iron-bearing rocks near Knob Lake, Quebec.

Grenville Province

Grenville Province includes a great mixture of rocks of the various Precambrian eras. The Grenvillian Orogeny is the youngest of the major Precambrian orogenies that affected them. The mineral associations are correspondingly varied. Large magnetite–hematite deposits occur in highly metamorphosed iron-formation that forms the southern extension of the Aphebian rocks of Labrador Geosyncline. Smaller titaniferous magnetite skarn deposits border mafic intrusions in the southern part of the province. Small, titanium-free magnetite deposits occur in limestone in southeast Ontario and Quebec. Large deposits of ilmenite, the principal ore of titanium, occur as magmatic segregations in the Elsonian anorthosites of southern Quebec as at Allard Lake and St. Urbain. Niobium-bearing deposits are found in carbonatites near Lake Nipissing, Ontario, and Oka, Quebec. Uraniferous pegmatites occur in metamorphosed gneiss at Bancroft, Ontario. Small deposits of gold, lead–zinc, and molybdenum as well as the non-metallic nepheline syenite and the magnesium ores dolomite and brucite are present locally.

Southern Province

Although one of the smaller structural provinces of the Shield, Southern Province of Ontario has yielded great mineral wealth. The Sudbury nickel deposits occur in a thick sill of norite, quartz diorite, and micropegmatite emplaced at the unconformity between the Aphebian strata and Archean crystalline rocks. In addition to nickel, large quantities of copper and smaller amounts of gold, silver, and platinum group metals are present. Huge tonnages of uranium ore are contained in the quartz–pebble conglomerates of the lower part of the Aphebian Bruce Group in the Blind River area. The ore minerals, brannerite, uraninite, pitchblende, and monazite, are considered to represent placer deposits filling erosional depressions in the basement rocks. In the Port Arthur area, silver deposits lie in Animikie sediments at the contact with Helikian mafic sills and dykes. Small lead–zinc deposits occur in Helikian red beds. The Gunflint iron-formation in the Thunder Bay district contains large submarginal iron deposits.

METALLOGENIC EVOLUTION OF THE CANADIAN SHIELD

Studies of geological and metallogenic relations in the Canadian Shield have led to the concept of crustal evolution by growth of protocontinents (Goodwin, 1968). According to this concept, the main facets of Precambrian crustal evolution were from thin unstable protocontinents to a thick stable continental mass, representing pre-geosynclinal and early geosynclinal stages, respectively. The sites of supracrustal accumulation progressed from basins and trenches with random to sublinear distribution in Archean time, to stable platforms, broad basins, and comparatively long linear primitive geosynclines in Proterozoic time. Preferred easterly and northerly tectonic directions of the Archean changed to preferred northeasterly and northwesterly directions of the Proterozoic. Orogenic, flysch-type Archean sedimentation changed to stable and mature, shelf-type Proterozoic sedimentation. Volcanic activity changed from compositionally simple and sequential in Archean time to compositionally varied and non-sequential in Proterozoic time. Correspondingly, the pattern of mineralization changed from comparatively simple native gold and silver—simple base metal sulphide–iron–nickel assemblages of Archean time to the complex polymetallic sulphide and sulpharsenide–uraniferous–lead–iron–nickel mineral assemblages characteristic of the Proterozoic.

Archean Crust

Archean protocontinents are considered to be primary crustal units, comprising granitic nuclei or cratonic elements and linear, supracrustal orogenic elements. Incipient protocontinents feature a random to sublinear distribution of small cratons and orogens. Mature protocontinents feature a peripheral distribution of longer linear orogens about larger central cratons. The proto-

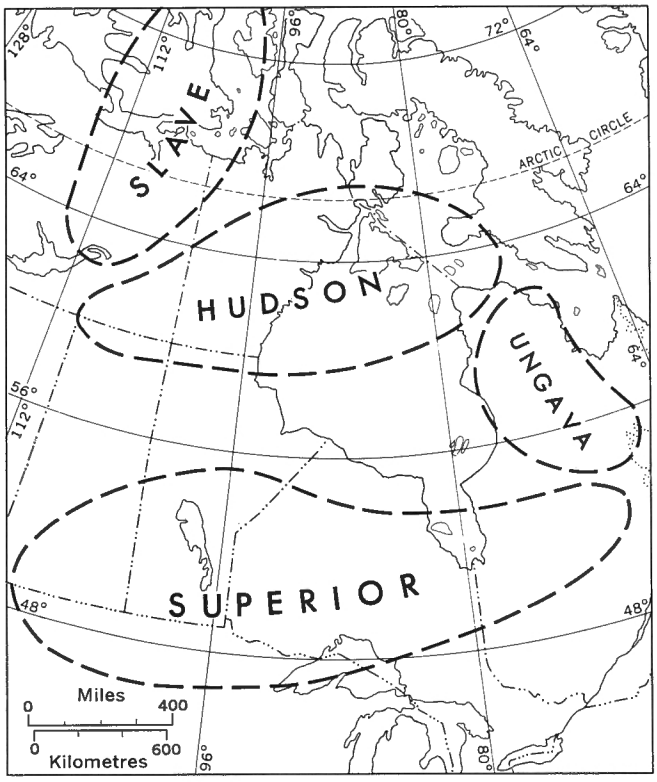


FIGURE V-1. Postulated Archean protocontinents (by A. M. Goodwin).

continents grew by surficial accretion of orogenic elements and by subcrustal addition of new, mantle-derived crustal material. With continued growth, the protocontinents, at various stages of development, merged to form a large stable continental mass. In the Canadian Shield, the remnants of at least four protocontinents in different stages of growth and development are evident: the Superior, Hudson, Slave, and Ungava Protocontinents (Fig. V-1).

Superior Protocontinent. This protocontinent contains a number of alternating volcanic- and sedimentary-rich orogenic belts together with irregular granitic masses. The southern part of the protocontinent contains three main east-trending volcanic-rich belts: the Uchi, Wabigoon, and Abitibi–Wawa belts (Fig. V-2). The northwestern part also contains three, though less well defined, volcanic-rich belts: the Windigo, Amisk, and LaRonge belts. The orogenic belts are distributed about a main granitic nucleus or craton that is partly exposed in northwestern Ontario and the adjacent part of Manitoba. The volcanic-rich belts in the southern part of the Superior Protocontinent, viewed from north to south, exhibit the following general progressions: 1) the belts increase in width; 2) individual areas of volcanic rocks within the belts increase in size; 3) the quantity of felsic volcanic rock increases; and 4) the volcanic assemblages as known increase in chemical diversity. The northwestern belts, though less well known, in part exhibit similar progressions to the north. These core-to-periphery progressions may also correspond to younger ages. Accordingly, a long and complex orogenic history involving many episodes of granite emplacement is postulated; however many local divergencies from these generalizations are undoubtedly present.

Hudson and Slave Protocontinents. Random to sublinear distribution of comparatively short discontinuous volcanic and sedimentary-rich belts is characteristic of the central and northern parts of the Canadian Shield. East to east-northeastern trends predominate in the Hudson Protocontinent and northern trends in the Slave Protocontinent. In the Hudson Protocontinent, on the basis of regional trends and location of predominantly granitic masses, cratonic elements are postulated for the areas immediately northwest of Hudson Bay and possibly northeast of Lake Athabasca. In the Slave Protocontinent, the northerly trend of the belts and distribution of granitic masses suggest the presence of a cratonic mass in the northern part, now mainly covered by post-Archean rocks.

Ungava Protocontinent. Very little is known about Archean volcanic and sedimentary rocks in the highly metamorphosed, structurally obscure part of the Shield east of Hudson Bay. Long linear belts of Proterozoic rocks girdle all but the southwestern part of a large area of predominantly granitic rocks in northern Quebec. On this basis the northerly trending Ungava Protocontinent is postulated. Other remnants of Archean protocontinents may be present in northeastern parts of the Shield such as Baffin Island and eastern Ungava.

Archean Mineralization

Nearly all known mineral occurrences in Archean rocks are located in the volcanic-rich or greenstone belts and most occurrences, whether they consist of sulphide masses, gold-bearing quartz veins, nickeliferous mafic intrusions, or iron sediments, are closely associated with the volcanic rocks themselves. Many mineral occurrences are associated with concentrations of felsic volcanic rocks, the sites of ancient volcanic centres. These closely associated metals and volcanic rocks are viewed as consanguineous igneous products (Goodwin, 1965). Base metal deposits occur preferentially in the outer, presumably younger parts of the Superior Protocontinent, whereas gold, iron, and nickel mineralization is more uniformly distributed in direct proportion to the quantity of volcanic rocks. Established mineral patterns in the Slave and Hudson Protocontinents resemble those in the inner or base-metal deficient part of the Superior Protocontinent. The metallogenic patterns of the Canadian Shield illustrated on Figures V-2 to V-5, are based on the present stage of exploration and hence are influenced by many variables (Goodwin, 1968b). In preparing these figures a balanced representation of occurrences rather than a demonstration of all occurrences and types of mineral deposits has been sought for each part of the Shield.

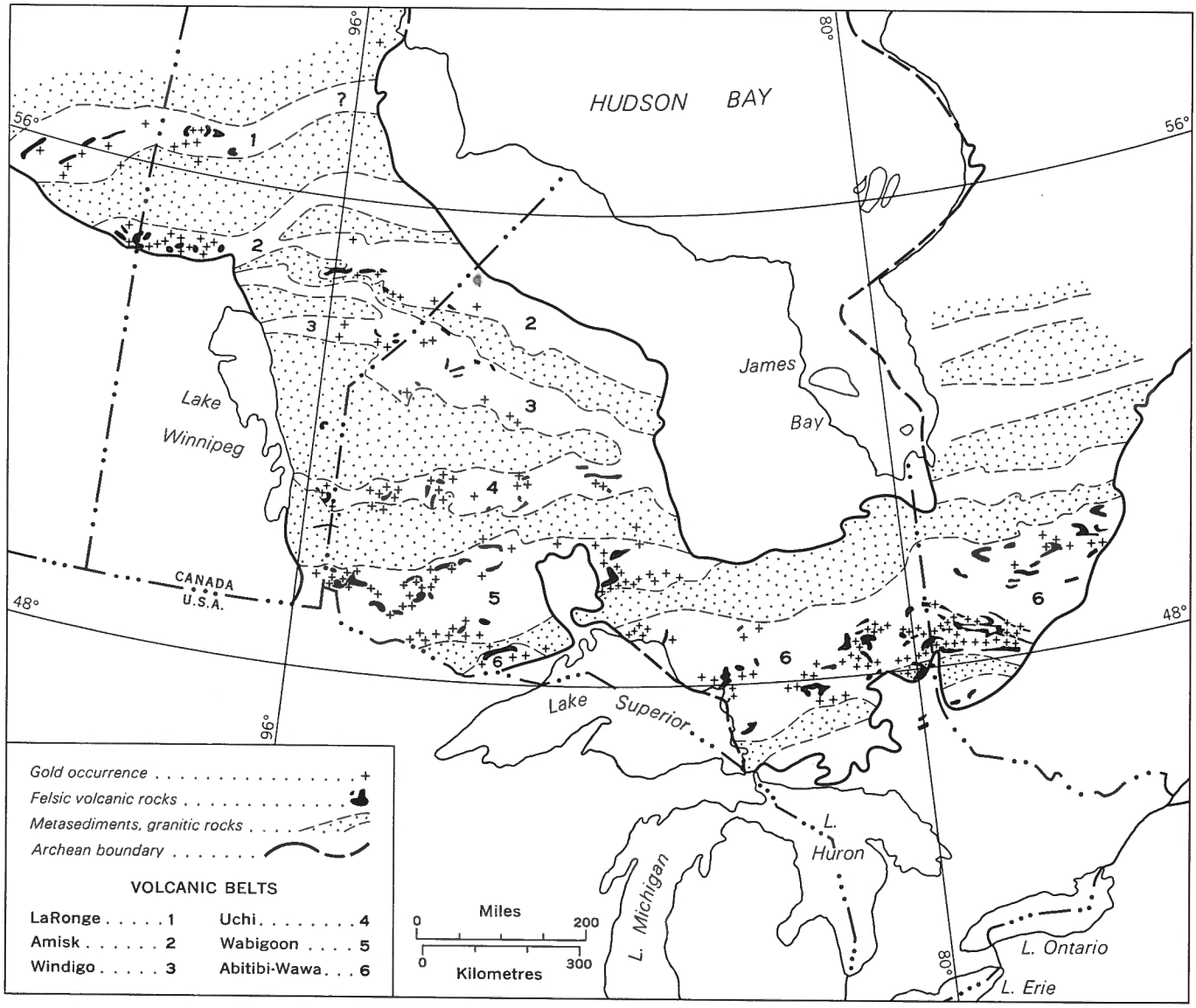


FIGURE V-2. Representative gold occurrences in the Superior Protocontinent (by A. M. Goodwin).

The distribution of one hundred and ninety-seven representative gold occurrences in the Superior Protocontinent is shown on Figure V-2. With few exceptions the occurrences lie in volcanic-rich belts and are preferentially associated with the felsic volcanic rocks. In general, the numbers of representative gold occurrences (Table V-2) in the six volcanic-rich belts are in direct proportion to the quantity of volcanic rocks in the belts indicating therefore that gold mineralization is fairly evenly distributed within the volcanic rocks of the protocontinent.

Of the thirty-three representative nickel and seventy-one representative iron occurrences (Fig. V-3) most occur in or near the volcanic-rich belts. Nickel distribution shows no preference for felsic volcanic rocks, but, on the contrary, is associated with mafic and ultramafic rocks. The ratios of nickel occurrences are in proportion to the relative sizes of the belts with the exception of the apparently nickel-deficient LaRonge belt. Most iron is in pyroclastic volcanic phases or nearby sediments. The ratios of iron occurrences are generally proportionate to the sizes of the three southern belts; however recognized iron occurrences are rare in the Amisk belt and absent in the LaRonge belt.

Most of the one hundred and eighty-three representative copper–zinc and seventeen representative zinc–lead occurrences (Fig. V-4) are associated with felsic volcanic rocks. A pronounced core-to-periphery increase within the protocontinent in the ratios of copper–zinc occurrences is apparent, an increase out of proportion to the relative sizes of the belts. Lead-bearing sulphide occurrences are almost restricted to the Abitibi–Wawa belt, the largest and presumably youngest volcanic-rich belt in the Superior Protocontinent. The inner part of the Superior Proto-

TABLE V-2 *Representative mineral occurrences and relative sizes of volcanic belts in Superior Protocontinent*

Volcanic Belt	Relative Size	Gold	Copper–Zinc	Zinc–Lead	Nickel	Iron	Total
a) Total Number, Superior Protocontinent							
LaRonge[1]	1.5	14	19	3	2	0	38
Amisk	2	20	30	1	7	3	61
Windigo	1	10	2	0	3	5	20
Uchi	1	21	5	0	4	13	43
Wabigoon	2	50	17	2	7	20	96
Abitibi–Wawa[2]	4	82	110	11	10	30	243
Total		197	183	17	33	71	501
Per cent		39	36	4	7	14	100
b) Total Number, Nucleus (Uchi and Windigo Belts)							
Windigo	1	10	2	0	3	5	20
Uchi	1	21	5	0	4	13	43
Total		31	7	0	7	18	63
Per cent		49	11	0	11	29	100
c) Total Number, Annulus (LaRonge, Amisk, Wabigoon and Abitibi–Wawa Belts)							
LaRonge	1.5	14	19	3	2	0	38
Amisk	2	20	30	1	7	3	61
Wabigoon	2	50	17	2	7	20	96
Abitibi–Wawa	4	82	110	11	10	30	243
Total		166	176	17	26	53	438
Per cent		38	40	4	6	12	100

1 Including Meades Lake area to the north
2 Including Ville Marie area to the south

continent, comprising the Uchi and Windigo belts, contains proportionately more gold, iron, and nickel and far less copper–zinc–lead sulphide than the outer and presumably younger part comprising the Abitibi–Wawa, Wabigoon, Amisk, and LaRonge belts (Table V-2).

Representative mineral occurrences in Archean rocks of the Hudson and Slave Protocontinents are shown on Figure V-5. The mineral occurrences are preferentially associated with the volcanic-rich belts. The gold, which occurs mainly in the volcanic belts of the Slave Protocontinent and in the Ennadai belt of the Hudson Protocontinent, is commonly associated with felsic rocks. Iron occurrences, both as magnetite iron-formation and iron sulphide, are present in eastern Hudson Protocontinent. Some iron and nickel lie in sedimentary rocks some distance from known volcanic rocks. Compared with the Superior Protocontinent the number of known mineral occurrences in the Slave and Hudson Protocontinents is much less, a metallogenic relationship in keeping with the smaller number and size of volcanic-rich belts. In addition, copper–zinc sulphide occurrences are scarce (Table V-3).

Thus, the geological and metallogenic relations in the Slave and Hudson Protocontinents appear to be similar to those prevailing in the inner part of the Superior Protocontinent and to represent gold–iron–nickel mineral provinces (Table V-4). The outer part of the Superior Protocontinent represents a more diversified and richly endowed base metal sulphide–gold–iron–nickel mineral province. The trend to metallogenic diversity in the outer, presumably younger parts of the protocontinent may be matched by increasing chemical diversity, particularly the alkalinity of igneous suites including the volcanic rocks.

Proterozoic Crust

The distribution of Proterozoic supracrustal rocks is dominated by comparatively long linear and broad basinal elements (Map 1250A). Northeasterly and, to a lesser degree, northwesterly trends predominate. Flat-lying to gently dipping strata are common. The main Proterozoic belts extend intermittently from Lake Superior to Lake Mistassini, across northern Quebec and Labrador, the northern tip of Ungava, and along the east shore of Hudson Bay. Proterozoic rocks underlie large areas south and east of Lake Athabasca, south and east of Great Slave Lake, 200 miles farther east at Dubawnt and Baker Lakes, and in the vicinity of Great Bear Lake and Bathurst Inlet. Additional Proterozoic belts lie in the Arctic Islands to the north and on Baffin Island to the east.

Proterozoic sedimentary assemblages include mature quartzite and sandstone, limestone and dolomite, carbonaceous shale, Superior-type iron-formation, and red beds. Because of distinctive lithology, substantial thickness, broad stratigraphic continuity, and abundant diagnostic criteria such as crossbedding, ripple-marks, raindrop impressions, mudcracks, intraformational conglomerates and breccia, and stromatolites, many Proterozoic assemblages are attributed to shallow-water, stable platformal, or miogeosynclinal environments of accumulation.

Proterozoic volcanic assemblages of the Shield are represented mainly by thick, tholeiitic basalts in the Lake Superior area, at Coppermine River in the Northwest Territories, and at Seal Lake in eastern Ungava. They resemble younger continental flood basalts characteristic

of stable continental sites. In addition, geosynclinal-type tholeiitic basalts characterized by high iron content (Baragar, 1960) are present in the eastern part of the Labrador Geosyncline, and trachytic volcanic rocks in the Dubawnt Lake area of the Northwest Territories (Donaldson, 1965). Other felsic volcanic rocks are present locally.

Granitic rocks of Proterozoic age are widespread in the Shield. Many Proterozoic intrusions and complexes have been indentified in the Northwest Territories in the southern and eastern parts of Churchill Province, in Grenville and Nain Provinces, and in certain parts of Superior Province. A reconnaissance geochemical study of Proterozoic granitic terrains has established an average composition (Fahrig and Eade, 1968; Shaw, *et al.*, 1967) that is close to that of common granodiorite. A comparison of average K_2O, U, and Th contents in Proterozoic and Archean crystalline rocks indicates a significant

TABLE V-3 *Representative mineral occurrences in the Slave and Hudson Protocontinents*

Protocontinent	Gold	Copper–Zinc	Nickel	Iron	Total
Hudson	16	7	8	15	46
Slave	30	7	3	5	45
Total	46	14	11	20	91
Per cent	51	15	12	22	100

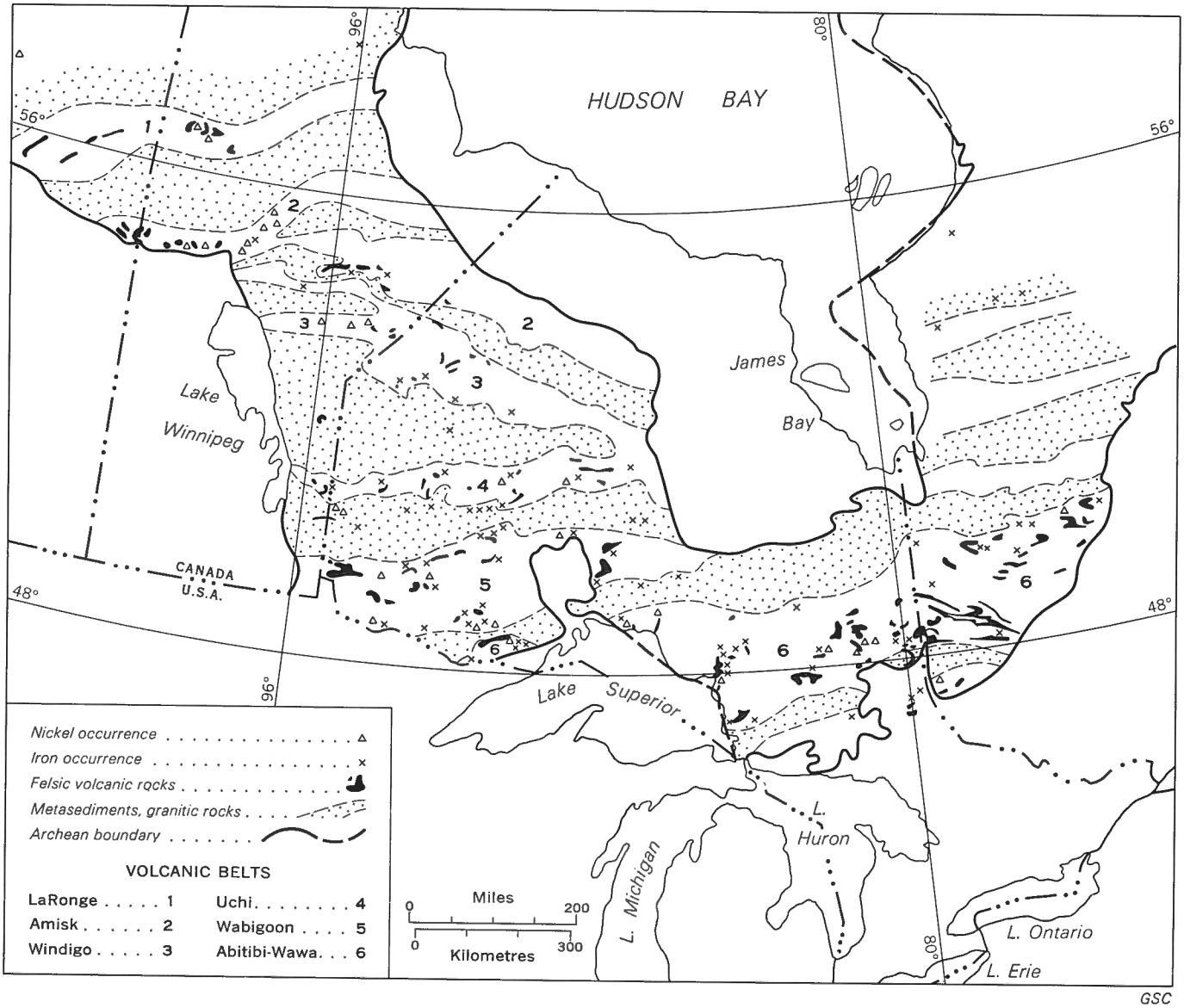


FIGURE V-3. Representative nickel and iron occurrences in the Superior Protocontinent (by A. M. Goodwin).

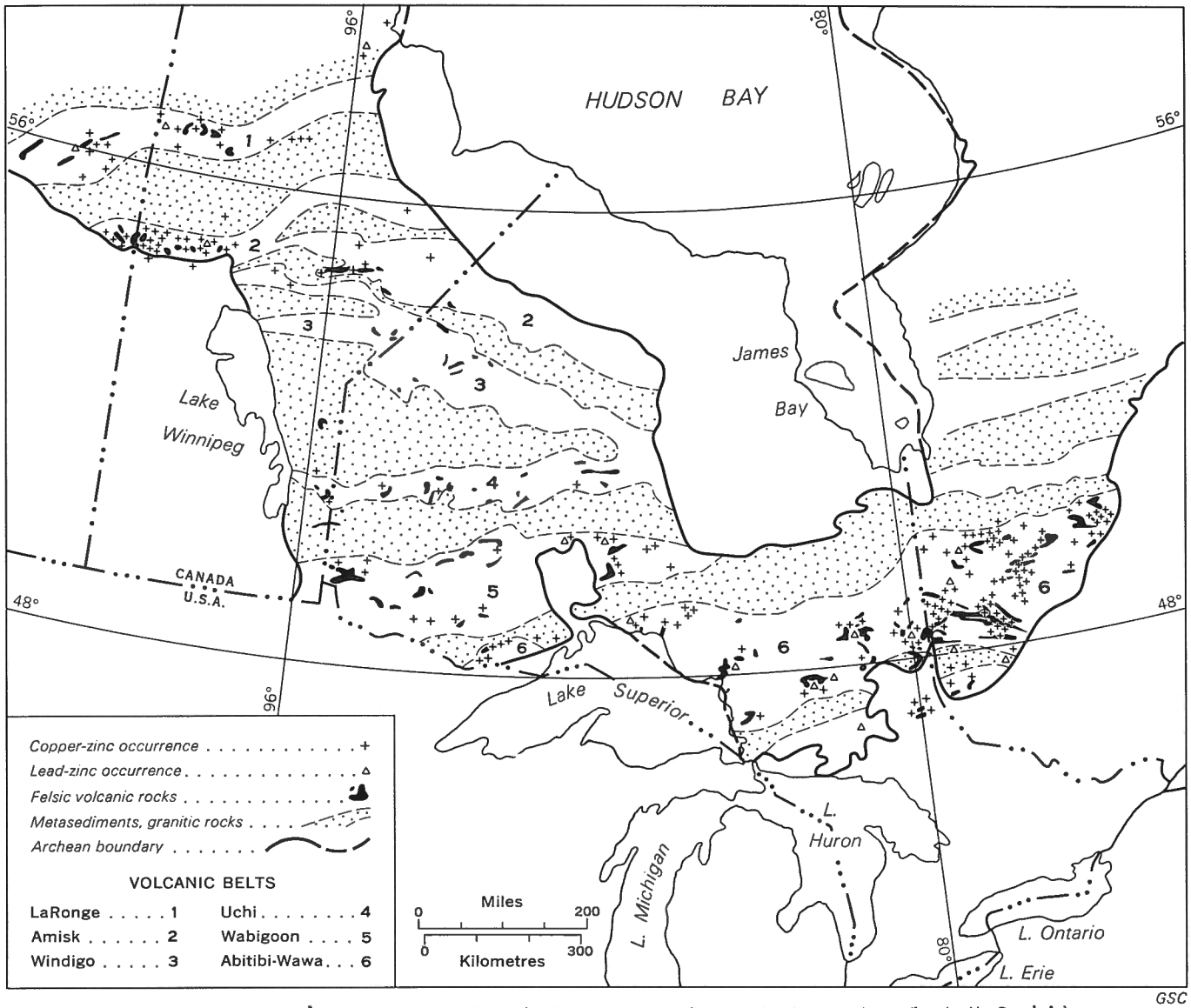


FIGURE V-4. Representative copper–zinc and zinc–lead occurrences in the Superior Protocontinent (by A. M. Goodwin).

increase of 1.8 times in radioactive level in the Proterozoic granitic rocks, notably in uranium.

Proterozoic Mineralization

Proterozoic rocks of the Canadian Shield contain a distinctive suite of mineral deposits: 1) uranium deposits of three main types: the quartz-pebble conglomerate type of Blind River, Ontario, and the northwestern Shield, the pegmatitic type of Grenville Province, and the vein type of the northwestern Shield (Lang, *et al.*, 1962); 2) granular iron-formation of the Superior type (Gross, 1965) in the Lake Superior, Labrador, Belcher Islands, and Lake Mistassini regions; 3) complex nickel–platinum ores associated with a mafic intrusion at Sudbury; 4) complex silver sulpharsenides associated with mafic intrusions at Cobalt and in the Thunder Bay area of the Lake Superior region; 5) copper deposits in Proterozoic volcanic and sedimentary rocks in the Lake Superior region and Coppermine River area of the Northwest Territories; and 6) minor lead sulphide occurrences in several regions.

Proterozoic mineral occurrences or groups of related occurrences are generally large and extensive corresponding in form and habit to the Proterozoic host rocks. The lithological character and distribution of Proterozoic rocks point to the presence of stable platforms, basins, and primitive geosynclines. These features imply substantial crustal stability and continuity of linear elements, hence the presence of a comparatively thick stable crust. Development of a thick Proterozoic crust is attributed mainly to growth and aggregation of Archean protocontinents and sub-crustal accretion of new, mantle-derived crustal material (Goodwin, 1968). In large

measure, Proterozoic tectonic history of the Canadian Shield reflects progressive crustal thickening and stabilization. Proterozoic mineral deposits, whether placer uranium concentrations in quartz-pebble conglomerates, granular iron-formations, lead–zinc deposits in reef carbonates, or complex sulpharsenides in mafic intrusions, are products of a comparatively thick, stable crustal regime.

In lithological character, tectonic pattern, and mineral habit Proterozoic rocks more closely resemble Phanerozoic rocks than Archean rocks. This is viewed within the context of a unidirectional evolving crust, the metallogenic associations constituting one important facet of this evolution.

TABLE V-4 *Percentages of representative mineral occurrences in Archean protocontinents*

	Superior Protocontinent			Hudson and Slave Protocontinents
	Total (%)	Annulus (%)	Nucleus (%)	(%)
Gold	39	38	49	51
Copper–zinc–lead	40	44	11	15
Iron	14	12	29	22
Nickel	7	6.	11	12

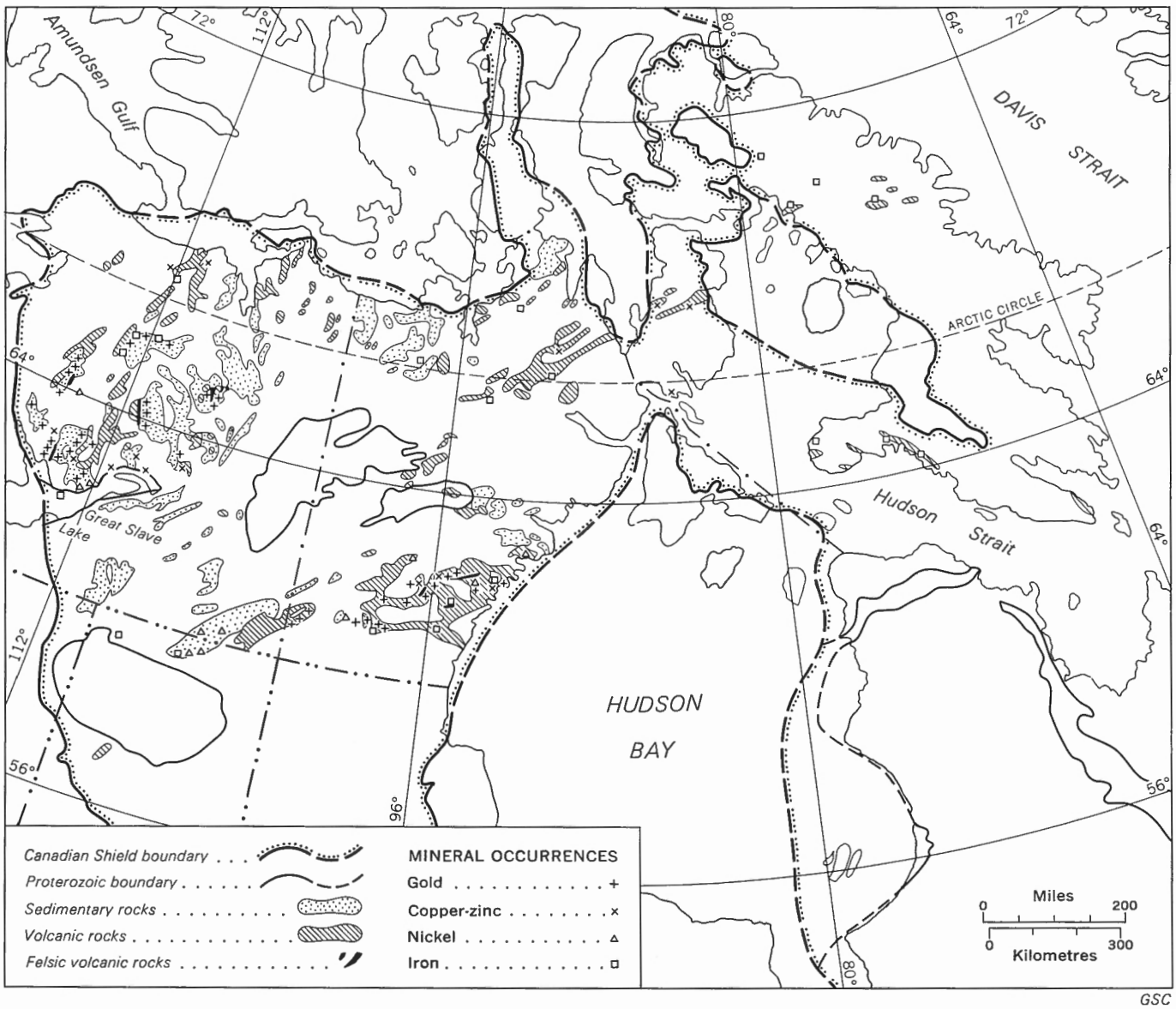


FIGURE V-5. Representative mineral occurrences in relation to Archean sedimentary and volcanic rocks in northern Canadian Shield (by A. M. Goodwin).

Uranium

Canada is one of the world's leading producers of uranium and has the largest known reserves of ore (Fig. V-6). All production has been from the Canadian Shield (Table V-1), and began in 1942 at the Eldorado mine on Great Bear Lake. Larger deposits were found in northern Saskatchewan at Beaverlodge in 1946 and in Ontario at Blind River in 1949. Uranium was actively sought from 1948 to 1955 and thousands of occurrences were found in various parts of the Shield. Expiry of contracts forced closure of several mines after 1960 and others closed as ore was exhausted. In recent years demand has been increasing again. Uranium is fairly common among the metals of the earth's crust, being more abundant than gold or silver but less so than copper, nickel, or zinc. Uranium has a strong tendency to form oxides rather than sulphides and to be dispersed rather than concentrated in high-grade deposits. In Canada, a content of 0.05 per cent uranium oxide (U_3O_8) or its equivalent has been chosen as the lower limit for substances to be classed as radioactive under the Atomic Energy Control Regulations. This limit has been followed in defining occurrences of uranium and thorium; smaller quantities are present at many places but are not considered occurrences. Prospecting and mining for uranium are governed by regulations of the Atomic Energy Control Board which require that discoveries be reported to the Director, Geological Survey of Canada, Ottawa and that permits for advanced exploration and mining be obtained from the Secretary, Atomic Energy Control Board, Ottawa. Export permits are required for shipments abroad.

Canadian reserves of ore were estimated in 1967 to contain 490,000 tons of U_3O_8. Most Canadian reserves are at Elliot Lake, Ontario. Other reserves are mainly in the Athabasca area. Many other areas contain partly tested prospects and generally favourable conditions (Roscoe, 1966). Non-producing areas that attracted most attention in 1966 and early 1967 were the Agnew Lake area between Blind River and Sudbury, Ontario, the Havre St. Pierre district in Quebec, the Makkovik district in Labrador, and Bancroft, Ontario.

Other common radioactive elements are thorium and radium. Demand for thorium is not large at present but it may become supplementary to uranium in the future. The ores of Elliot Lake average about half as much thorium oxide as U_3O_8, mainly because monazite is present. The Bancroft ores carry thorium, mainly in uranothorite. The high price of radium is mainly due to the long, careful processes required to separate it from uranium. Artificially produced radioactive isotopes such as cobalt-60 are cheaper and more effective for many purposes, causing such a decline in demand for radium that the plants at most uranium mines were not designed to save the radium, and the value of contained radium is not considered in estimating ore. Limited amounts of thorium and certain rare-earth elements are recovered as byproducts from some of the Elliot Lake ores.

Types of Deposits

About fifty mineral species or varieties containing uranium have been found in Canada. The number of ore minerals from which uranium has been obtained commercially is much smaller, consisting essentially of: uraninite (uranium oxide) and its variety pitchblende; 'brannerite'; monazite; uranothorite (silicate of thorium and uranium); and uranophane (hydrated silicate of calcium and uranium). Some ores contain a little uranium-bearing hydrocarbon called 'thucholite'. The distinction between uraninite and pitchblende has proven useful in Canada as the crystalline uraninite and the cryptocrystalline or botryoidal pitchblende occur in different types of deposits. Also, crystalline uraninite contains appreciable amounts of thorium and rare earths, whereas pitchblende carries less than 0.1 per cent of these elements. About half the uranium minerals are supergene but of these only uranophane has been found in commercial quantities in Canada. Almost all the supergene minerals have an earthy appearance and a distinctive canary-yellow colour and occur in small amounts at or near the surface of many primary occurrences. They are referred to as 'uranium stain' and serve as a useful guide in prospecting although they are difficult to identify individually except by laboratory methods.

In order of productivity, Canadian uranium ores are: 1) quartz–pebble conglomerates containing "brannerite", uraninite, and uraniferous monazite; 2) disseminations, stringer-systems, and veins of pitchblende; and 3) 'dykes' of syenite and granite, with pegmatitic facies, containing uraninite and uranothorite (Fig. V-6). The conglomeratic ores, which first became productive in 1955, are in the Elliot Lake area of Ontario and lie near the unconformity at the base of the Huronian strata. They contain about 0.1 per cent U_3O_8 and are mined successfully at this relatively low content because of the accessibility of the district and because they form extensive, thick, and uniform bodies that are more amenable to underground mining than deposits of the pitchblende type.

Pitchblende-bearing veins at Great Bear Lake were mined at first for radium as early as 1933. Copper, silver, and cobalt were also produced, along with a little uranium. The deposit was relatively small but of exceptionally high grade, as was necessary in that remote region. Larger but lower grade pitchblende deposits are those of the Lake Athabasca district, Saskatchewan, where production began at the Ace mine in 1953 and the Gunnar mine in 1955. The Gunnar ore was virtually unique in that it contained, along with pitchblende, much

supergene uranophane. The granitic type ores were mined near Bancroft, Ontario, where the Bicroft, Faraday, and Dyno mines and the Greyhawk were operated between 1956 and 1964. Uranium has also been found in several other kinds of deposits. Pegmatitic and granitic rocks occasionally contain accessory grains of radioactive minerals. Uranium has also been found in metasomatic and related deposits, quartzites, sandstones, and dolomites.

Elliot Lake district, Ontario (72–75)[1]. The leading

[1] Number assigned to the deposit on Mineral Deposits of Canada, Map 1252A.

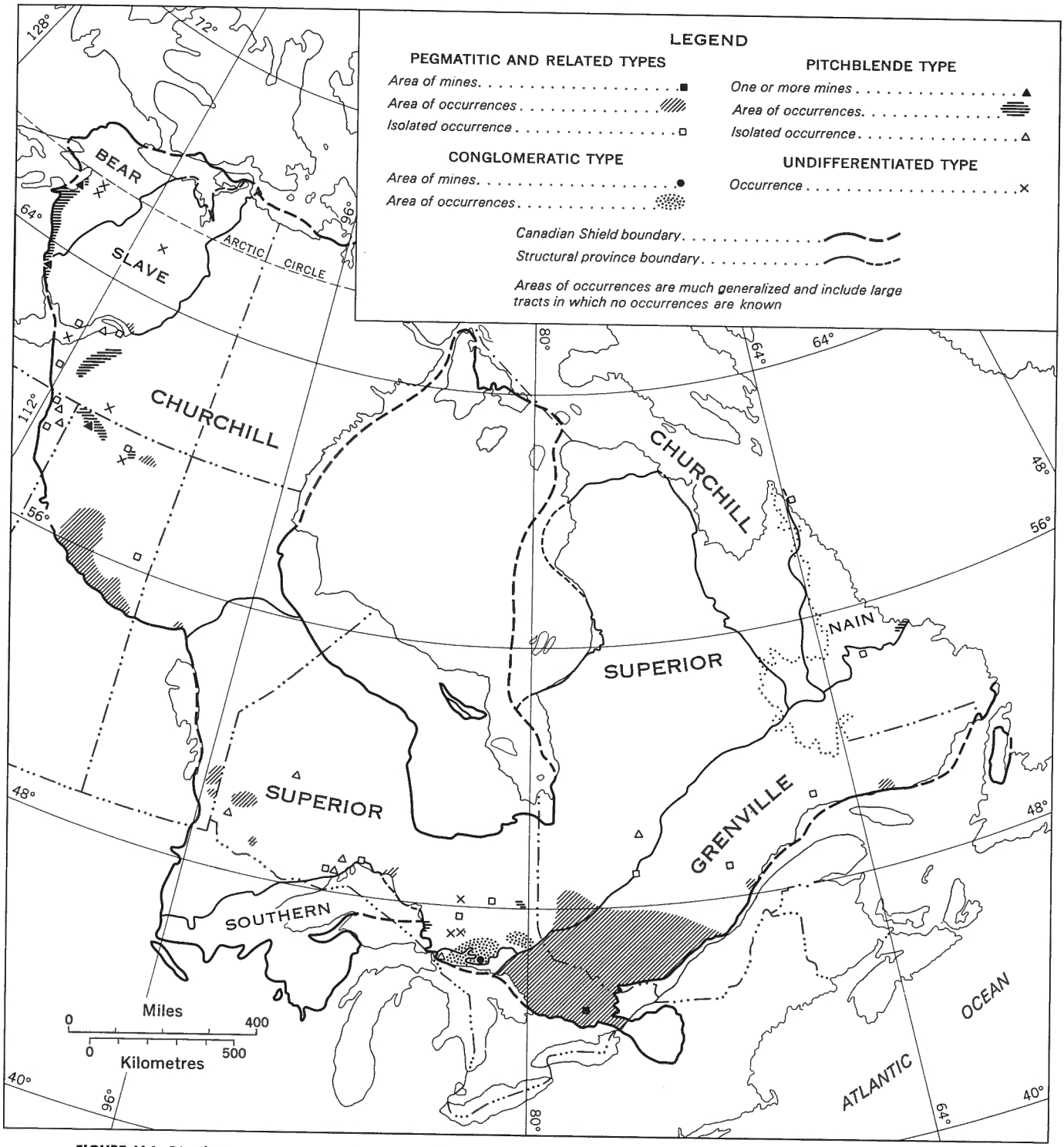


FIGURE V-6. Distribution of present and former uranium mines and known occurrences in the Canadian Shield (by A. H. Lang).

Canadian uranium-producing district is about 25 miles northeast of Blind River, Ontario. Huronian conglomerate outcropping near the north shore of Lake Huron, 11 miles east of Blind River, was found to be radioactive in 1949 and subsequently the Pronto mine was developed there in 1955. This discovery led to the finding of two large belts of deposits of the same general type as the Pronto in the vicinity of Elliot Lake some 15 to 20 miles to the north. These were discovered by diamond drilling to be structurally and stratigraphically favourable zones based on the general mode of occurrence of the Pronto ore. Eleven mines were developed from these two deposits, the largest being the Denison. The deposits are so nearly similar, except in size, that they are not described separately. The matrix of the conglomerate contains grains of pyrite 'brannerite', uraninite, monazite, hematite, magnetite, zircon, and other minerals. Some monazite is unusually high in uranium, possibly because of admixed uranothorite. Grains are commonly rounded but some grains of pyrite and uraninite are only slightly so. The material classed as brannerite presents a problem not yet fully solved. Hand-picked samples containing as much as 41 per cent U_3O_8 and 6 per cent ThO_2 are metamict and before heating yield X-ray patterns comparable to those of anatase and rutile, but after prolonged heating at high temperature they yield patterns characteristic of brannerite. Microscopically, many grains are seen to be mixtures, possibly of uraniferous anatase and rutile that were partly converted to synthetic brannerite by heating.

The beds of ore-bearing conglomerate of the Matinenda (lower Mississagi) Formation are composed of closely packed, subrounded quartz pebbles and cobbles. Most pebbles are a quarter inch to 2 inches in diameter. Those of the Elliot Lake sector (Fig. V-7) are in the Quirke Lake syncline which strikes easterly, plunges westerly, and contains Huronian strata over a width of nearly 10 miles. The syncline comprises as much as

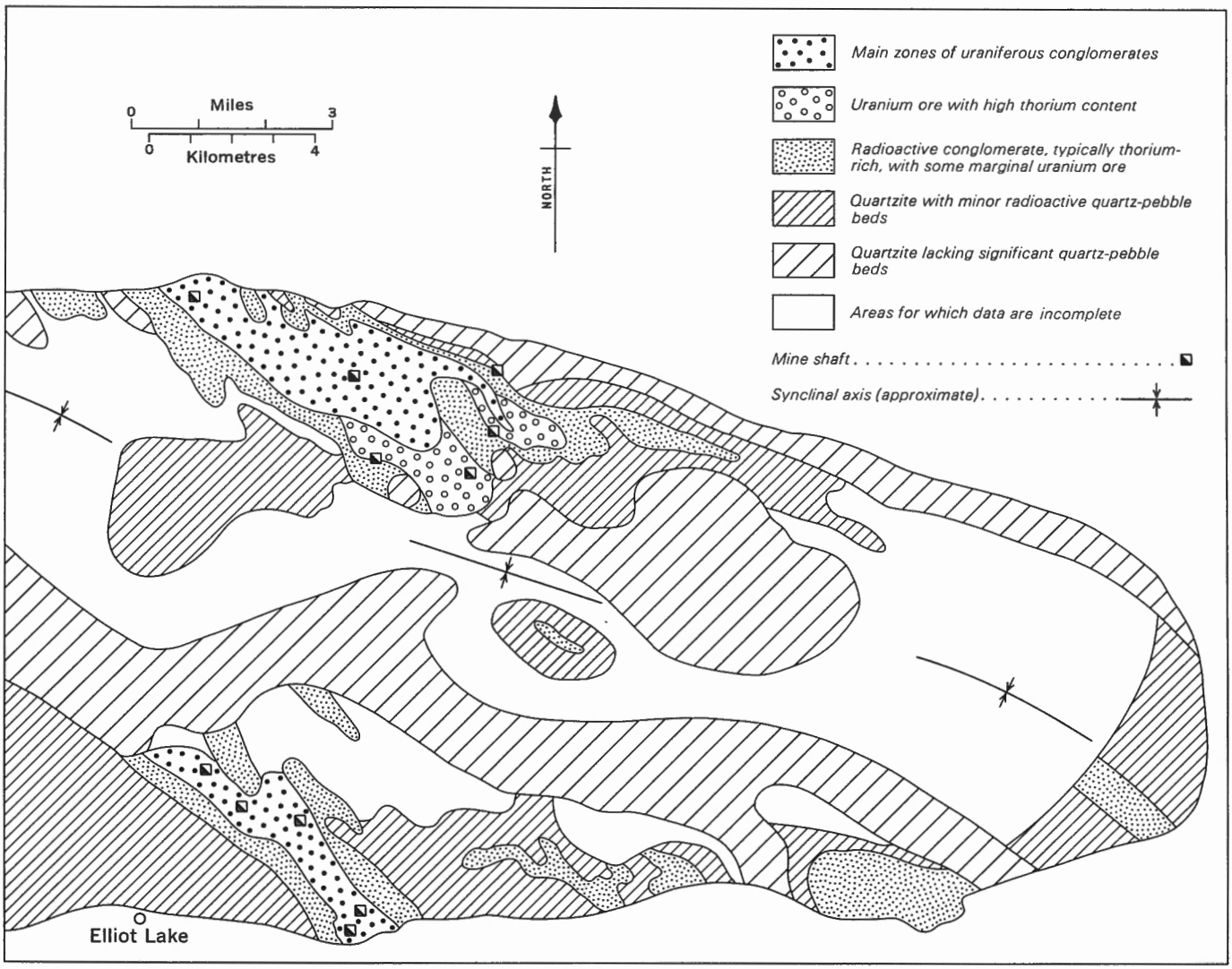


FIGURE V-7. Distribution of uranium ore, other radioactive conglomerates, and other strata in the Matinenda Formation, Quirke Lake syncline, Elliot Lake, Ontario (by S. M. Roscoe).

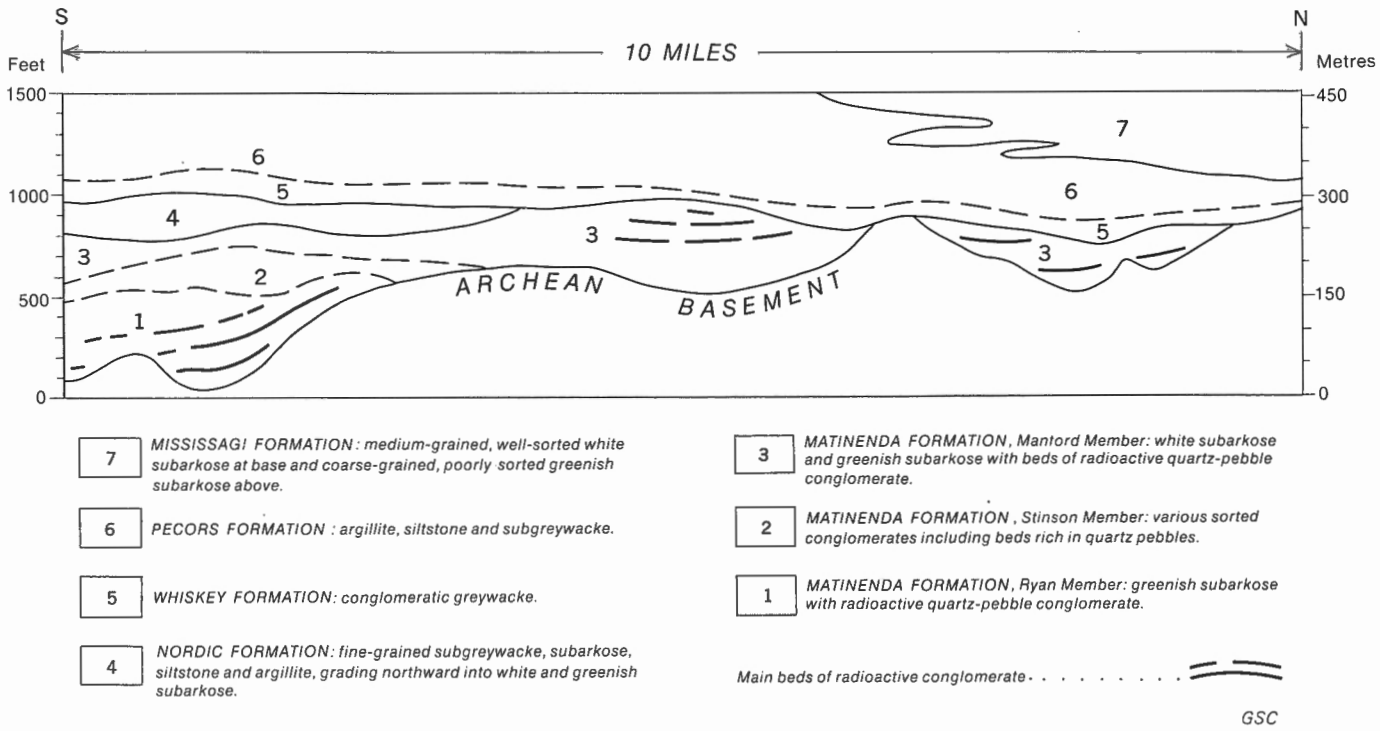


FIGURE V-8. Restored stratigraphic section through main uranium-bearing part of Quirke Lake syncline before Hudsonian folding (by S. M. Roscoe).

4,000 feet of strata, including up to 1,000 feet of Matinenda beds lying on the Archean basement (Fig. V-8). South of an anticline in basement rocks are southerly dipping Huronian rocks broken by the Murray fault, near which is the *Pronto* (72) mine. Most of the Pronto ore and some ore in the Quirke Lake syncline lie at the contact with the basement rocks, but most of the ore in the Quirke Lake syncline occurs in beds as much as 150 feet above the basement that vary from a single bed of conglomerate or grit to fairly intricate lenses of conglomerate and lean quartzite.

The average uranium content of the ores of the area as a whole is about 2.4 pounds of U_3O_8 a ton. The lowest grade material that is mined has about 0.06 per cent U_3O_8. The Elliot Lake ore-bearing conglomerates occur as accumulations in shallow valleys in the basement. Production has been confined to the southeasterly striking Quirke and Nordic belts. The Quirke belt contains six present or former producers: the Denison, Quirke, Spanish American, Panel, Can-Met, and Stanrock. The Denison mine, with a daily capacity of 6,000 tons of ore, is the largest in the entire district. The Nordic belt contains five present or past producers: the Nordic, Stanleigh, Milliken Lake, Lacnor, and Buckles. The belts end because of the interfingering of the conglomerate with other strata, the presence of ridges in the basement, and disconformable relationships with overyling strata. Studies of crossbedding and other features indicate that the sediments were derived from the northwest (Pienaar, 1963).

Explanations for the origin of the Elliot Lake deposits range from purely alluvial to purely hydrothermal. Most features of the deposits are consistent with a placer origin, whereas they are difficult to explain as a result of introduction of uranium-bearing fluids with resultant permeability-controlled deposition or replacement into the rocks. Roscoe and Steacy (1958) and others have shown that the deposits are restricted to a specific host rock, namely quartz-pebble conglomerate. Sharp, vertical and lateral variations in uranium content and mineralogy occur within the conglomeratic zones. The ore minerals are high-temperature types unlikely to have been formed in situ in the little-altered Huronian sediments and, moreover, occur in discrete, commonly rounded grains that resemble undoubted detrital minerals such as zircon, and are invariably associated with them. The high pyrite and uraninite content, the low magnetite content, and the low thorium to uranium ratio are features uncommon to placers. They could be explained by postulating special sources and depositional conditions during sedimentation or post-depositional modifications of the ores, including the introduction of hydrogen sulphide of biogenic or other origin to convert magnetite to pyrite. Little evidence has been presented for such special processes, however, and some geologists, therefore, consider hypotheses involving an epigenetic origin for the uranium more credible. Davidson (1957) suggested that the ores were deposited from uranium-bearing solutions derived from a deep-seated source but later proposed connate waters containing uranium dissolved from Huronian sediments as the source.

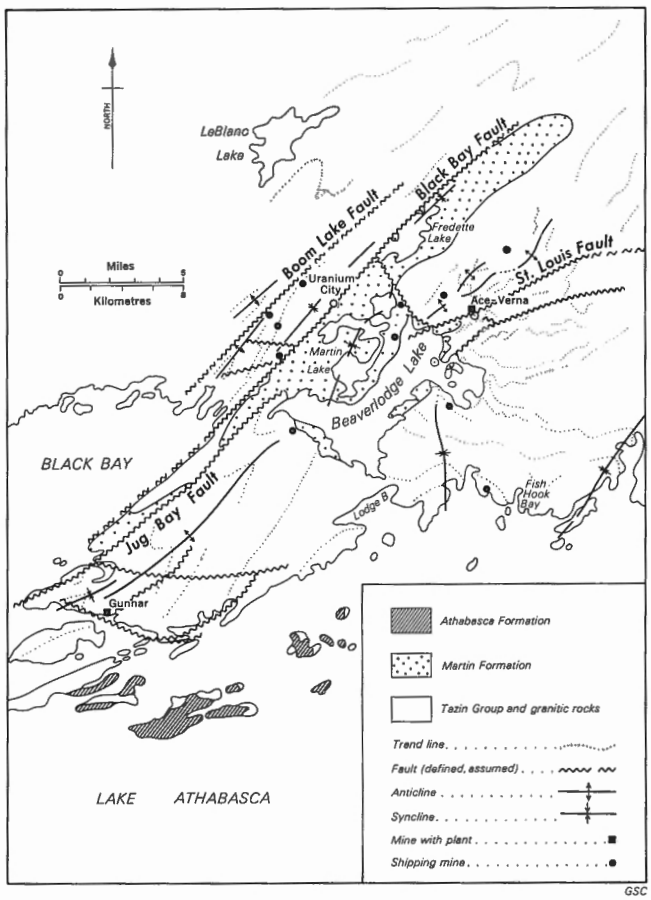


FIGURE V-9. Generalized geology and location of uranium mines, Lake Athabasca, Saskatchewan (by L. P. Tremblay and C. K. Bell).

Derry (1960) believed that the uranium was dissolved from weathered Archean rocks during deposition of the Huronian sediments and deposited biochemically from groundwater percolating through the unindurated gravels.

Athabasca district, Saskatchewan (1–4). The Athabasca or Beaverlodge district (Fig. V-9) on the north side of Lake Athabasca is underlain by highly metamorphosed sedimentary and volcanic rocks of the Archean or Aphebian Tazin Group, granitic gneiss and granite, and the Helikian Martin and Athabasca Formations. The distribution of uranium deposits in the area is controlled by regional and local structures and by lithology. The mineralized area is in a belt of folded and faulted rocks with a prevailing northeasterly trend. Faults are of two main ages, one group having originated after granitization of the Tazin rocks and the other just before and after deposition of the Martin strata. Zones of fracturing, brecciation, and mylonitization are commonly associated with the earlier faults. Some of the later faults may have followed the earlier fault traces.

The pitchblende deposits lie along the northeasterly trending faults, particularly at intersections with southeasterly trending faults and anticlines. They rake or plunge southwesterly, apparently partly because of the angles of intersection between the fault and fracture systems. Many show preference for rocks easily brecciated and mylonitized, and for argillites and mafic rocks. Most deposits are mineralogically simple—pitchblende with hematite as an alteration product. Some veins at the Nicholson property are more complex, having a mineralogical resemblance to those of the Port Radium, District of Keewatin. A vanadium mineral, nolanite, is fairly common in parts of the Ace orebodies and at the Fish Hook Bay prospect. The Gunnar ore contains much uranophane, also common at Fish Hook Bay. Bodies of migmatite and pegmatite in areas bordering the pitchblende occurrences carry uraninite and other radioactive minerals. The uraninite appears to have crystallized during the period of granitization in Tazin rocks and to have been followed by an early period of pitchblende mineralization. Younger pitchblende occurrences, some in the Martin Formation, probably represent redeposition of earlier material.

Production of the *Ace-Verna* (2) mine was begun in 1953 at the Ace orebody after an intensive prospecting program and diamond drilling of several of the better discoveries. The workings were connected underground with the Verna orebodies found nearby, mining being done at intervals for a distance of 14,000 feet and a maximum depth of 4,000 feet. The treatment plant has a daily capacity of 2,000 tons of ore that have averaged about 3 to 4 pounds of U_3O_8 per ton. The orebodies are in Tazin rocks along the St. Louis fault, which here strikes N60–70°E and dips 50° south. The Ace ore zones are in the footwall and those of the Verna are in the hanging-wall. The Ace ore is in brecciated feldspathic rock cemented by pitchblende, quartz, calcite, and chlorite, or in groups of narrow veins that occur as much as 300 feet from the fault. The Verna ores are groups of veins and veinlets in a mass of argillite 350 feet from the fault. Much of the Ace ore is sharply limited by the fault gouge on one side but is based on arbitrary grade limits on the other. Ore reserves in 1966 were reported to be 2 million tons grading 0.21 per cent U_3O_8.

The *Gunnar* (4) mine produced at a daily rate of 1,650 to 2,000 tons of ore from 1955 to 1964. The shape of the orebody permitted open-pit mining to a depth of 380 feet, below which underground operations were carried to a total depth of about 1,200 feet, mainly below lake-level. The ore was in brecciated porous albite–calcite–chlorite rock close to the intersection of two faults, the Zeemel which strikes northeast and the St. Mary's which strikes southeast along the shoreline. The general orebody plunged southward at 45° and comprised a group of small bodies elongated along the plunge, each with a clearly marked footwall and hanging-wall but irregularly limited in other directions. The ore consisted of disseminated pitchblende and uranophane in a gangue of albite, hematite, calcite, chlorite, and quartz.

Uranophane was unusually abundant even at low levels, probably because of circulation of groundwater through the porous rock. The grade of ore was fairly consistent at about 3.4 pounds of U_3O_8 a ton.

Bancroft district, Ontario (107, 108). The four productive mines of the Bancroft district are in the southwestern part of Grenville Province (Fig. V-6). The district has long been famous for rare minerals, mainly those of pegmatites and related deposits. Many radioactive minerals had been found and studied before uranium became important and an attempt to mine radium was made in 1929. After 1948 many more deposits were found, but not until 1952 were deposits of consistently higher uranium content discovered at *Bicroft* (108a), *Faraday* (107a), and *Canadian Dyno* (108b). Ore was shipped to the Faraday plant from the *Greyhawk* (107b) mine from 1957 to 1959. All mines were closed by 1964. The Bancroft area is underlain by three fairly circular masses, called the Cheddar granite, Cardiff complex, and Faraday granite; each is about 6 miles in diameter, and is composed of granite, syenite, gneisses, and related rocks. They are separated by metamorphic rocks of various kinds that exhibit a concentric structure. These are mainly marble, paragneiss, para-amphibolite, and meta-gabbro. The principal uranium deposits are in bodies of granite and syenite with pegmatitic and metasomatic phases that either cut the wall-rocks or replace them. The most favourable rocks were pegmatitic pyroxene granite or syenite, leucogranite, and cataclastic quartz-rich granite–pegmatite. All are commonly high in sodium. Many ore shoots were associated with concentrations of mafic minerals and magnetite.

The Bicroft mine is in paragneiss and amphibolite on the east flank of the Cardiff complex. Its ores were in an *en echelon* swarm of lenticular dykes in a zone half a mile wide and 3½ miles long. The ore minerals were uranothorite and uraninite, the ore averaging about 0.1 per cent U_3O_8. The Faraday mine is in a belt of meta-gabbro and amphibolite on the south flank of the Faraday granite. The ores were parts of a zone of irregular pegmatitic granite dykes, with metasomatic phases, extending at intervals over about 6,000 feet. The ore minerals were uraninite and uranothorite, with minor uranophane; ore averaged about 0.1 per cent U_3O_8. The Canadian Dyno mine is in a belt of paragneiss and other rocks on the east flank of the Cheddar granite. The property contains a series of pegmatitic granite dykes only some of which were of ore grade. The ore minerals were uraninite and uranothorite. The Greyhawk mine is in meta-gabbro on the south flank of the Faraday granite. Its ores were in dykes of pegmatitic granite containing uraninite and uranothorite.

Great Bear Lake area, District of Keewatin. The *Eldorado* (3) mine at Port Radium on the east shore of Great Bear Lake produced from 1933 to 1940 and 1942 to 1960. Although almost on the Arctic Circle the richness of the veins and the high price of radium permitted mining, and throughout its life the mine had the distinction of being the most northerly metal mine in Canada. Radium was the significant commodity during the earlier years, but silver, copper, and cobalt were produced as well. For several years the mine was one of the two main sources of radium in the world and the principal one on this continent. Although the pitchblende contained far more uranium than radium there was virtually no demand for uranium until 1942. Output was then increased to 300 tons of ore daily.

The deposits were in Bear Province; the host rocks were metamorphosed tuffs and coarse fragmental material interbedded with cherty sedimentary strata, all part of the Echo Bay Group of Aphebian age (Fig. V-35). Masses of feldspar porphyry are interbanded and are at least partly intrusive. These rocks are invaded by granite encountered in the lower levels. Diabase dykes are younger than the granite and are transected by structures associated with the ore. The youngest rocks are flat-lying sheets of diabase, apophyses of which cut ore. The Eldorado orebodies were in an area roughly 2,000 feet wide and a mile long, mined in places to 1,650 feet below the surface. The rocks are traversed by a series of steep northeasterly trending faults, some coalescing and some branching. The orebodies were veins and breccia-fillings in the faults, individual ore shoots being a few inches to 15 feet wide and 50 to 700 feet long; one was worked for a depth of 1,100 feet. The ore shoots were confined almost entirely to the stratified rocks or to places where the fault-zones followed contacts between them and diabase. They extended only a little way into parts of the fault-zones that crossed porphyry, granite, or massive tuff. Ore shoots appeared to be grouped around 'noses' of porphyry.

Forty metallic minerals were present including cobalt or nickel arsenides and sulpharsenides. Gangue consisted of altered rock, quartz, and carbonate minerals. The nickel minerals, native silver, botryoidal pitchblende, the supergene minerals, and red hematitic alteration occurred only in the upper levels. Mineralization took place in four stages, the pitchblende apparently being deposited first and at fairly low temperatures and pressures. All evidence indicated that the mineralization was related to the granite. Some veins extend into it, suggesting that the mineralization was a late stage of the intrusion itself rather than redistribution, by metamorphism or granitization, of metals contained in the invaded rocks.

In southern Bear Province, the *Rayrock* (5) mine was operated for two years with a 150-ton plant. Pitchblende occurs in shoots in a northeasterly trending large quartz stockwork. These are locally called 'giant quartz veins' and are fairly abundant in southern Bear Province.

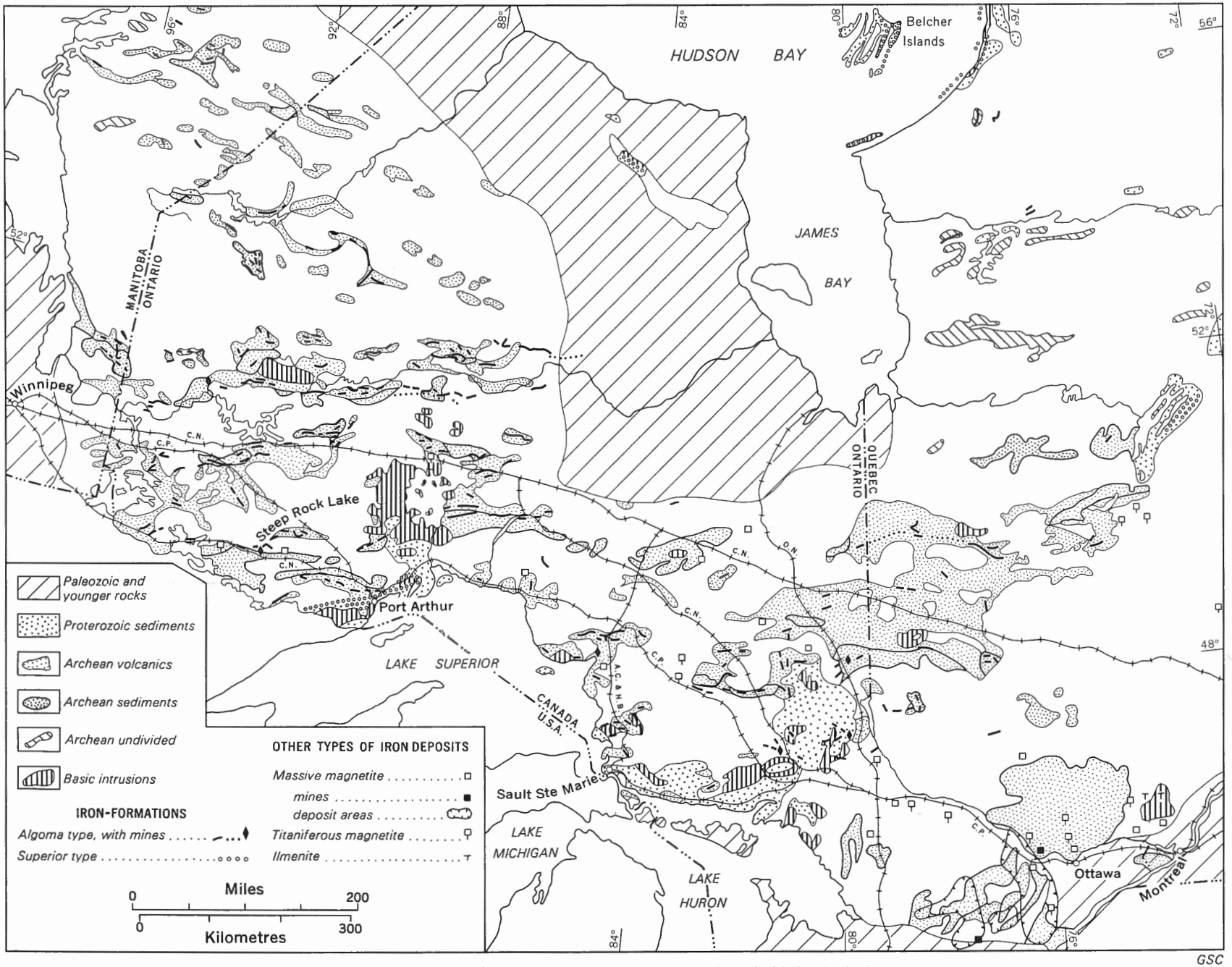


FIGURE V-10. Iron deposits and mines of the southern Canadian Shield (by G. A. Gross).

Iron

Canada produced 36 million long tons of iron ore in 1966, valued at $420 million, and was the fourth largest producer in the world after France, the United States, and the Soviet Union. Iron ore exceeds nickel in value, and was surpassed only by copper as the most valuable metallic commodity produced in Canada in 1966. More than 94 per cent of the iron produced was derived from the Precambrian rocks of the Canadian Shield (Table V-1). The opening of the Helen mine in the Michipicoten district of Ontario in 1939 marked the beginning of a new period of development and rapid growth in the Canadian iron ore industry after a non-productive stage of nearly 20 years. The Steep Rock iron range was developed during World War II to help offset increased continental market needs. In the decade after 1950 as a result of a phenomenal expansion in exploration and growth in the industry, ore production rose from 3.5 million tons to 22 million tons. The first ore was shipped from the Labrador–Quebec ranges in 1954 and since then production from that area has increased greatly.

A profound change has taken place during the past 10 years, especially in North America, in the grade and structural specifications for iron ore and hence in the processing required. Direct-shipping quality ore is no longer considered to be premium material, and pelletized high-grade concentrates, beneficiated and sintered ores are now in greatest demand. Research is being done on methods for direct reduction and partial reduction on ore concentrates with the aim of providing even higher grade furnace burdens. These revolutionary changes in processing technology and in market specifications are of major significance to Canada as it has become economical to concentrate and pelletize the widely distributed low-grade siliceous iron-formations that comprise the huge reserves of potential iron ore in the Canadian Shield. Iron is also recovered as a byproduct in the treatment of nickel and copper sulphide ores. More than one million tons of pellets and iron oxide calcine were produced in 1965 in the Sudbury area.

Types of Iron-formation

There are many types of iron deposits in Precambrian rocks in addition to the low-grade iron-formations. The classification of deposits as shown in Table V-5 is based on empirical and descriptive data (Gross, 1965). The general distribution of iron-formations and iron-producing areas in the southern part of the Shield is shown on Figure V-10. Some of the features of these siliceous iron-formations are summarized on Figure V-11 and are compared to other types of iron-bearing sediments that occur in Canada in rocks younger than Precambrian. The composition of four different types of iron-bearing sediments are compared in Table V-6 and general features

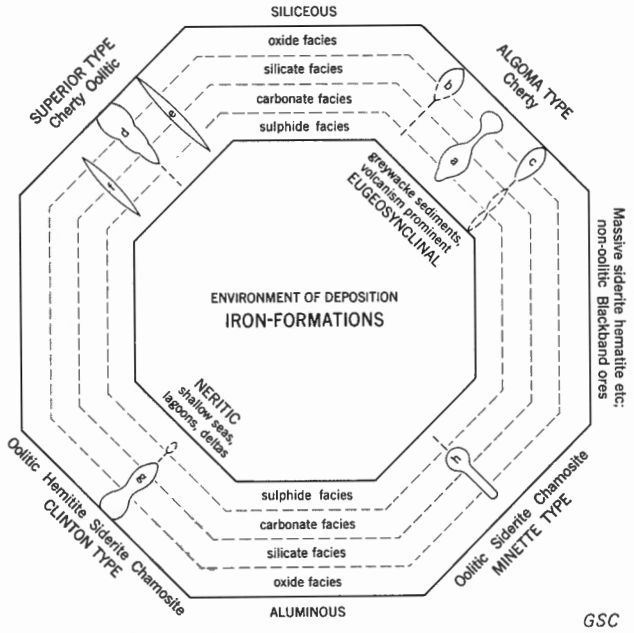


FIGURE V-11. Diagram showing major types of chemically precipitated iron-formation with their sedimentary facies and depositional environment. Examples are from: a, Michipicoten, Ontario; b, Moose Mountain, Ontario; c, Timagami, Ontario; d, Knob Lake, Labrador and Quebec; e, Iron River, Michigan; f, Gunflint and Biwabik iron-formation, Ontario and Minnesota; g, Wabana, Newfoundland; h, Clear Hills, Alberta (Gross, 1967).

are noted for the two prominent types found in the Canadian Shield. The geology of iron-formations and other ferruginous sedimentary rocks has become increasingly important during the last two decades. It is the basis for guiding prospecting, evaluating deposits of higher grade ore within the iron-formations formed by natural enrichment processes, and for selecting, exploring, and developing low-grade formations for production of high-grade concentrates. Almost every belt of volcanic and sedimentary rock in Superior and Churchill Provinces contains some iron-formation of the Algoma type. Younger Aphebian rocks, such as those of the Labrador Geosyncline or those near Mistassini Lake and west of Lake Superior, contain iron-formations of the Superior type. They were deposited in continental shelf environments.

The Algoma type of iron-formation was deposited with volcanic and sedimentary rocks characteristic of eugeosynclinal environments (Goodwin, 1965). Typical beds are composed of grey or red jasper chert interlayered with magnetite and hematite-rich layers (Pl. V-1), but other mineral facies, such as massive siderite and pyrite–pyrrhotite beds, occasionally form part of the formation. Algoma type iron-formations range in thickness from a few inches to more than 100 feet and are usually less than a few miles long. They were commonly deposited during a volcanic cycle near the close of a rhyolitic stage and before andesitic or basic extrusion. Carbon-bearing



PLATE V-1. Algoma-type, Archean, iron-formation of magnetite, siderite, and chert. Helen mine, Michipicoten district, Ontario.

schists containing iron sulphide minerals are closely associated with the cherty iron-formations in this environment. Algoma type beds are abundant in Precambrian volcanic belts, and similar beds also occur in the Appalachian and Cordilleran Geosynclines.

Iron-formations of the Superior type are typically thin-banded cherty rocks with granular to oölitic textures, and contain little or no clastic material. Lithological facies, either cherty magnetite and hematite or the cherty iron silicate and carbonate varieties, are prominent and usually separated in well-defined stratigraphic units. The beds show abundant evidence of deposition in shallow water and many were deformed when the sediment was in a hydroplastic state. Superior type beds are commonly associated with quartzite, dolomite, black ferruginous slate, argillite, chert, and various volcanic rocks, including tuffs and lava flows and the sedimentary rocks derived from them. They were deposited on continental shelves or in miogeosynclinal basins or troughs occurring within narrow belts of sediments bounded by ancient cratonic areas on the one side and by volcanic belts on the other. These iron-formations are hundreds of feet thick and may extend continuously along the margin of the geosyncline for hundreds of miles. Superior type formations are the protore throughout the world for most of the large, naturally enriched, hematite–goethite ore deposits.

Iron Deposits in Superior Type Iron-formation

Superior type iron-formations in the Labrador Geosyncline extend more than 700 miles from west of Ungava Bay southeast beyond *Schefferville* (75), Quebec, and *Wabush Lake* (1), Labrador, and thence southwest for an additional 200 miles within Grenville Province (Fig. V-12). In the central and northern part of the belt they were folded and deformed during the Hudsonian Orogeny. Rocks of the southern part of the geosyncline within Grenville Province have been redeformed by the Grenvillian Orogeny and possibly also by the Elsonian Orogeny. The iron deposits differ in the northern, central, and southern parts of this belt. In the northern part of the geosyncline, west of Ungava Bay, rocks are highly metamorphosed and the recrystallized medium- to coarse-grained iron-formations form extensive metataconite deposits (Type IIE) suitable for beneficiation.

Iron-formations in the central part of the Labrador Geosyncline are not highly metamorphosed and form the host rock for hematite–goethite ore deposits (Type IA). The unoxidized iron beds constitute a vast reserve of

potential taconite ore (Type IID). The iron-formations are composed mainly of iron oxide and silicate–carbonate and because of their fine-grained cherty character were especially amenable to natural enrichment processes that produced pockets and masses of direct-shipping hematite–goethite ore of Type IA (Fig. V-13). During Mesozoic time groundwater is believed to have penetrated the iron-formations along permeable zones connected with deeply weathered erosion surfaces. It leached silica, and oxidized and remobilized some of the iron causing secondary enrichment and formation of ore. Existing ore deposits are the roots of larger deposits that were partly eroded during

TABLE V-5 *Classification of iron deposits (Gross, 1967)*

GROUP

I HEMATITE–GOETHITE DEPOSITS IN IRON-FORMATION. Direct-shipping ores and wash ores formed in iron-formation by leaching of silica and concentration of iron by natural processes.

Type A: i) Lake Superior type in Minnesota, and Knob Lake type in Quebec and Labrador. Concentrations of hematite, goethite, and other iron minerals in cherty iron-formation. Cherty iron silicate, carbonate, and cherty iron oxide iron-formations are the common kinds of protore for this type.

ii) Semitaconites, partly leached iron-formations associated with Type Ai) that provide "wash ore". These are considered as a special subtype in this group.

Type B: The Old Helen mine deposit, Michipicoten district, Ontario. Concentrations of hematite and goethite of direct shipping quality in siderite beds.

II IRON-FORMATIONS. Various kinds or types of iron-formation that may be utilized without concentration or beneficiation, or that are suitable for beneficiation.

Type A: Clinton type of iron-formation, Wabana deposits, Newfoundland. Hematite–siderite–chamosite beds with distinctive oölitic texture.

Type B: Minette or Lorraine oölitic siderite–chamosite iron-formations, mined in the Northampton district of England.

Type C: Helen mine and Sir James mine, Michipicoten district, Ontario. Siderite iron-formation consisting of stratigraphic units of fairly massive siderite.

Type D: Taconite iron-formation, Moose Mountain mine, Ontario, and Mesabi Range taconites in Minnesota. Included here are all iron-formations that must be ground finer than 100 mesh to be concentrated. Common types include iron oxide, iron-silicate, iron-carbonate, and iron silicate-iron oxide types with a banded cherty matrix.

Type E: Metataconite iron-formation, Wabush Lake, Labrador, Newfoundland, and Lac Jeannine, Quebec, or the itabirite deposits in Minas Gerais, Brazil. Included here are metamorphosed iron-formations that have undergone extensive recrystallization and can be concentrated without grinding finer than 100 mesh.

Type F: Pyrite-rich iron-formations of potential ore value.

Type G: Other types of iron-formation of distinctive character found in local areas.

III RESIDUAL DEPOSITS, AND CHEMICALLY AND MECHANICALLY TRANSPORTED SURFACE DEPOSITS

Type A: Steep Rock Lake type, Steep Rock Range, Ontario. Buried residual accumulations of iron minerals on regoliths, with subsequent alteration of iron minerals and redistribution of iron. Mainly massive to crudely banded red and brown hard rubbly hematite and goethite ores of direct-shipping quality.

Type B: Residual deposits, Londonderry area, Nova Scotia; Bilbao, Spain.

Type C: Laterite deposits developed by surficial oxidation and supergene enrichment of rocks other than iron-formation, Conakry, French Guinea.

Type D: Bog Iron deposits, St. Maurice, Quebec. Iron is transported chemically and precipitated near or on the surface.

Type E: Placer deposits and deposits formed by mechanical concentration of iron minerals. Iron sands of the North Shore of the St. Lawrence River, or the Charleson hematite gravel deposit south of Steep Rock Lake.

IV DEPOSITS DIRECTLY ASSOCIATED WITH PLUTONIC ROCKS

Type A: The St. Charles deposit, Bourget township, Quebec; or Chaffey mine at Newboro Lake, Ontario. Layered, disseminated, interstitial, and injected magnetite with minor hematite and ilmenite in basic and ultrabasic rocks.

Type B: Allard Lake deposits, Quebec; or Taberg type of deposit in Sweden. Layered, disseminated, interstitial, and injected ilmenite with magnetite and hematite in anorthosite rocks.

Type C: Stephenville, Newfoundland. Deposits in acid intermediate rocks or in alkalic rocks. Massive and disseminated ores in these rock types. Further subdivision may be based on types of host rock.

V SKARN, CONTACT METASOMATIC, VEIN, AND STRUCTURALLY CONTROLLED EMPLACEMENTS OF IRON MINERALS

Type A: Texada Island, British Columbia; Marmora deposit, Ontario. Skarn, contact metasomatic deposits or high temperature replacement deposits.

Type B: Forsyth deposit, near Hull, Quebec. Veins with minor replacement, deposited under structural control.

Type C: Disseminated magnetite, iron oxides, and iron minerals that are probably of replacement origin and deposited under structural control in shear zones and other structural features.

Type D: Noranda deposits, Quebec. Massive and disseminated iron sulphide bodies, may contain base metals and other ore minerals.

VI OTHER TYPES OF DEPOSITS

Pleistocene glaciation. More than 45 deposits of this kind, some of which contain as much as 50 million tons of ore, are present at Schefferville near the border of Quebec and Labrador (Pl. V-2). Conditions are favourable for the occurrence of this type of ore throughout most of the central part of the belt. These ores are composed of porous, friable to rubbly, red, blue-black, or brownish yellow masses of hematite and goethite formed within bands of cherty oxide and silicate–carbonate iron-formation. Ore of direct-shipping quality usually contains from 51.5 to 65% iron, as much as 12% silica, 0.045% phosphorus, 1.5% alumina, and small amounts of lime and magnesia. Manganese content varies, averaging less than 3%.

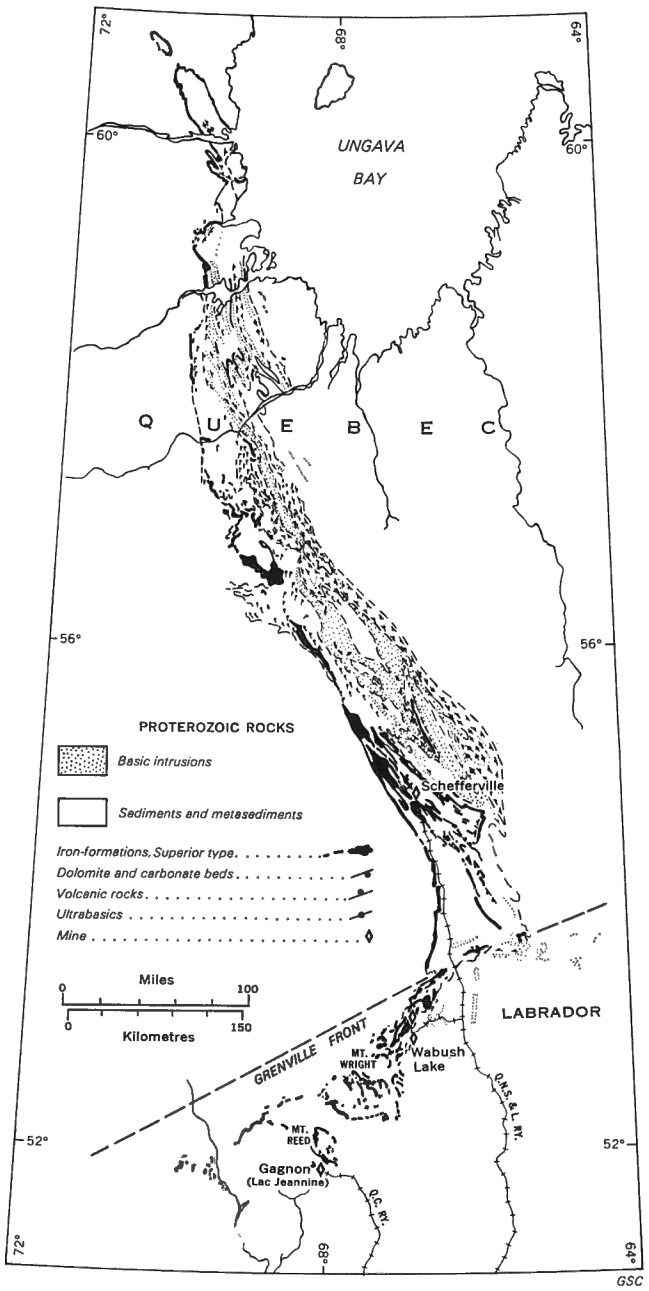


FIGURE V-12. Iron-formations of the Labrador Geosyncline (by G. A. Gross).

TABLE V-6 *Compositions of iron-formation types*

Type	Algoma	Superior		Clinton	Minette
Facies	Oxide (1)	Oxide (2)	Silicate carbonate (3)	(4)	(5)
Fe	33.52	33.97	30.23	51.79	30.97
SiO_2	47.9	48.35	49.41	11.42	28.06
Al_2O_3	0.9	0.48	0.68	5.07	5.79
Fe_2O_3	31.7	45.98	16.34	61.83	29.81
FeO	14.6	2.33	24.19	11.00	13.08
CaO	1.45	0.1	0.1	3.32	1.92
MgO	1.8	.32	2.95	0.63	1.54
Na_2O	0.2	.33	0.03	ND	0.33
K_2O	.32	.01	.07	ND	.53
H_2O^+	.47	2.0	5.2	1.94	13.10
H_2O^-	.1	0.04	0.38		
TiO_2	.05	.01	.01	0.015	0.18
P_2O_5	.1	.04	.08	1.96	1.59
MnO	.3	.025	.65	0.17	0.16
CO_2	ND	.03	.22	2.15	2.89
S	ND	.013	.05	0.023	ND
C	ND	.08	.15	ND	ND
Location	Timagami	Knob Lake		Wabana	Peace River

(1) Average of four analyses for 50 feet of section sampled systematically; Algoma type, mainly oxide facies, magnetite–quartz iron-formation; Timagami Lake area, Ont. (Analyses, Geol. Surv. Can. Laboratories.)
(2) Average of six analyses for 335 feet of section sampled systematically; Superior type, oxide facies, hematite–magnetite–quartz iron-formation; Knob Lake iron ranges, Quebec and Labrador. (Analyses, Geol. Surv. Can. Laboratories.)
(3) Analysis of 50 feet of section sampled systematically; Superior type, silicate–carbonate–chert facies; Knob Lake iron ranges, Quebec and Labrador. (Analysis, Geol. Surv. Can. Laboratories.)
(4) Analysis of large composite sample; Clinton-type iron-formation; Wabana, Newfoundland. (Courtesy, Wabana Mines.)
(5) Average of eleven analyses representing 15 feet of section; Minette type, oolitic siderite–chamosite–limonite iron-formation; Peace River District, Alberta. (Mellon, 1962.)

The iron-formations in the southern part of the Labrador Geosyncline (Fig. V-12), from the Wabush Lake area through Mount Wright and southwest to Matonipi Lake, are deposits of metataconite (Type IIE). Exceedingly complex structures are developed in this area as a result of deformation during both the Hudsonian and Grenvillian Orogenies. The coarse-grained texture in the highly metamorphosed iron-formations and the thickening and repetition of the iron beds by folding and faulting are controlling factors in the formation of large masses of easily concentrated metataconite ore.

The metamorphosed hematite and magnetite–quartz iron-formations derived from the sedimentary oxide facies supply good ore but the iron in beds rich in iron–silicate minerals and iron carbonates is difficult to concentrate. The oxide formations contain 30 to 40 per cent iron and pelletized concentrates produced from them contain 64

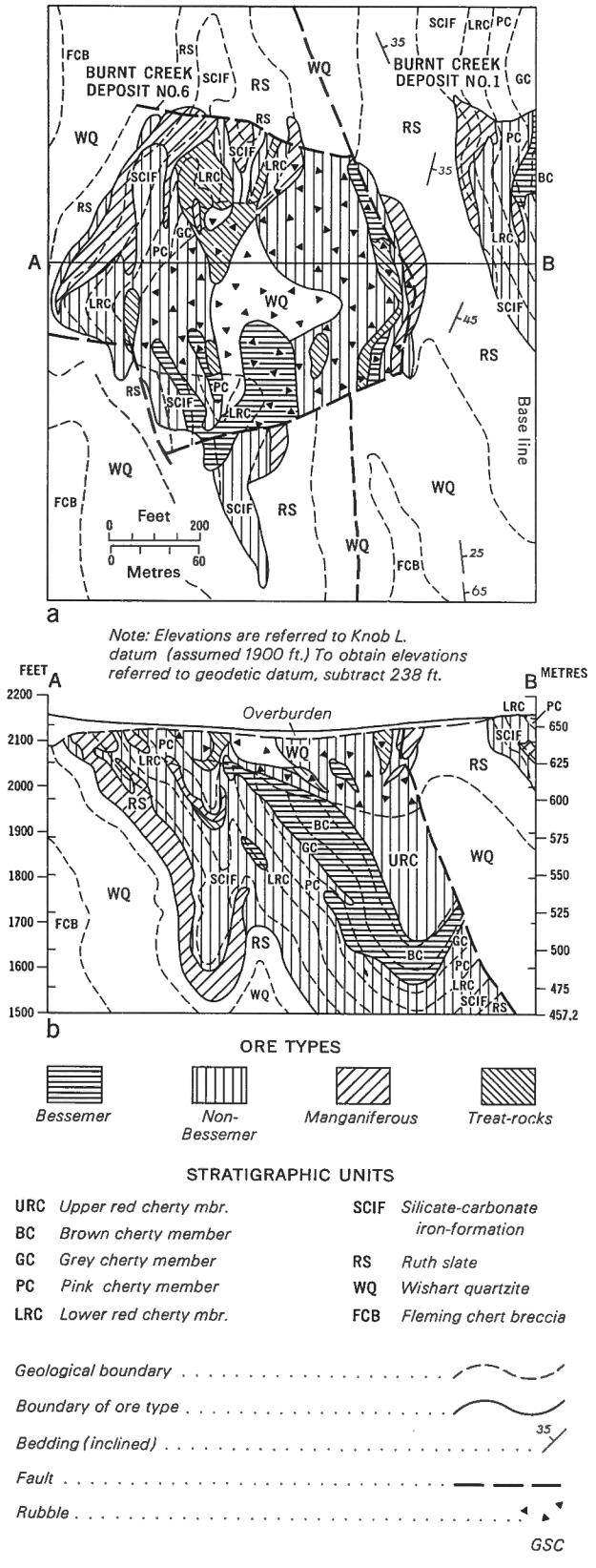


FIGURE V-13. French mine, Schefferville–Knob Lake area, Quebec (by G. A. Gross): (a) plan, 4B section of Burnt Creek deposit; (b) section along line A–B.

to 65 per cent iron with 4 to 5 per cent silica. Iron-formation mined in the Wabush Lake area contains about 38 per cent iron in the form of specular hematite and magnetite with quartz. Deposits at Mount Wright and Fire Lake, not yet produced, and *Lac Jeannine* (74) are typical of a group composed of specular hematite and quartz. Those such as the one at Lac Jeannine that lie well within Grenville Province are the coarsest grained. Concentration is done exclusively in gravity spirals. Other iron-formations composed of magnetite, specular hematite, quartz, and intermixed iron–silicate minerals, such as one at Mount Reed, contain billions of tons of potential ore but they are fine grained and difficult to concentrate. Mines in the Wabush area produce ore concentrate for 15 million tons of pellets per year and the Lac Jeannine mine near Gagnonville annually produces 8 to 9 million tons of concentrate. Conservative estimates indicate that 4 to 5 billion tons of ore concentrate can be obtained from 10 billion tons of metataconite ore mineable by open-pit methods in the southern part of the Labrador Geosyncline.

Superior type iron-formations are also present in Proterozoic rocks in the Cape Smith belt, along the east side of Hudson Bay on the Nastapoka and Belcher Islands, and near Sutton Lake in Ontario. They are also exposed for 35 miles along strike in the iron ranges near Mistassini and Albanel Lakes in central Quebec. The Gunflint iron range extends along the west shore of Lake Superior and southwest to link up with the Mesabi iron range in Minnesota. These iron-formations are relatively unmetamorphosed and contain large reserves of potential taconite ore. Some have been prospected for direct-shipping ores, but no commercial deposits have been found.

Iron Deposits in Algoma Type Iron-formation

Algoma type iron-formations in Superior Province of Ontario are becoming increasingly important as taconite ore (Type IID). Siderite beds with minor pyrite and pyrrhotite (Type IIC) are mined in Michipicoten district and the geology of the *Helen* (37) mine is typical of many iron deposits in the area (Fig. V-14). The siderite–pyrite ore zones are considered to be sedimentary facies of the iron-formation. The stratigraphic succession at the Helen, George McLeod, and Sir James mines consists of basal altered rhyolite volcanics that grade upward into massive or crudely banded beds of buff or black siderite with pyrite and pyrrhotite in the upper part. These grade upward into thinly banded chert, carbonate, and magnetite iron-formation, capped by andesitic pillow lava. The carbonate member ranges in thickness from 50 to 250 feet at Helen mine and although disrupted by numerous faults has been traced for miles along strike. The siderite–pyrite ore as mined contains about 33% iron, but after beneficiation in the sink-float plant and sintering, it contains 50 to 51% iron, 2.9% manganese, 11.4% silica, 3.32% lime, 7.37% magnesia, 1.84% alumina, 0.01% phosphorus, and 0.10% sulphur.

Magnetite–quartz facies predominate in most of the other Algoma type iron-formations of Ontario. The iron-formation at *Moose Mountain* (84) north of Capreol, Ontario, was first mined between 1901 and 1920 where the magnetite concentrate was briquetted and roasted in an attempt to provide a suitable high-grade product. The enterprise was not successful economically but it anticipated modern beneficiation methods for iron ore by 40 years. Several open-pit mines have been worked since 1959 on various fault segments of the iron-formation which averages about 33% iron. The pelletized concentrate contains 63 to 64% iron with less than 8.75% silica.

The *Adams* (60a) mine near Kirkland Lake, Ontario, has produced pelletized ore concentrate since 1964. Thinly laminated magnetite–quartz iron-formation with an average of 22% iron is mined from two pits in a structurally complex belt of iron-formation that extends for 6 miles. The pellets contain 62 to 63% iron and the rail haul of 700 miles to Pittsburgh where they are used is probably the longest for iron ore anywhere in the world.

The iron-formation at *Sherman* (95) mine in the Timagami area of Ontario is probably the least metamorphosed of any of the Algoma type. The magnetite–quartz beds average 25 to 30% iron and 1,200,000 long tons of pellets containing about 65% iron are expected to be produced annually.

Very highly metamorphosed iron-formations occur in Superior Province between Lake of the Woods, Lake Nipigon, Red Lake, and the Paleozoic rocks of Hudson Bay Lowland (Fig. V-10). Magnetite–quartz beds occur in garnet–biotite–quartz schists which are derived from Archean sedimentary and volcanic rocks. The Griffith mine is being developed in one of these iron ranges along the west shore of Bruce Lake; production commenced in 1968. It is expected that about 1,500,000 long tons of pellets containing 65 per cent iron will be produced yearly from magnetic taconite containing 30 to 33 per cent iron. Some of the important ranges in this area are the Lake St. Joseph, Bending Lake, Kapico, and Kowash ranges north of Nakina.



PLATE V-2

Ruth Lake mine, Labrador, Newfoundland. Hematite–goethite orebody in Knob Lake iron range, extending in background into Burnt Creek mine, Quebec.

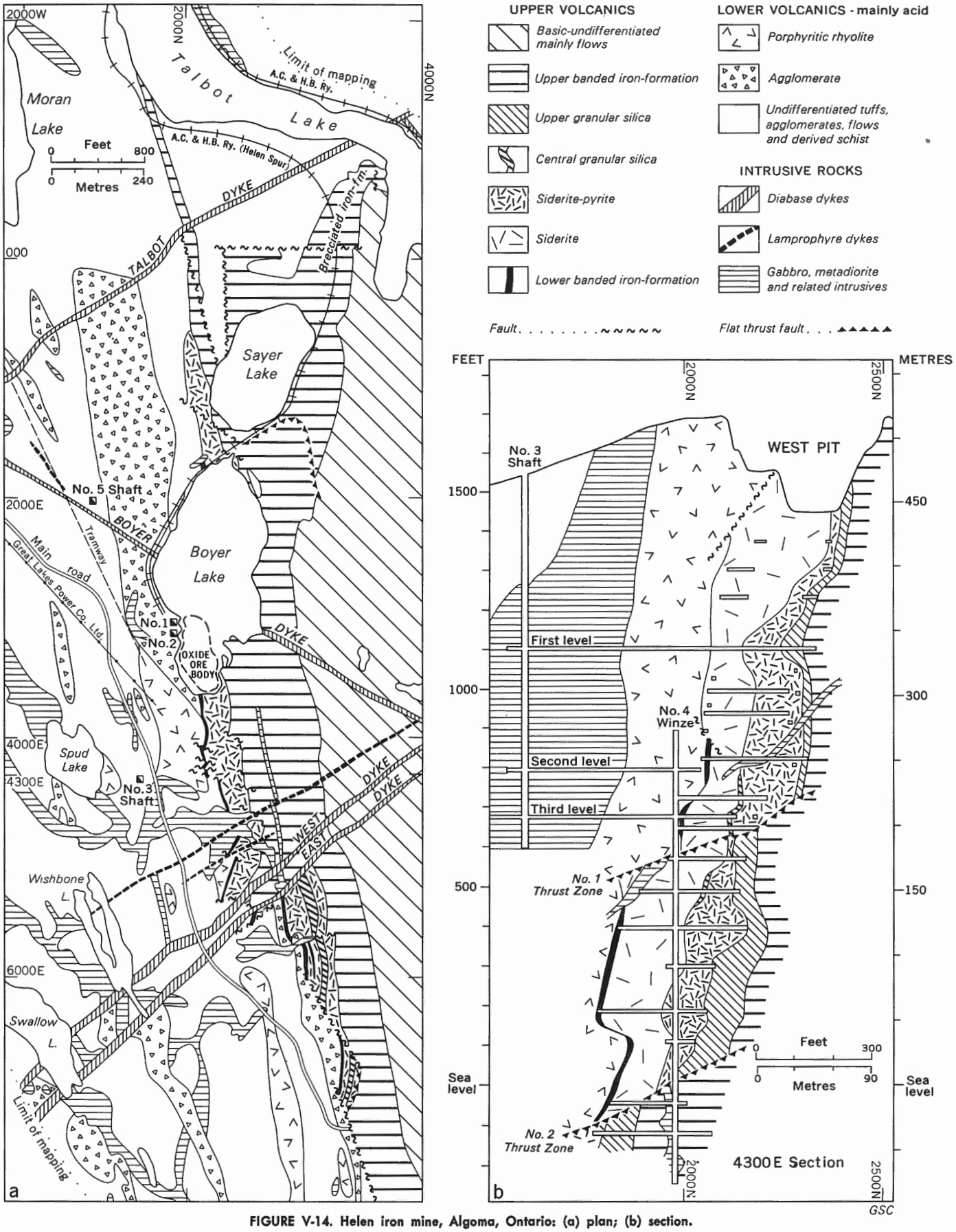


FIGURE V-14. Helen iron mine, Algoma, Ontario: (a) plan; (b) section.

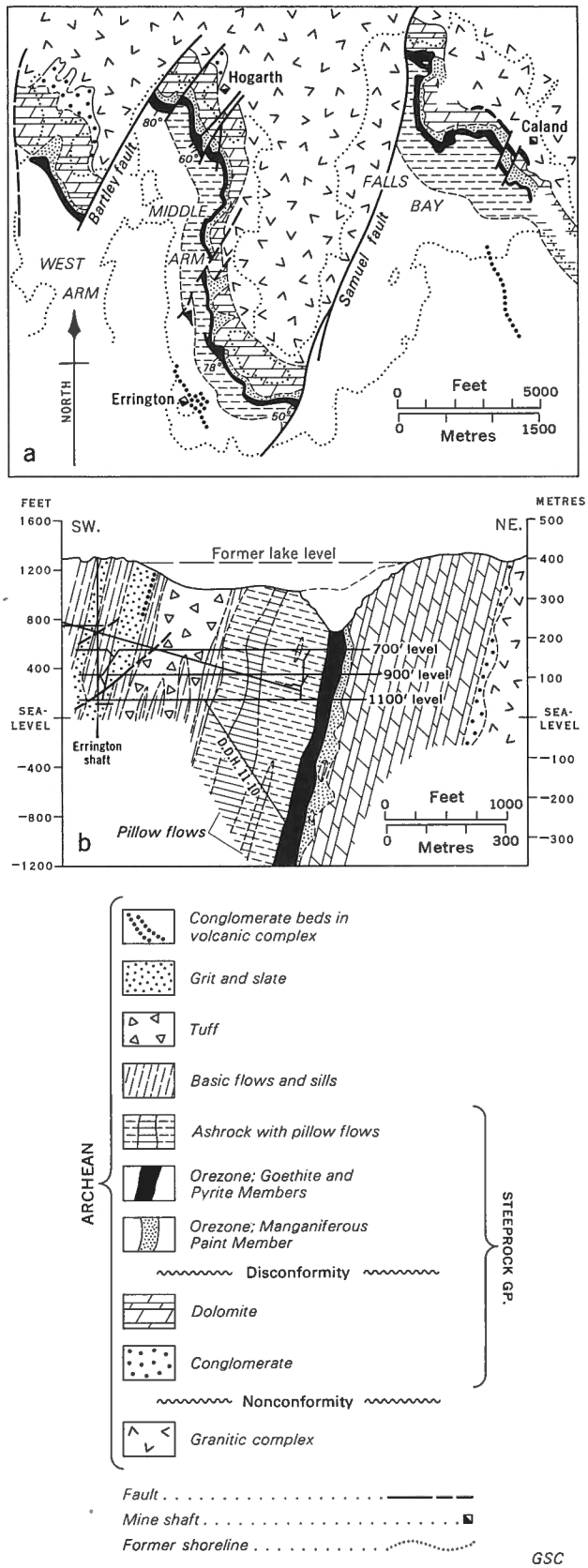


FIGURE V-15. Steeprock Lake area, Ontario (Jolliffe, 1966): (a) geological map; (b) section through Errington mine.

The hematite–goethite ore deposits of the *Steep Rock* (21) iron ranges are classified as Type IIIA. The tabular orebodies (Fig. V-15) are faulted segments of one stratigraphic zone. Jolliffe (1966) postulates that the interlayered goethite, minor pyrite lenses with graphite, and thinly banded chert were deposited on an old erosion surface where iron, in the form of colloidal oxide and hydroxide gels, accumulated in a manner akin to modern bog-iron deposits. The iron-rich bed was then covered by basic tuff, "ash rock". The rock succession was later folded and deformed in the Kenoran Orogeny. Oxidation and alteration took place in the ore zone during a late stage in its history, and part of the tuff and banded chert was leached and transformed to a clay or paint-rock that resembles many other altered iron-formations. This fact and other evidence suggest that the hematite and goethite were derived from siderite–pyrite and banded chert iron-formation similar to that at Michipicoten. The sediments were later transformed to ore by supergene enrichment processes. Neither hypothesis accounts fully for all the facts known, but new information from more extensive and deeper mine workings is helping to resolve the problem of ore genesis. Ore from the mines, before processing or pelletizing, contains about 55% iron, 5% silica, 1.5% alumina and lime combined, and 0.027% phosphorus. The dried ore contains about 60% iron.

Iron-formations in Churchill Province are of the Algoma type and most are highly metamorphosed. Very high-grade ore deposits have recently been discovered in northern Baffin Island near Milne Inlet. The deposits consist of zones of metamorphosed magnetite and hematite iron-formation, unusually rich in iron, that were further altered by secondary enrichment processes and then subjected to later metamorphism. Two kinds of ore are present: hard, dark grey to blue hematite and high-grade stratified bodies of massive or banded magnetite and specular hematite. Large reserves of potential ore containing 67 to 69% iron and less than 1% silica have been proven.

Iron Deposits of Southwest Grenville Province

Numerous deposits associated with anorthosite and gabbro (Types IVA and B) are present in Grenville Province. Some of these, like the famous large orebodies in anorthosite at *Lac Tio* (73), Quebec, are predominantly ilmenite and the ore is processed for titanium dioxide. More than one million tons of iron are produced each year as a byproduct. Numerous large low-grade bodies of titaniferous magnetite in gabbroic rocks are not mined because an economic metallurgical process for separating titanium from the iron minerals has yet to be developed.

Numerous contact metasomatic magnetite deposits (Type V) occur in the southern part of Grenville Province. These deposits are relatively small and many were worked during the 19th century. The *Marmora* (111b), Ontario, deposit was discovered beneath 125 feet of Paleozoic limestone by an aeromagnetic survey. The mass of

ore is 2,400 feet long and 400 feet wide and contains an average of 40% iron. It has been mined since 1955 and about half a million tons of pellets containing 65% iron are produced yearly. A similar orebody has been worked at *Hilton* (42) near Shawville, Quebec, since 1958. Nearly a million tons of pellets containing 66% iron are produced yearly from crude ore grading about 18% iron. Deposits of this type are related spatially to gabbro intrusions. The limestones and amphibolites are the most favourable host rocks.

Nickel

Canada has been a world leader in production of nickel since exploitation of the huge Sudbury deposits commenced in 1887. In 1966, Canada produced more than 70 per cent of the non-communist world's nickel with the year's production valued at $400 million. More than 99 per cent of the total Canadian production of 234,061 tons was derived from mines in the Canadian Shield. Since 1953, nickel discoveries have been made in other parts of the Shield, at Thompson, Manitoba, in Ontario, in Quebec, and at Rankin Inlet, District of Keewatin. Ore reserves in 1966 were in excess of 380 million tons, averaging 1.5 per cent nickel. Despite the large tonnages now mined, these reserves are being increased annually, a trend that is expected to continue for some time as new discoveries are made, and as technological improvements and increases in metal prices enable presently known reserves of marginal grade ore to be mined.

The orebodies in the Canadian Shield are all of the sulphide type. They may be considered primary, as they are little altered by surface weathering processes. The principal ore mineral is pentlandite, almost invariably associated with pyrrhotite. Other sulphides of nickel, copper, and iron are commonly present, their species and relative proportions being dependent both on the bulk composition and on the temperature-pressure environment within the orebody during its formation. All important nickel deposits in the Canadian Shield are located in or near magmatic rocks of mafic or ultramafic composition, which were the source of the nickel. Copper is generally present and the ratio of nickel to copper has been found to be relatively consistent for a given deposit. Also the nickel–copper ratios have shown a direct correlation with the magnesium content of the associated mafic or ultramafic rock. Thus, sulphide minerals associated with dunite and peridotite have higher nickel to copper ratios than those associated with gabbro and norite. In the descriptions that follow, the nickel deposits of the Canadian Shield are grouped on the basis of their mafic or ultramafic affiliation.

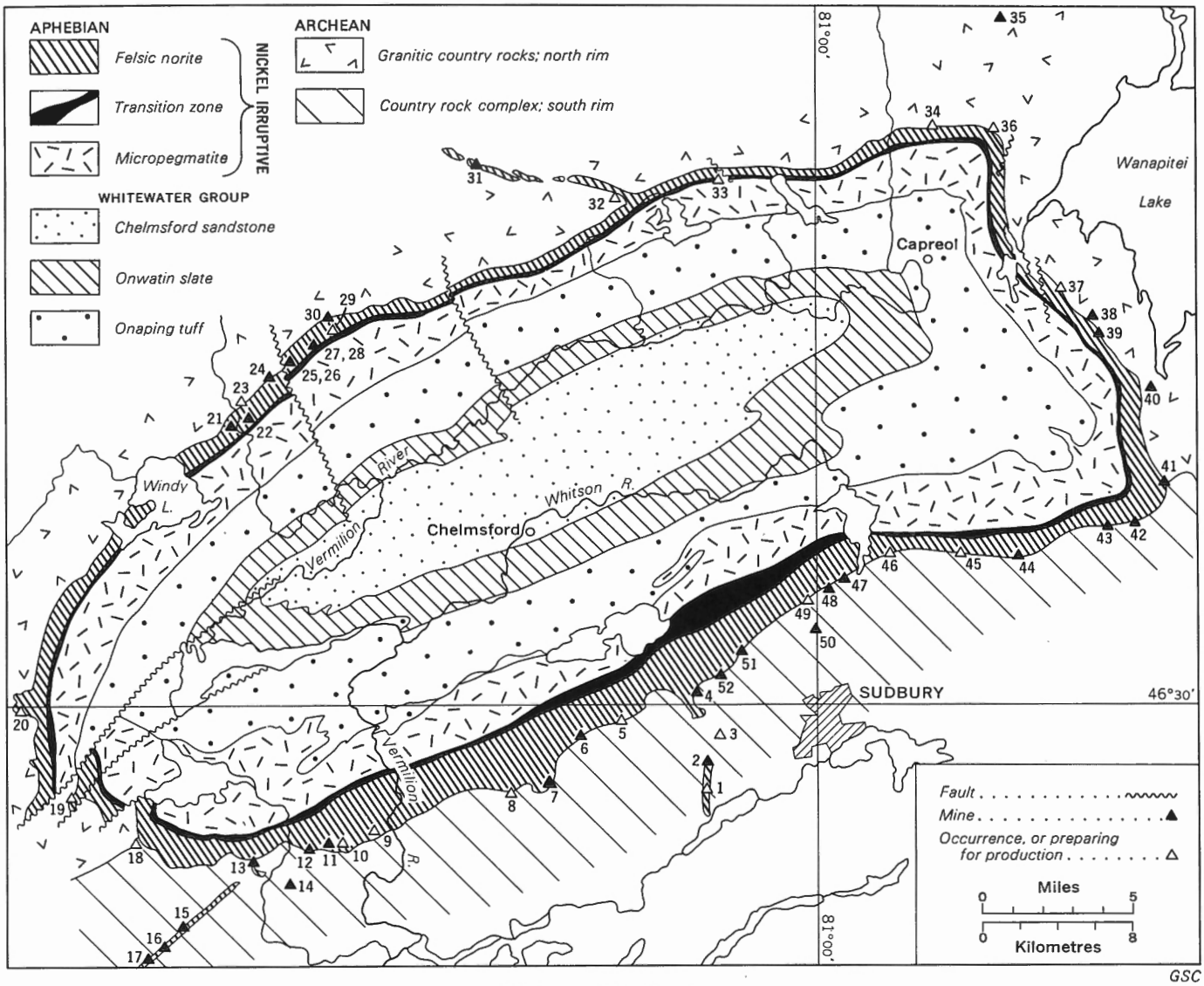
Deposits of Mafic Affiliation

Sudbury district, Ontario (78 to 90). Nickel ores affiliated with mafic rocks are typified by the deposits of the Sudbury district. About 185,000 tons, or more than 70 per cent of Canada's total annual nickel production, is derived from this district. A comparable tonnage of copper is also extracted, which makes Sudbury the largest copper producing area as well. The nickel sulphide orebodies occur at or near the outer margins of a body of igneous rock known as the Sudbury Irruptive. In plan, the irruptive forms an elliptical basin, 37 miles long and 17 miles wide (Fig. V-16). The outer part is composed of coarse-grained felsic norite and the inner part is a granophyre. The outer contact of the norite is marked by a zone of shearing and brecciation. The dip varies from 35 to 45 degrees south on the north side of the basin to vertical on the south (Fig. V-17). The rocks inside the irruptive, the Aphebian Whitewater Group, consist of tuffs, slate, and sandstone. They are generally synclinal but are broken by several east-west striking faults. Older rocks north of the irruptive are principally Archean granite. The southern contact of the irruptive is with the Sudbury Group, a sequence of metamorphosed and locally brecciated sediments and volcanics of Aphebian age intruded by granites and gabbros of various ages. The irruptive itself is metamorphosed, locally metasomatized, faulted, brecciated, and transected by diabase dykes. The origin of the Sudbury Irruptive and its associated ores is still a matter of conjecture. Cases can be made for the irruptive being a spoon-shaped sill, a lava lake remnant, a ring-dyke, a funnel-shaped intrusion, or even a structure produced by meteorite impact. The presence of abundant inclusions of ultramafic rock in many of the massive sulphide zones points to a deep-seated source. This and other less direct evidence suggest that some form of root or feeder exists beneath the Sudbury Irruptive.

The Sudbury ores consist of massive sulphides, breccia sulphides, and disseminated or stringer sulphides in silicates. The most abundant sulphide minerals are pyrrhotite, pentlandite, chalcopyrite, and cubanite. A host of less common sulphides and arsenides is the source of such byproduct metals as those of the platinum group, gold, silver, and cobalt. The overall ratio of nickel to copper is very close to 1:1, but in some orebodies there are significant departures, generally in the direction of increasing nickel.

The most commonly accepted theory on the origin of the sulphide minerals is that they were present in the main felsic norite magma and segregated as fluid droplets from the silicate fraction, settling toward the lower contact, to form pools of ore in depressions in the basement. In places, subsequent tectonic activity caused the sulphides to migrate into the adjacent country rocks, forming orebodies offset from the main irruptive bodies. Some recent observations (Souch and Podolsky, 1968; Naldrett and Kullerud, 1967) suggest that this theory needs revision. A complex group of intrusions and intrusive breccias occurs as a semi-continuous layer along the base of the main felsic norite and, in places, projects as dykes into

the footwall. These "sublayer" rocks, which vary in composition from quartz diorite to mafic norite, contain or lie adjacent to all known orebodies in the Sudbury district. They contain abundant inclusions and are intrusive into the felsic norite. Whole rock analyses indicate that the "sublayer" rocks contain about fifty times as much nickel and copper as the felsic norite. The implication is that the ores were deposited during a later intrusion of sulphide-rich silicate magma along the basal contact of the irruptive.



INDEX TO PROPERTIES

(Numbers in brackets according to Map 1252A)

1. Copper Cliff South
2. (89b) Copper Cliff No. 1
3. (89c) Copper Cliff North
4. (89a) Clarabelle
5. Tam O'Shanter
6. (90b) North Star
7. (90a) Creighton
8. Gertrude
9. McVittie Graham
10. Lockerby
11. (78b) Ellen
12. (78a) Crean Hill
13. (78c) Victoria
14. Vermilion
15. Kidd
16. (79c) Worthington
17. (79b) Totten
18. Chicago
19. Sultana
20. Trillabelle
21. (80a) Hardy
22. (80b) Onaping-Boundary
23. Levack West
24. (82b) Levack
25. (82a) Fecunis Lake
26. (82d) North
27. (82e) Strathcona
28. Coleman
29. Longvack South
30. (82c) Longvack
31. (83) Nickel Offset
32. W.D. 150, 155
33. North Range Shaft
34. New Dominion
35. (85) Milnet
36. Whistle
37. Capre Lake
38. Victor
39. (86) Nickel Rim
40. (87e) Maclennan
41. (87f) Norduna
42. East
43. (87a) Falconbridge
44. (87c) Garson
45. (87d) Kirkwood
46. Sheppard
47. (88a) Blezard
48. (88c) Mount Nickel
49. Little Stobie
50. (88b) Frood-Stobie
51. (89g) McKim
52. (89h) Murray

FIGURE V-16. Nickel deposits of the Sudbury district, Ontario (by J. A. Chamberlain).

Lynn Lake, Manitoba (1). The mines at Lynn Lake annually produce about 13,000 tons of nickel, or approximately 5 per cent of Canada's total. The orebodies, ten in all (of which three have been mined out), occur mainly in a variety of intrusive rocks of gabbroic character, but one orebody is in a more siliceous, fine-grained, variable quartz hornblende diorite. Peridotite is present in the vicinity of all orebodies, and sometimes close to them, and although some ore minerals are present in the peridotite at one locality, no orebodies have been developed. The sulphide minerals occur as massive or breccia type concentrations, disseminations, or stockworks. The main ore minerals are pyrrhotite, pentlandite, and chalcopyrite. Pyrite is minor. According to Ruttan (1955), the nickel–copper ratios range from 1:1 to 3.5:1 in the various orebodies, and average about 2:1. No significant change in ratio has been observed with depth. The orebodies are irregular to lenticular in plan, and pipe-like in section. They occur in zones that show a high incidence of fracturing and faulting. Some of the faulting is later than deposition of the ore. The sulphide minerals replace granite, pegmatite, feldspar porphyry, felsite, and quartz veins that cut the gabbros. A considerable time must have elapsed between emplacement of the mafic rocks and the period of mineralization. This is incompatible with formation of sulphides in situ by processes of magmatic segregation.

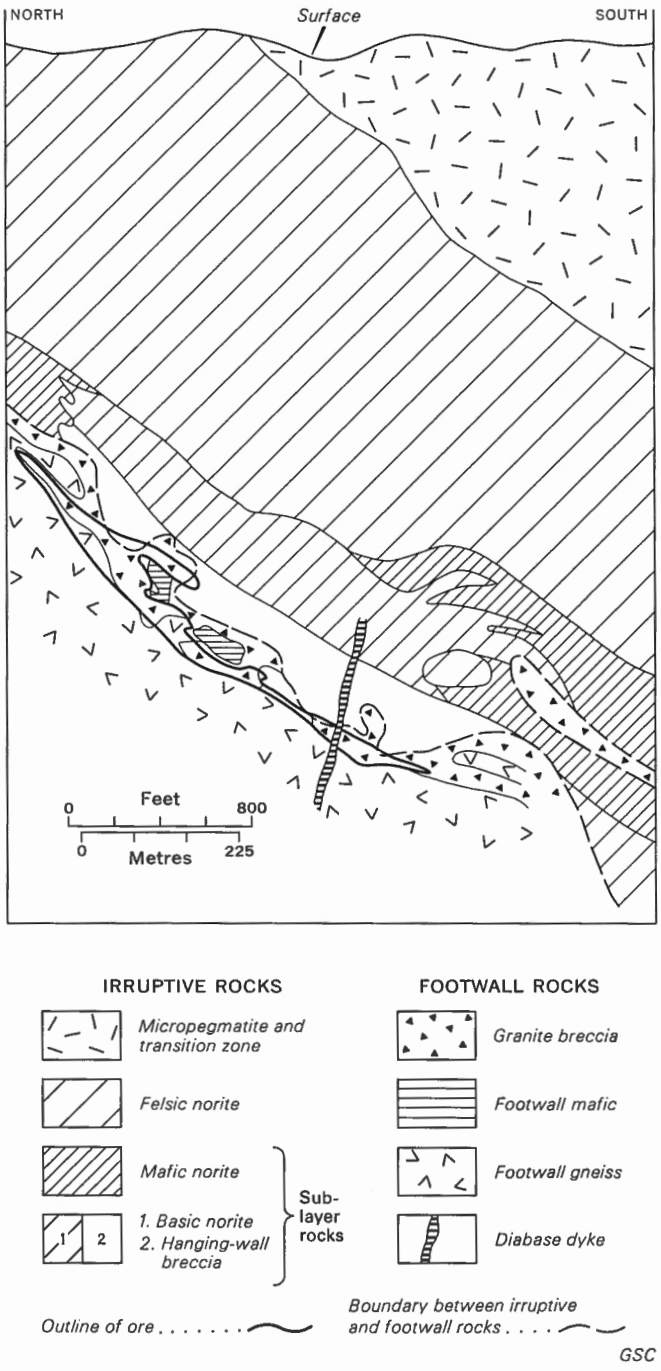


FIGURE V-17. Vertical section through the Strathcona deposit, north rim of the Sudbury basin, Ontario (Naldrett and Kullerud, 1967).

Other nickel deposits associated with mafic rocks are fairly common in the Canadian Shield, but few have proved to be economic. In 1965, the *Lorraine* (40) mine in the Belleterre area, Quebec, produced 900 tons of nickel and 2,200 tons of copper. The orebody is small, steeply dipping, and pipe-shaped and is present in Archean greenstones near a diorite intrusion. The ratio of nickel to copper in the sulphides is fairly consistent at 0.4:1. The *Kenbridge* deposit in the Kenora district, Ontario, occurs in a gabbro lens intrusive into greenstone. The sulphides occur in a series of flat sheets that have an average nickel to copper ratio of 2:1. In *Pardee Township,* south of Port Arthur, Ontario, pyrrhotite, chalcopyrite, and pentlandite occur along the lower margin of sill-like bodies of gabbro that intrude Aphebian sediments. Recent drilling has outlined large tonnages of low-grade sulphide ore occurring largely as disseminations and veinlets in coarse-grained gabbroic matrix. A uniform nickel–copper ratio of about 0.1:1 is indicated.

Deposits of Ultramafic Affiliation

Thompson district, Manitoba (12, 13). The largest nickel deposits associated with ultramafic rocks are those of the Thompson belt, in Manitoba, that were discovered in 1956 (Fig. V-18). The present annual nickel production from the Thompson mine of 50,000 tons accounts for nearly one fifth of Canada's total. Three other mines, the Pipe, the Soab, and the Birchtree, are currently being prepared for production. The orebodies of the Thompson belt are present for 90 miles along the boundary between Superior and Churchill Provinces. The deposits are very nearly colinear and occur in or near lenticular pods of Alpine-type peridotite within metasediments and gneisses. The peridotites are commonly intruded by dyke-like bodies of granite and pegmatite, and the sulphide minerals in places invade these younger intrusive rocks. The nickel ore is present in two principal forms: (1) as uniformly disseminated sulphides or sulphide veinlets in peridotite, and (2) as sharply defined zones of breccia sulphide in metasedi-

mentary rocks. Orebodies of the first type occur at Mystery Lake, Moak Lake, Pipe Lake, Grass Lake, Soab, and near Wabowden. They are relatively low grade and appear to be indigenous to the peridotites, being regarded by most geologists as magmatic in origin.

Orebodies of the second type occur at Birchtree Lake and Hambone Lake, and are best exemplified by the Thompson deposit (Fig. V-19). There, they occur in a continuous sheet that is grossly conformable to a large anticline and syncline in the metasediments and gneisses. The sulphides on the east limb of the anticline occur more or less continuously for more than 15,000 feet along strike within a narrow band of biotite–sillimanite schist. Although the schist thins and disappears on the nose and west limb of the anticline, the sulphides persist along the contact between quartzite and skarn rock. West of the main anticline the sulphides have been traced to a schist-encased peridotite lens (Zurbrigg, 1963). In places, the peridotite itself contains sufficient disseminated sulphide minerals to be considered as ore.

Conclusions concerning the origin of the Thompson ore are summarized as follows:

1. The continuity of the sulphide horizon suggests a strong structural control on the ore localization.
2. The close spatial relationship of nickel-rich peridotite to all ore zones in the Thompson belt suggests that the nickel is derived from these rocks. This is corroborated by the fact that the ores have high nickel–copper ratios, in excess of 15:1 regardless of whether they occur in peridotite or in the metasediments.
3. As the sulphide minerals clearly replace pegmatite which is intrusive into the peridotite, the formation of ore zones in the country rocks is considered to have taken place after emplacement of the peridotites.

The mechanism by which the nickel sulphide was transported away from the peridotite and deposited as a continuous sheet over many thousands of feet is not clear. As there is no evidence of wall-rock alteration, transportation via hydrothermal fluids is unlikely. On the basis of current knowledge, the most plausible explanation is that, as the temperature and pressure increased during a period of folding and regional metamorphism, a fluid consisting essentially of iron, nickel, and sulphur developed in the peridotite. This fluid seeped down a shear formed in the relatively incompetent schists or at the contact between the quartzite and skarn where it ultimately congealed to form the orebodies.

At the *Marbridge* (27) mine, Quebec, the ore is localized in massive, acidic tuff adjacent to a lens of altered ultramafic rock. Currently producing about 2,300 tons of nickel annually, this small but high-grade deposit is believed to have formed by multiple injections of a silicate–sulphide fluid derived from the peridotite (Clark, 1965). The nickel–copper ratio in these ores is about 30:1.

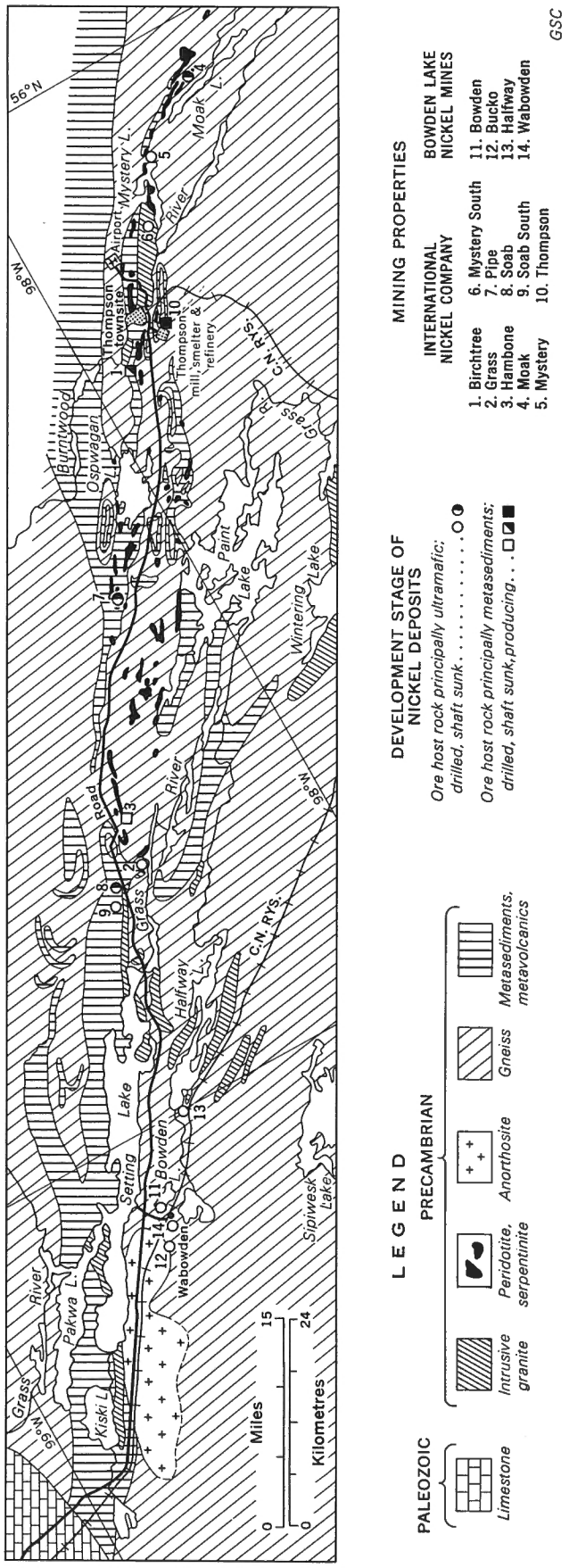


FIGURE V-18. Geology of the Thompson nickel belt, Manitoba (compiled by J. A. Chamberlain).

Gordon Lake, Ontario (13). The mine presently produces 2,800 tons of nickel annually. The sulphide minerals occur as disseminations, stringers, or flat-dipping tabular masses in or adjacent to peridotite in the footwall of a prominent fault. Lenticular or dyke-like pegmatitic intrusions transecting the peridotite are commonly mineralized. The sulphide assemblages in both the peridotite and pegmatite are much the same, indicating a common origin. The nickel–copper ratio in these ores is approximately 2:1.

The Cape Smith belt in northern Quebec contains several small, high-grade nickel–copper deposits. Although the remoteness of the region has hindered development, one orebody of 3.5 million tons is estimated to contain 3 per cent of combined nickel and copper at a ratio of 5:1. These sulphides occur as replacements at the base of sill-like bodies of altered peridotite (Philpotts, 1961). Between 1957 and 1962, a similar deposit at *Rankin Inlet* (11), District of Keewatin, produced more than 10,000 tons of nickel and 2,900 tons of copper. Other nickel deposits of ultramafic affiliation that attained production for short periods include the *Alexo* (54) and *Trebor* mines in the Timmins and Timagami areas of Ontario, respectively. The total production of nickel from these two properties was less than 200 tons.

Cobalt

All Canadian cobalt is obtained from the Shield, the greatest part as a byproduct of the nickel–copper ores at Sudbury, Ontario, and Lynn Lake, Manitoba. Pentlandite, the main source of cobalt in these ores, has an average cobalt content of about 1 per cent (Hawley, 1962). Minor amounts were produced during refining of silver ore from the Cobalt and Gowganda areas of Ontario. Total production since 1955 has remained constant at about 1,500 to

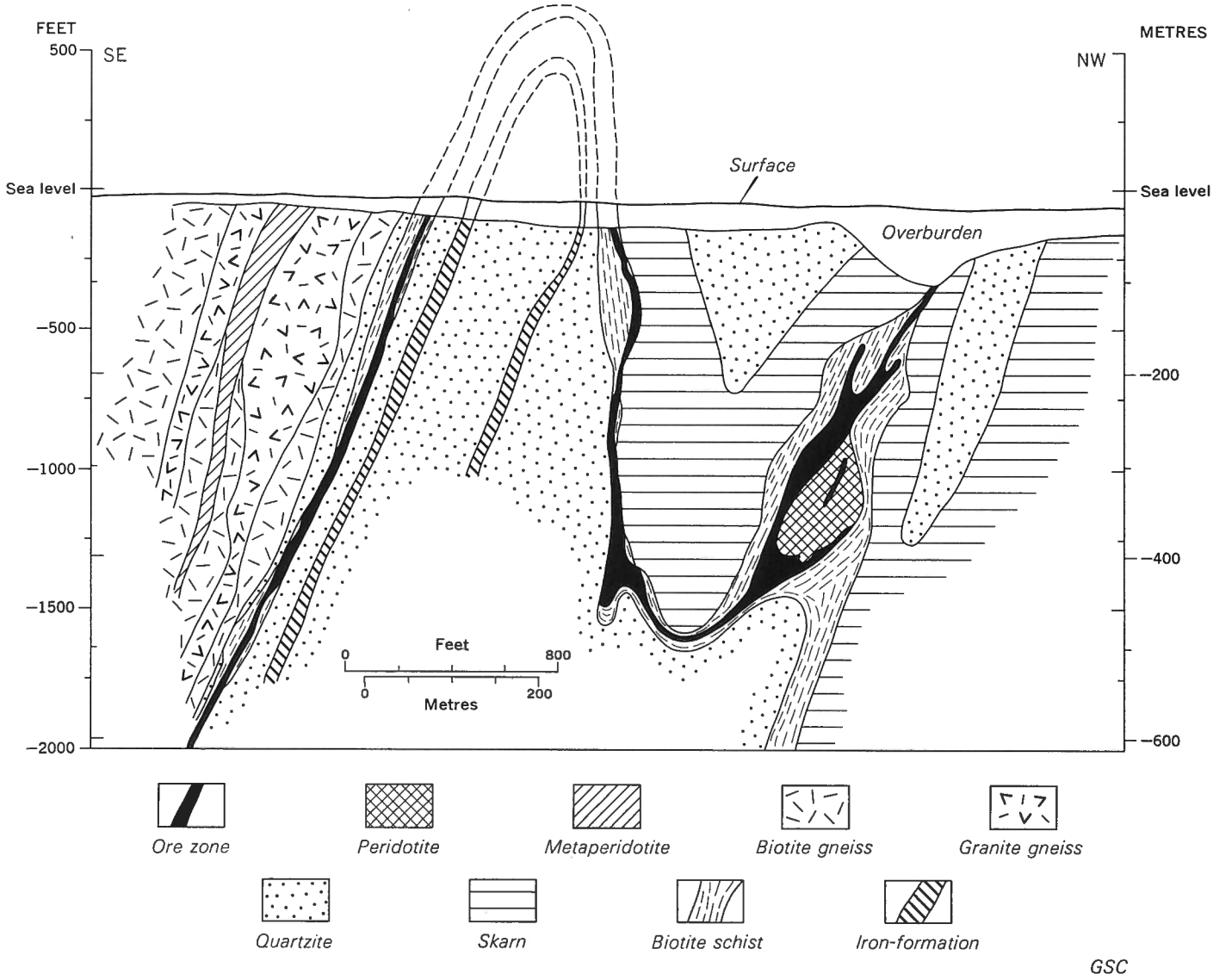


FIGURE V-19. Distribution of ore at the Thompson mine, Manitoba (Zurbrigg, 1963).

1,800 tons per year; 1966 production was 1,755 tons, about 7 per cent of which came from the Cobalt and Gowganda areas. Cobalt production in Canada has had a varied history, linked until recently with the fortunes of the Cobalt silver camp. Cobalt ores were mined from that area, and salvage operations were carried on during inactive periods. In 1951, the federal government raised the price for cobalt contained in cobalt ores or concentrates in order to stimulate exploration and production in the Cobalt and Gowganda areas. This premium price plan was terminated in 1956, the same year that the United States ended the price support in effect during the Korean war. No cobalt ores have been produced as such in Canada since then, although byproduct recovery from silver ores has continued. Shipping of byproduct cobalt from nickel mines began from Sudbury in 1954, and from Lynn Lake in 1955.

Chromium, Chromite

No chromite is being produced from the Shield. In the years 1934–37 a few thousand tons were mined from a deposit near Obonga Lake, Thunder Bay district of Ontario. There a mass of serpentinite composed of serpentine, chlorite, and talc in varying proportion underlies an area about 3 miles long and three-quarters mile wide. Vein-like bodies of chromite occur near the eastern end of the mass that carry an average of 27 to 34 per cent Cr_2O_3. The best concentrate that could be made of this material carried 42 per cent Cr_2O_3 and had a chrome–iron ratio of 1.3:1. In 1942 a large, low-grade chromite deposit was discovered north of Bird (Oiseau) River in southeastern Manitoba. The deposits have received intermittent attention but there has been no commercial production. A great sill-like mass of gabbro is intrusive into Archean rocks. It has been differentiated, presumably by gravity, into products ranging from augitite at the base to feldspathic quartz diorite at the top. Slightly below the middle of the sill small grains of chromite form a band that parallels the strike and dip. The chromiferous material occurs in alternate narrow bands of high- and low-grade material. The main band averages 7 feet wide, and has been traced several miles. The average grade is reported to be about 26 per cent Cr_2O_3, and the best chrome–iron ratio is 1.4:1.

Titanium

Titanium is the eighth most abundant metallic element on earth. It is estimated to average about 0.4 per cent by weight of rocks of the earth's crust, and it probably forms about 0.5 per cent of the rocks of the Canadian Shield, where it occurs mainly as the dark iron–titanium oxide minerals. It is about one-tenth as abundant as iron, and much more abundant than the common base metals, copper, nickel, lead, and zinc. Titanium is used on an increasing scale throughout the world in the white oxide form as a pigmenting material. Since 1950 Canada has become a leader in the production of ilmenite from the Lac Tio deposit near Allard Lake and TiO_2-rich slag is recovered at the smelter at Sorel, Quebec. In 1957 Canada became the world's leading ilmenite producer; by 1966 production reached more than a million tons per year. About 98 per cent of the titanium ore mined in Canada has gone into the production of TiO_2 pigment. Some is used as heavy aggregate in the concrete employed in laying natural gas pipelines and to form nuclear reactor shields. Only a small amount of titanium metal has been produced as yet; production may improve as use of metal increases, particularly in the construction of supersonic aircraft.

In the Canadian Shield, titanium occurs as an ilmenite–hematite mixture and as titaniferous magnetite (Pl. V-3). It is associated with large anorthositic and gabbroic intrusions that range in age from Archean in Superior Province, to Helikian in Grenville and Nain Provinces. This close association of ilmenite–hematite with anorthosite, and of titaniferous magnetite with gabbro, is also typical of many other parts of the world. Nowhere, however, is there a larger or more widespread development of anorthositic rocks, with associated gabbroic rocks and titaniferous deposits, than in Grenville and Nain Provinces.

Ilmenite Deposits

Ilmenite–hematite (titanic iron) deposits were first found in Canada by French colonists west of *St. Urbain* (70) in 1666. The first attempts to mine them were financially unsuccessful, but later attempts during World Wars I and II were more rewarding. Less than a million tons of ilmenite have been produced from the Furnace, General Electric, Coulombe, and Bignell deposits that contain about 22 million tons of ore averaging about 22 per cent titanium and 35 per cent iron. The deposits are dyke-like bodies composed almost entirely of massive, coarsely crystalline, blue-black intergrowths of ilmenite and exsolved hematite. The ore dykes, 10 to 100 feet wide and 100 to 1,000 feet long, are intrusive into the anorthosite on the hill west of St. Urbain. The deposits have been intersected by diamond drill at depths of more than 1,000 feet. Minor rutile pyrite, zeolites, calcite, and sapphirine occur sparingly, as well as inclusions of altered anorthosite. Some biotite is developed at the contacts of the ilmenite dykes with the anorthosite. Rutile is most abundant in the General Electric deposit where it is concentrated in a calcite segregation within the ilmenite–hematite lode. Many ilmenite deposits in the Morin anorthosite are similar to those at St. Urbain. The *Ivry* deposit has yielded about 52,000 tons of ilmenite since 1910. High-grade ore averages about 20 per cent titanium and 40 per cent iron. It consists predominantly of ilmenite with exsolved hematite, and minor pyrite, chalcopyrite,

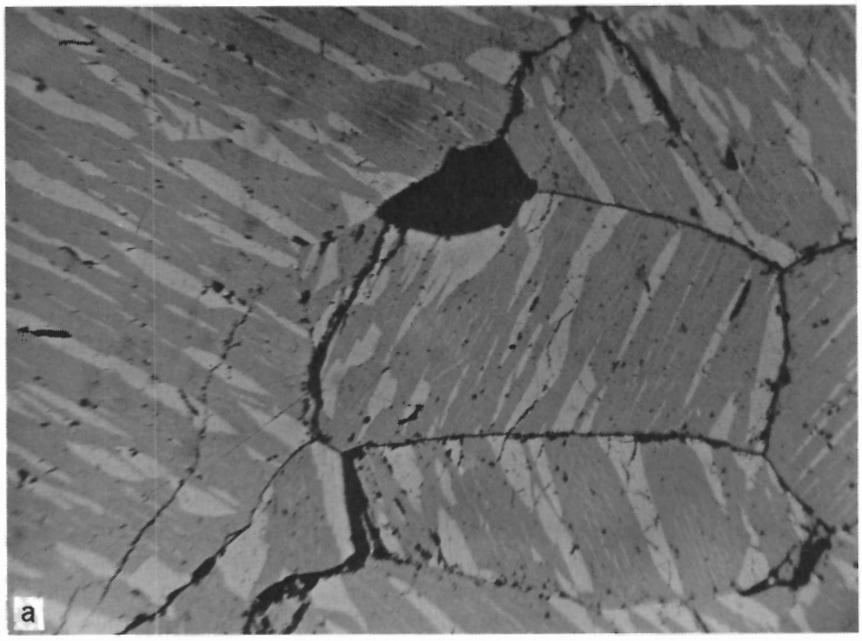
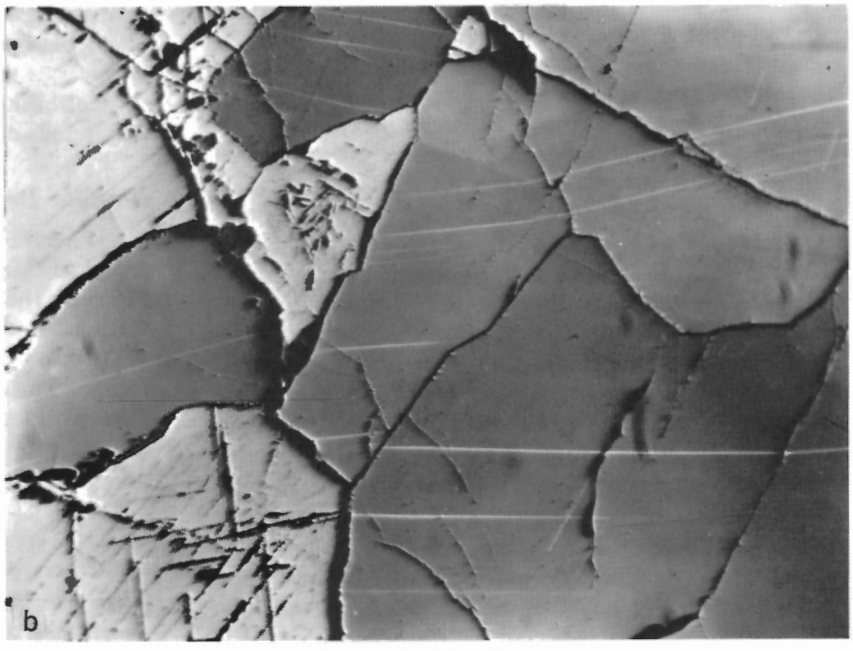


PLATE V-3
PHOTOMICROGRAPHS OF POLISHED SURFACES OF TITANIUM ORES.

(a) Lac Tio ilmenite–hematite deposit, Allard Lake, Quebec, showing hematite exsolution lamellae (light) in ilmenite (dark). Parallel nicols, low power.



(b) Titaniferous magnetite, south of Mattawa, Ontario, showing ilmenite grains (smooth, grey) with blades of exsolved hematite (white) and grains of titanomagnetite (rough, light grey) with exsolved ilmenite (dark). Crossed nicols, low power.

and magnetite. Mineralization is intermittent within a zone 100 feet wide and 1,000 feet long, and probably extends to depths of several hundred feet.

A large deposit of massive ilmenite–hematite was discovered in anorthosite west of Allard Lake, at *Lac Tio* (73). This deposit contains more than 125 million tons of ilmenite. The orebody forms a saddle and gently dipping sill-like mass, 25 feet to more than 200 feet thick, that intrudes anorthosite on the west, and is capped by gabbroic anorthosite on the east. The deposit has several dyke-like tongues and satellitic deposits associated with it. The ore consists almost entirely of massive, coarsely crystalline ilmenite with exsolved hematite (Pl. V-3a), averaging about 20.4 per cent titanium and 41 per cent iron. Inclusions of anorthosite, and crystals of plagioclase, pyroxene, pyrite, and chalcopyrite form the gangue or contaminate the orebody. About 10 million tons of ore have been mined up to 1967 in open-pit operations, from which about 5 million tons of TiO_2 slag and 3½ million tons of iron (Sorel metal) have been recovered.

Titaniferous Magnetite Deposits

Titaniferous magnetite deposits form a large undeveloped potential source of titanium and iron. More than

50 titaniferous magnetite deposits are known in the Grenville Province of Quebec alone, and many others are known elsewhere. Information is lacking regarding the full extent and grade of these occurrences, but they appear to range in size from several hundred thousand tons to more than a billion tons, and to average, conservatively, about 20 per cent iron, 5 per cent titanium, and 0.25 per cent vanadium. Representative of the better known titaniferous magnetite occurrences are the Seine Bay–Bad Vermilion Lake deposits in northwestern Ontario, the St. Charles and the Sept Iles Bay deposits of Quebec, and the Steel Mountain and Indian Head deposits of western Newfoundland. The Newfoundland deposits occur in Precambrian inliers within the Paleozoic rocks of the Appalachian Region. All deposits are associated with anorthositic and gabbroic rocks. This is also true of the less well known, recently discovered, titaniferous magnetite deposits around Lac Pin Rouge north of Montreal, at Magpie Mountain (Awater–Lapointe) north of Mingan, Quebec, and at Pipestone Lake near Cross Lake, Manitoba. Large tonnages of low-grade material are available in all these deposits, and many others undoubtedly remain to be discovered.

The main ore minerals, titanomagnetite, ferrian ilmenite, and titanhematite, commonly show exsolution textures characteristic of deposits formed at magmatic temperatures (Pl. V-3b). Because of the intimacy of the intergrowths it is commonly difficult, or impossible, to separate them satisfactorily by mechanical methods. Most deposits, even the low-grade ones may be readily concentrated magnetically, especially where the ilmenite and titanomagnetite occur as discrete grains. Most titaniferous magnetite deposits carry traces of vanadium, chromium, manganese, cobalt, and nickel. More than one per cent vanadium has been detected in the deposits at Pipestone Lake, Manitoba, at Millbridge and Mattawa, Ontario, and from Indian Head, Newfoundland.

Other potential ores of titanium in the Canadian Shield include occurrences of rutile, titanite (sphene), and perovskite, but none has yet been found in deposits of economic importance. Concentrations of heavy dark minerals in sands of Recent age may carry all these minerals and be potential ores of iron, titanium, or other rare metals, as they are in Australia, India, the United States, and Brazil. Notable concentrations of black magnetic sands have been found at localities along the north shore of the St. Lawrence River.

Locating Deposits

Aeromagnetic surveys have proved to be of value in locating concentrations of magnetic minerals, and also help to outline the gabbroic phases of anorthositic massifs. The gabbroic phases generally have a much higher magnetic intensity than the anorthosite proper, and the titaniferous magnetite concentrations within them are commonly associated with very strong positive aeromagnetic anomalies. On the other hand, the ilmenite–hematite occurrences that are commonly associated with the anorthositic phases are also generally represented by negative magnetic anomalies, as at Lac Tio in which the polarity of the deposit is reversed, with the north-seeking pole inclined upwards. However, the typical magnetic intensities may be muted by mixtures of the two types of minerals; furthermore, ilmenite–hematite occurrences are normally (positively) polarized and associated with positive aeromagnetic anomalies, as at St. Urbain, and titanomagnetite is in places reversely (negatively) polarized, as at Magpie Mountain. Ground magnetometer, gravimeter, and dip-needle surveys, are also used to advantage in outlining titaniferous deposits after they have been located or their presence suspected.

BASE METALS

Copper and Zinc

Copper produced from the Canadian Shield in 1966 was valued at $464 million, and formed 73 per cent of the total Canadian production of this metal; similarly, zinc was valued at $285 million and formed 45 per cent (Table V-7). About half the copper came from nickel–copper mines; these are described in the section on nickel. Most of the remaining copper and almost all the zinc (Tables V-8 and -9) came from lenticular, massive to disseminated, strata-bound deposits in metavolcanic and metasedimentary rocks. These deposits contain mixtures, in various proportions, of pyrite, pyrrhotite, sphalerite, chalcopyrite, and, in some instances, galena. They also contain significant amounts of gold and silver. Zinc is not recovered from some deposits, nor is copper recovered from others, even though the deposits are all essentially of the same type. Most production has been from the Abitibi belt in Superior Province, which extends from Timmins, Ontario, to Matagami and Chibougamau, Quebec (Fig. V-20). Important production from strata-bound deposits is also obtained in the Manitouwadge area of Ontario, and from the southern part of Churchill Province at Flin Flon, Manitoba, and within the surrounding district. Several copper–gold deposits in the Chibougamau region, Quebec, which is in Superior Province but east of Matagami, occur in basic plutonic rocks. A copper deposit mined in the northern part of Chibougamau region is in Aphebian dolomite.

Smaller amounts of either copper or zinc, or both, are produced from other deposits of various kinds includ-

TABLE V-7 *Estimated distribution of copper and zinc production in 1966 (based on preliminary figures for 1966, including some assumptions because of different ways of reporting)*

	COPPER			ZINC		
	Pounds (millions)	Per cent of Canadian Production	Per cent of Production from Shield	Pounds (millions)	Per cent of Canadian Production	Per cent of Production from Shield
Total Canadian production	1,033			1,889		
Production from Canadian Shield	751	73		862	45.5	
Production from nickel–copper mines	337	33	45			
Production from non-nickel mines	414	40	55			
Superior Province	322.5			709		
Timmins–Noranda–Matagami (by difference)	*129.5*	12	17	*571*	31	66.5
Manitouwadge	*60*	6	8	*125*	6.5	14.5
Chibougamau	*102*	10	13.5			
Other mines	*31*	3	4	*13*	0.5	1.5
Churchill Province	90	9	12	140	7	16
Grenville Province				13	0.5	1.5
Slave Province	1.5	<0.5	<0.5			

ing strata-bound ones in the Superior and other provinces of the Shield. These include a porphyry copper deposit at Timmins, Ontario, and a few chalcopyrite-bearing quartz veins near Kirkland Lake, Ontario, the Zenmac sphalerite-bearing lenses in gabbro north of Lake Superior, the North Coldstream copper deposits in basic volcanics near Port Arthur, the Copperfields deposits near a rhyolite–diorite contact at Timagami Lake, and chalcocite-bearing veins in Keweenawan strata near Sault Ste. Marie (Tables V-8 and -9). Although not at present mined, a large low-grade copper dissemination in a basic phase of an alkalic syenite complex occurs near Marathon, Ontario. Zinc is produced from a sphalerite deposit in metamorphosed impure limestone in Grenville Province northwest of Ottawa. Copper is obtained along with silver from veins at Great Bear Lake, in Bear Province, and much testing is being done on the copper occurrences in the basalts of the Coppermine River region.

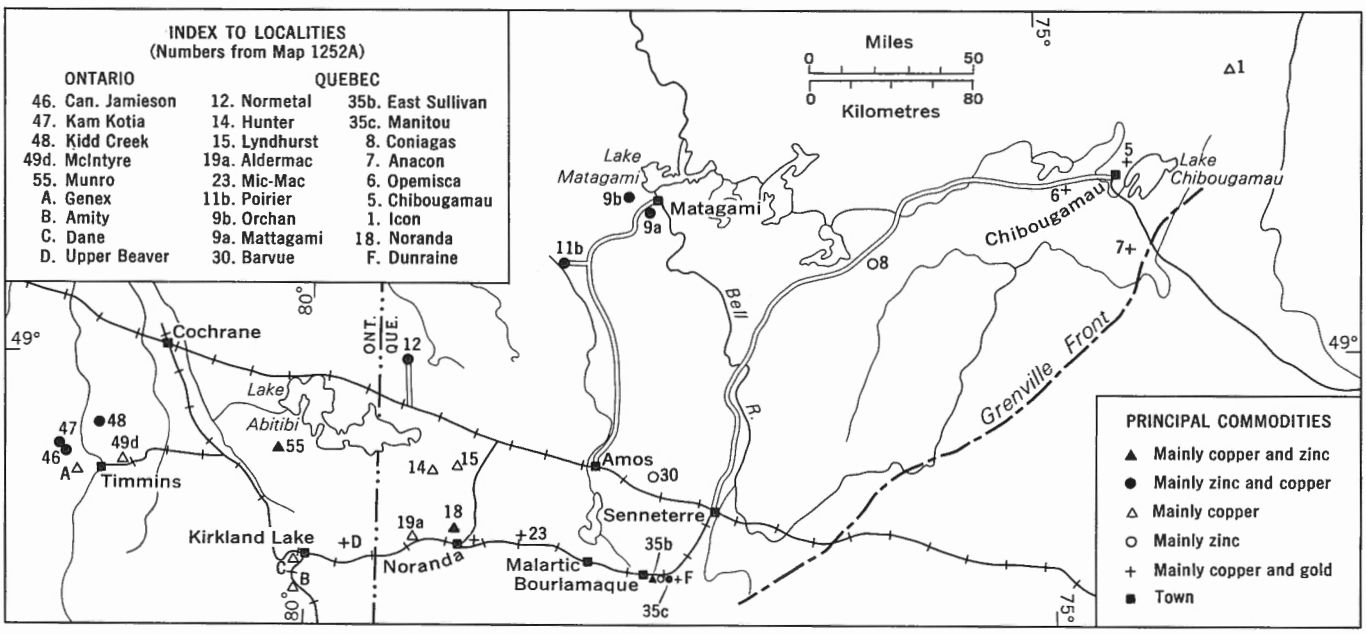


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FIGURE V-20. Copper and zinc mines of the Abitibi Belt, Ontario and Quebec (by A. H. Lang).

TABLE V-8 | *Copper and zinc mines in Superior Province*

Map No.[1]	Name	Production Began	Production Ended	Rate[2] (t.p.d.)	Main Tenor[3] % Cu	% Zn	Other	Type	Host Rock[4]	Remarks
ONTARIO										
	TIMMINS AREA									
48	Kidd Creek	1966		9,000	1.33	7.08	Ag	concordant massive	rhyolite	shipped to Kam-Kotia mill 1965-66
47	Kam-Kotia	1943, '61		1,700	1.73	2.94		shear zone	contact between basic and acid volcanics	
46	Can. Jamieson	1966		400	2.95	4.46			rhyolite	
49d	McIntyre (copper)	1963		1,800	0.88			porphyry copper	quartz–feldspar porphyry	contains molybdenite
	Genex[s]	1966	1967	200	1.95					
55	Munro	1967		700	2.09	1.69			greenstone near peridotite	
59b	Upper Beaver[s]	1965		150	1.19		Au	quartz veins	greenstone and syenite porphyry	shipped to Upper Canada mill
	Dane[s]	1911	1914	small				quartz vein	greenstone	Kirkland Lake
	Amity[s]	1928	1932	small					tuff, iron-formation	south of Kirkland Lake
QUEBEC										
	NORANDA AREA–MAIN									
18e	Horne	1927		3,500	2.2		Au	ellipsoidal bodies	rhyolite, etc.	
18f	Joliet				0.95		Au			operated by Noranda
19a	Aldermac	1932	1943	1,000	1.65			dissemi-nated masses	acid tuff and agg.	
18a	Amulet	1930	1962	1,000				domical masses	rhyolite	
	'A'				5.12	5.47		domical masses	rhyolite	
	'C'				2.2	8.5		domical masses	rhyolite	
	'F'				3.4	8.6		domical masses	rhyolite	
	'NW'				1.5			domical masses	rhyolite	
18c	D'Eldona		1952			7.7	Au	pyritic zone	volcanic	
18j	Quemont	1949			1.5		Au			
18k	West Macdonald[s]	1955	1959			3				zinc–pyrite shipper
17f	Waite, 'Old'	1928, '37	1961	700	4.7	2.98		lenses	rhyolite	
	Waite, 'East'	1952	1961		4.13	3.26				
17e	Vauze	1961	1965		5.2	3.7		tabular body in pipe	rhyolite	beneath diorite
18g	Lake Dufault	1964			3.3	6.6	Au			
	NORANDA AREA–OUTLYING									
40	Lorraine	1965		400	Cu		Ni			Belleterre area
23	Mic-Mac	1942	1944	600	Cu		Au	sheeted quartz stringers		Bonsquet tp.
35b	East Sullivan	1949	1966	2,800	0.51	0.51	Au	massive and dissemi-nated zones	tuff	Bourlamaque tp.

TABLE V-8 | *Copper and zinc mines in Superior Province (cont.)*

Map No.[1]	Name	Production Began	Production Ended	Rate[2] (t.p.d.)	Main Tenor[3] % Cu	% Zn	Other	Type	Host Rock[4]	Remarks
35c	Manitou	1942	1962	1,000	0.82	2.93		masses and disseminations	tuff	two distinct Cu and Zn zones
	Dunraine (Rainville)	1956	1958	500	1.22			at contact of diabase sill	tuff	two distinct Cu and Zn zones
14	Hunter[s]	1955	1957		1			disseminated lenses	rhyolite	Duparquet tp.
15	Lyndhurst[s]	1956	1957		1.93					Destor tp.
30	Barvue	1952	1957	5,300	0.82	2.93			tuff and agglomerate	Barraute area
12	Normetal	1937		900	1.07	7.1			fragmental volcanics	La Sarre area
MATAGAMI AREA										
9a	Mattagami Lake	1963		3,800	0.69	10.4		concordant lenses	rhyolitic	
9b	Orchan	1963		1,900	1.23	11.1		concordant lenses	rhyolitic	
10	New Hosco	1963		900	1.1	3.5		concordant lenses	rhyolitic	shipper to Orchan
11b	Poirier	1966		2,500	1.74	3.63		concordant lenses	rhyolitic	
11a	Joutel[s]	1967		2,358	2.35	9.5				shipper to Poirier; Zn ore has 0.25% Cu
8	Coniagas	1961	1967	350		4–7	Pb, Ag	lenses	agglomerate and tuff	at Bachelor Lake; contained considerable Pb
CHIBOUGAMAU REGION										
	Campbell-Chibougamau									
5b	Main Mine	1955		35% of total	2.06	0.03		shear zone	meta-anorthosite	3,000 t.p.d. concentrator includes Merrill Id. claims
5e	Kokko Creek[s]	1959	1966		2.24	0.01		shear zone	meta-anorthosite	
5c	Cedar Bay[s]	1958		24% of total	2.50	0.07		shear zone	meta-anorthosite	
2a	Henderson[s]	1960		39% of total	2.45	0.06		shear zone	meta-anorthosite	
	Grandroy[s]	1967		500 t.p.d.	1.2	0.03		stockwork	granite	15 miles to northwest
	Copper Rand									
3b	Main Mine	1960		46% of total	2.34	0.025		stockwork	meta-anorthosite	1,800 t.p.d. concentrator
3c	Jaculet[s]	1960		24% of total	2.05	0.025		stockwork	meta-anorthosite	
2b	Portage[s]	1960		21% of total	2.20	0.16		stockwork	meta-anorthosite	
5h	Que. Chib.[s]	1963	1967	3% of total	2.01	0.08		stockwork	meta-anorthosite	5 miles west of Main mine
3a	Bouzan[s]	1962		5% of total	3.22	0.25		stockwork	meta-anorthosite	eastern extension of main deposit
5a	Bruneau[s]	1966		135	1.60	0.02		stockwork	gabbro and pyroclastic volcanics	6 miles NW at Chibougamau
5g	Un. Obalski[s]	1964	1966	200				veins	granophyre	
6	Opemiska	1953		2,099	3.00	0.02		veins	gabbro	2,000 t.p.d. concentrator 20 miles SW of Chibougamau

TABLE V-8 | *Copper and zinc mines in Superior Province (conc.)*

Map No.[1]	Name	Production Began	Production Ended	Rate[2] (t.p.d.)	Main Tenor[3] % Cu	Main Tenor[3] % Zn	Main Tenor[3] Other	Type	Host Rock[4]	Remarks
1	Icon	1967		600	5.94 82.4				dolomitic	38 miles NE of Chibougamau; grades are for open-pit and underground ore
7	Anacon	1956	1960	500	0.39	0.21		shear zone and quartz veins	meta-volcanics	30 miles south of Chibougamau 500–750 t.p.d. concentrator
ONTARIO										
MANITOUWADGE AREA										
36	Geco	1957		3,300	1.95	4.15		concordant	meta-sediments	fairly high grade metamorphism
35	Willroy	1957		1,700	1.49	2.38		concordant	meta-sediments	fairly high grade metamorphism
34	Willecho[s]	1965		1,000	0.48	4.29		concordant	meta-sediments	fairly high grade metamorphism
OTHER MINES										
	Wendigo	1936	1943					quartz vein in shear zone	basalt	Lake of Woods
	Andowan	1903	1903					breccia zone	granite–pegmatite contact	Port Arthur area (Pb, Zn)
25	North Coldstream	1957	1967	1,000	1.95		Au, Ag	lenses	basic volcanics	Port Arthur area intermittent production 1903-17
	Zenmac	1966		100		23.3		lenses	meta-gabbro	Schreiber area
71	Bruce	1846	1876					quartz veins	diabase sills	Sault Ste. Marie area int. prod. 1908–1921
69	Jardun	1954	1957	300		3.52	Pb, Ag	veins	schist and sheared basic dykes	Sault Ste. Marie area int. prod. 1878–1904
70	Rock Lake	1899	1903		Cu			calcite quartz	sediments	Sault Ste. Marie area
68	Coppercorp	1965		500	2.1			veins	basalt and conglomerate	Sault Ste. Marie area[5]
67	Tribag	1967		400	2.2			stockwork		
	Cheney	1916	1916		4.65			vein		
76	Pater	1961		700	1.97			sulphide zone	metasediments and volcanics	Blind River area
	Parkway	1906	1906		Cu			veins	metasediments	Blind River area
81	Lake Geneva	1941	1944			9 to 13	Pb	concordant	metasediments and breccia	north of Sudbury
	Pax Intern'l	1950	1965	150	1.25		Mo	shear zones	peridotite	Matachewan area
63b	New Ryan	1948	1953		Cu		Au, Ag			Matachewan area
	Matarrow	1952	1953			1.59	Pb, Ag	lenses	iron-formation	Matachewan area
94	Temagami	1959		150	6.17		Au, Ag	zones near contact	rhyolite and diorite	Timagami area
	Trebor	1936	1936		0.84		Ni			Timagami area

[1] Number is that assigned to the deposit, according to Province, on Mineral Deposits of Canada, Map 1253A (in folio). Mines without a number are not shown on the map.

[2] Production rates per day, generally for latest producing year.

[3] Grades are derived from 1966 reserves or the latest producing year, and are intended to be a general indication only.

[4] Host rocks are the principal rock types immediately enclosing the main deposits.

[5] In Southern Province.

[s] Shipping mine.

Exclusive of nickel operations, 45 mines produced copper or zinc or both as main products in 1966. During that year there were 41 other mines classified as former producers. A few mines began to produce in 1967, but about the same number were closed. The Horne mine has produced for 40 years, the Flin Flon for 37, and the Normetal for 30. At the other extreme a few small shipping mines were operated for a year or less. Most periods of operation are about 10 to 20 years. The deposits, many of which comprise several orebodies within the compass of a single mine, range downwards from the huge Kidd Creek deposit being mined at a daily rate of 9,000 tons; most are in the range of 500 to 3,000 tons per day. Depths vary from 1,000 to 4,000 feet below surface; one mine, the Normetal, extends to 8,000 feet. The long axes of many deposits or clusters of deposits pitch steeply; thus the depths of some mines are two or three times the longest lateral dimensions of the deposit. The tenors of the strata-bound ores are about 0.5 to 5 per cent copper and 0.5 to 11 per cent zinc, but many mines have sections or separate bodies having different tenors.

Many of the mines are centred around the large concentrators and smelters at Noranda, Quebec, Flin Flon, Manitoba, and Sudbury, Ontario. In addition to treating ore from the company's mines, these plants also treat ore

TABLE V-9 *Copper and zinc mines in Churchill Province*

Map No.[1]	Name	Production Began	Production Ended	Rate[2] (t.p.d.)	Approx. Main Tenors[3] % Cu	% Zn	Other	Type	Host Rock[4]	Remarks
MANITOBA										
2	Sherridon	1931, 1937	1951	3,000	2.2	0.88		massive and disseminated bodies	quartzite–gneiss contact	
4b	Schist Lakes[s]	1954		300	4.76	9			andesite	
4a	Mandy[s]	1917	1920		7.3	12.9			schist and breccia	
5	Cuprus	1948	1954	300	3.25	6.4			andesite and tuff	
6b	North Star[s]	1955	1958						andesite and rhyolite	
6a	Don Jon[s]	1955	1958						andesite and rhyolite	
8	Chisel Lake[s]	1960		1,000	0.59	10.3	Pb		schist and gneiss	
10	Stall Lake[s]	1964			4.5	1.6				
	Waden Bay	1966		900	2.4			lenses	arkose–greywacke contact	Anglo-Rouyn Mines Ltd.
	Rottenstone Lake	1965		150	2	2			ultrabasic rock	
	Hanson Lake	1967		350	0.4	7.35	Pb	lenses	tuff	
SASKATCHEWAN										
8	Flin Flon	1930		6,300 (capacity)	2.23	4.57		massive and disseminated bodies	andesite and tuff	operated at 4,630 t.p.d. in 1966
9	Birch Lake[s]	1957	1960		6.2				intermediate lava flows	
10	Coronation[s]	1960	1965		4.25	0.2			basic flows and tuff	

[1] Number is that assigned to the deposit, according to Province, on Mineral Deposits of Canada, Map 1252A (in folio). Mines without a number are not shown on the map.
[2] Production rates per day, generally for last producing year.
[3] Grades are derived from 1966 reserves or the latest producing year, and are intended to be a general indication only.
[4] Host rocks are the principal rock types immediately enclosing the main deposits.
[s] Shipping mine.

or concentrates from smaller mines in their respective regions. Some concentrators also treat ore shipped from smaller nearby mines that are similar mineralogically. For example, three concentrators at Chibougamau are treating or have treated ore from 14 mines, and the concentrators at Matagami and Poirier treat ore from neighbouring mines. This system permits the mining of some deposits that otherwise could not have been economically worked, or could only have been worked at greater capital costs. The principal deposits are described briefly in the following sections. They are grouped together into regions of similar geology and according to the geological provinces of the Shield.

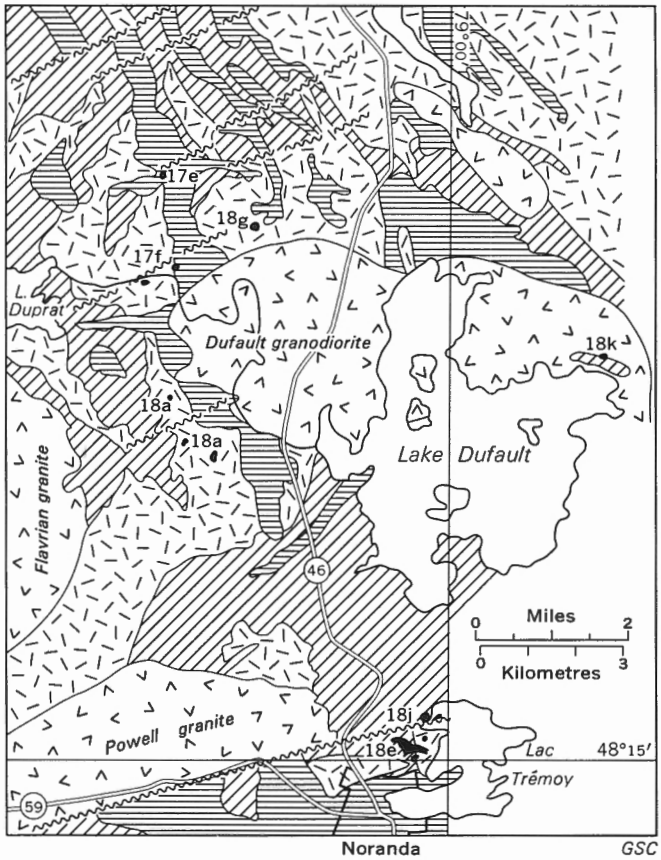
Copper–Zinc Deposits of Superior Province

Abitibi Belt, Ontario and Quebec

This region contains thick eugeosynclinal successions of metavolcanic and metasedimentary strata intruded by large and small masses of acidic, intermediate, and basic plutonic rocks. The belts of stratified rocks generally strike easterly but are internally complicated by local folds. The southern part of the belt extends eastward from Timmins, Ontario, through Noranda and Val d'Or to Bourlamaque, Quebec. This part contains the bulk of the mines. One hundred miles north the recently developed mines of the Poirier and Matagami areas are aligned with the Munro and Normetal mines.

Timmins, Ontario. In the Timmins area, but about 15 miles northwest of the main gold mines, are the *Kam-Kotia* (47), *Jamieson* (46), and *Genex* deposits of copper with some zinc and gold. They occur in Archean volcanic rocks on the flank of a large body of diorite and gabbro. Twelve miles northeast is the *Kidd Creek* (48) deposit, found in 1964 by geophysical prospecting in a terrain mainly covered by overburden. The deposit consists of massive sulphides concordant with enclosing steeply dipping rhyolite breccia that is overlain by andesite. An early estimate of open-pit ore indicated an orebody at least 400 feet wide and 2,000 feet long, with an exceptionally high content of zinc and silver and 1.33 per cent copper. At the *McIntyre* (49d) gold mine an important copper deposit was also found. It is of a type unusual in the Canadian Shield, being a dissemination of chalcopyrite with some tennantite and molybdenite in the Pearl Lake quartz–feldspar porphyry.

Noranda, Quebec. The sulphide deposits in the main part of Noranda area are in a thick pile of volcanic rocks containing much rhyolite, dacite, and related tuffs and breccias as well as andesite. Interbedded sediments are rare in comparison with the amount that occurs to the east and west. Most deposits are in acidic volcanics overlying a more widespread basic volcanic sequence; this leads to the speculation that the area had been one of abundant volcanism in an Archean eugeosynclinal belt,

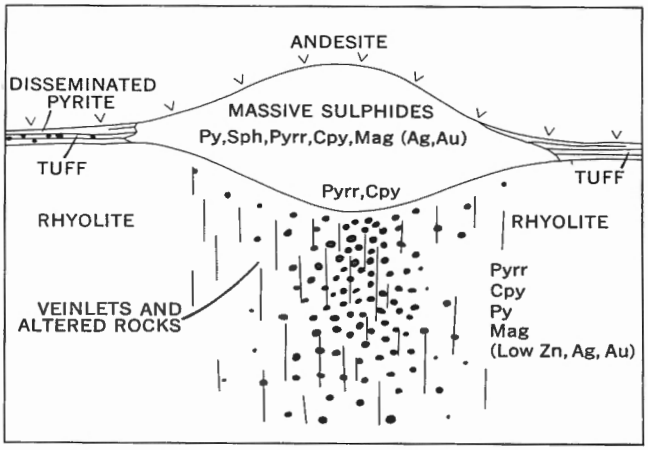


INDEX TO DEPOSITS
(Numbers from Map 1252A)

17e. Vauze	18g. Lake Dufault
17f. Waite	18j. Quemont
18a. Amulet	18k. West Macdonald
18e. Horne	

FIGURE V-21. Main copper and zinc producing part of Noranda area, Quebec. Some deep-seated deposits are in rocks other than shown on this surface map.

and that the volcanic host rocks of the deposits were laid down at the close of the main sequence of geosynclinal strata, or a little later. Many deposits occur peripheral to the Dufault Lake granodiorite body, those on the west side being in a remnant of volcanics bounded on the west by the Flavrian Lake granite (Fig. V-21). Most are lenticular, dome shaped, or ellipsoidal (Fig. V-22). All occur in acidic volcanics and most are at contacts of rhyolite with overlying andesite. Several are in rhyolite breccia beneath massive rhyolite. The *Waite* (17f), *East Waite*, and *Lake Dufault* (18g) orebodies are aligned and together



Pyrite	Py	Magnetite	Mag
Pyrrhotite	Pyrr	Zinc	Zn
Chalcopyrite	Cpy	Silver	Ag
Sphalerite	Sph	Gold	Au

GSC

FIGURE V-22. Features common to many sulphide deposits in Noranda area, Quebec (by S. M. Roscoe).

with that at *Amulet* (18a) appear to lie along northwesterly and northeasterly trending lineaments that form a rectilinear set. Several deposits north of the Horne mine are at the top of pipe-like zones of alteration to chlorite, sericite, and silica. Age determinations, geological settings, the fact that host rocks near deposits are deficient in the metals contained in the deposits, and other evidence cause several of the most recent students of the origin of the deposits (Gilmour, 1965; Roscoe, 1965) to conclude that the deposits are related genetically to the volcanism, but that they and their enclosing rocks are only slightly younger, and that subsequently in the Kenoran Orogeny the deposits and their host rocks were folded and tilted to their present attitudes.

The deposits are of two main mineralogical types: pyrrhotite–chalcopyrite–magnetite and pyrite–sphalerite. Both types carry minor amounts of gold and other metals. They commonly occur together as parts of a single orebody or as separate bodies at a single mine. In the latter case the pyrite–sphalerite bodies lie above the pyrrhotite–chalcopyrite bodies. This suggests that copper deposits might be encountered below known zinc orebodies in other areas.

The *Horne* (18e), the oldest and largest in the Noranda area, has yielded more than 50 million tons of ore containing more than one million tons of copper, 8 million ounces of gold, and large amounts of silver and pyrite. Reserves of the main orebodies at the end of 1966 were 3,831,600 tons carrying 2.47 per cent copper and 0.19 ounce of gold per ton. The orebodies are of irregular shapes, with long axes vertical or dipping steeply (Fig. V-23). They are in a fault block containing rhyolite and related tuff and agglomerate, and bodies of 'metadiabase' or 'older gabbro' that also may be related to the volcanism. Most ore is in the 'H' and Lower 'H' bodies. Some others contain siliceous flux ore. An alteration pipe has not been found beneath the Horne cluster of deposits. Some geologists suggest that the alteration found in places at the edges of deposits may be remnants of a faulted pipe. The largest body, the 'No. 5', contains pyrite, sphalerite, chalcopyrite, and gold and silver, but is low grade. It lies to the north, beyond the group shown on Figure V-23, and has been explored from surface to a depth of 8,000 feet. The feasibility of mining sections of it is being investigated. Gilmour (1965) suggests that it is concordant with the strata and stratigraphically above the copper-bearing bodies.

The *Poirier* (11b) mine, 60 miles north of Amos and 35 southwest of Matagami, is in a belt of volcanics

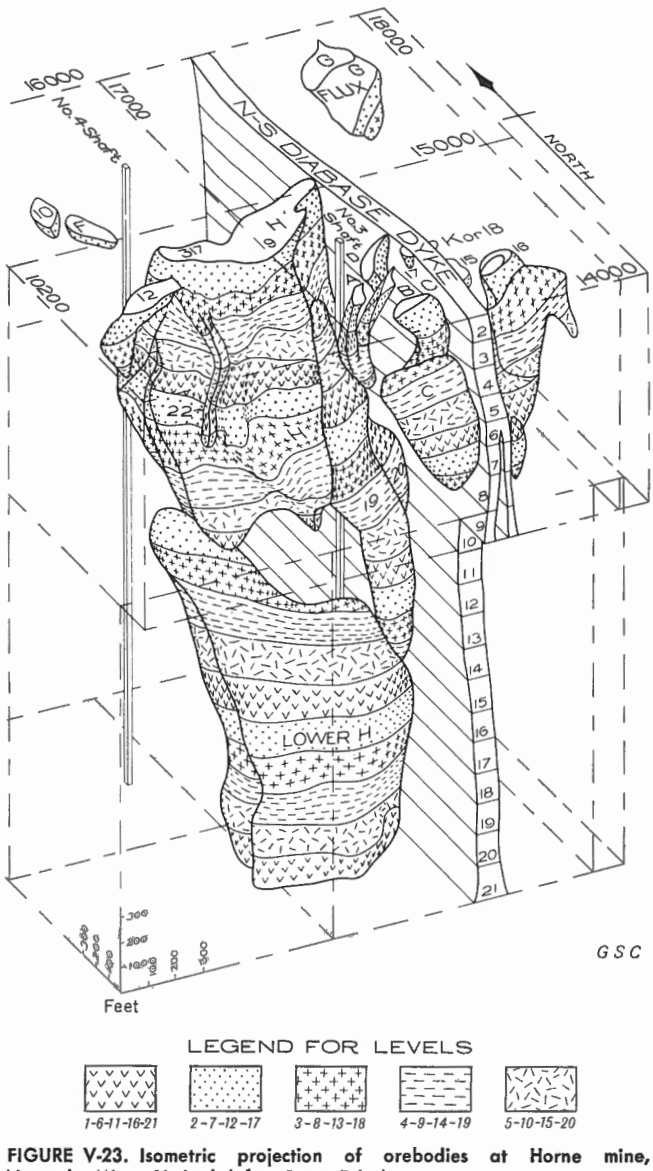


FIGURE V-23. Isometric projection of orebodies at Horne mine, Noranda Mines Limited (after Peter Price).

a mile from a granitic body. A zone 300 feet wide and 1,500 feet long lies in the steeply dipping contact between rhyolite and rhyolite tuff. It contains lenticular bodies in rock much altered to chlorite and talc. The lenses are of three kinds: predominantly pyrrhotite with sphalerite; chalcopyrite with some pyrrhotite; and chalcopyrite–sphalerite. The nearby *Joutel* (11a) deposit has a copper and zinc–copper section.

Matagami area, Quebec. The Matagami area is underlain by mafic and rhyolitic volcanics and the northwestern part of the Bell River complex of anorthosite, gabbro, and other rocks (Fig. V-24). Around the nose of the main part of the complex are sill-like bodies so intimately associated with the volcanics that a common origin has been suggested. The area contains three producing zinc–copper mines, the *Mattagami Lake* (9a), *Orchan* (9b), and *New Hosco* (10). These and several prospects are in a sequence of mainly rhyolitic rocks near some of the smaller bodies of the Bell River complex. The three mines and nine prospects have most of the following characteristics in common (Latulippe, 1966): they are in the contact zone of the Watson Lake group of rhyolitic rocks with the Wabassee group of dacite, andesite, and basalt; most host rocks are little metamorphosed; the sulphide masses are lensoid, some being deformed; the tops of the masses abut sharply against overlying strata but their basal parts grade into the host rocks; the deposits contain both massive and disseminated pyrite, pyrrhotite, sphalerite, chalcopyrite, and magnetite. The largest mine, the Mattagami, is at the nose of a minor drag-fold along the flank of the main anticline of the area. This property and the adjoining Orchan contain five orebodies as thick as 400 feet, over a total length of 6,700 feet. One of the Orchan bodies extends to a depth of 1,200 feet.

Manitouwadge area, Ontario. The *Geco* (36) discovery in Manitouwadge area, north of Lake Superior, was confirmed in 1953 by prospectors investigating an occurrence of sulphides noted on a map issued by the Ontario Department of Mines in 1932. Production was begun in 1957. Further prospecting and testing led to the discovery of the adjoining *Willroy* (35) and *Willecho* (34) orebodies farther northwest. In 1966 the Geco yielded 46,120 tons of zinc, 26,770 tons of copper, 840 tons of lead, 2,203,500 ounces of silver, and some gold. The combined Willroy and Willecho production was 10,696,766 pounds of zinc, 2,250,136 pounds of copper, 852,574 pounds of lead, 311,003 ounces of silver, and 429 ounces of gold.

The area is underlain by a synclinal roof-pendant of metavolcanic and metasedimentary strata of Keewatin type, surrounded by biotite–granodiorite gneiss and migmatite prevalent in the district. The development of the area is noteworthy because such small, strongly metamorphosed 'greenstone' belts were generally considered unfavourable for base metals. The syncline plunges northeast causing the strata to nose into the granitic rocks to the southwest and to assume a crescentic shape to the northeast, the two extremities being 6 miles apart where they taper into the granitic rocks. The older rocks, exposed in the western and southern parts of the area, are amphibolite and massive and laminated hornblende

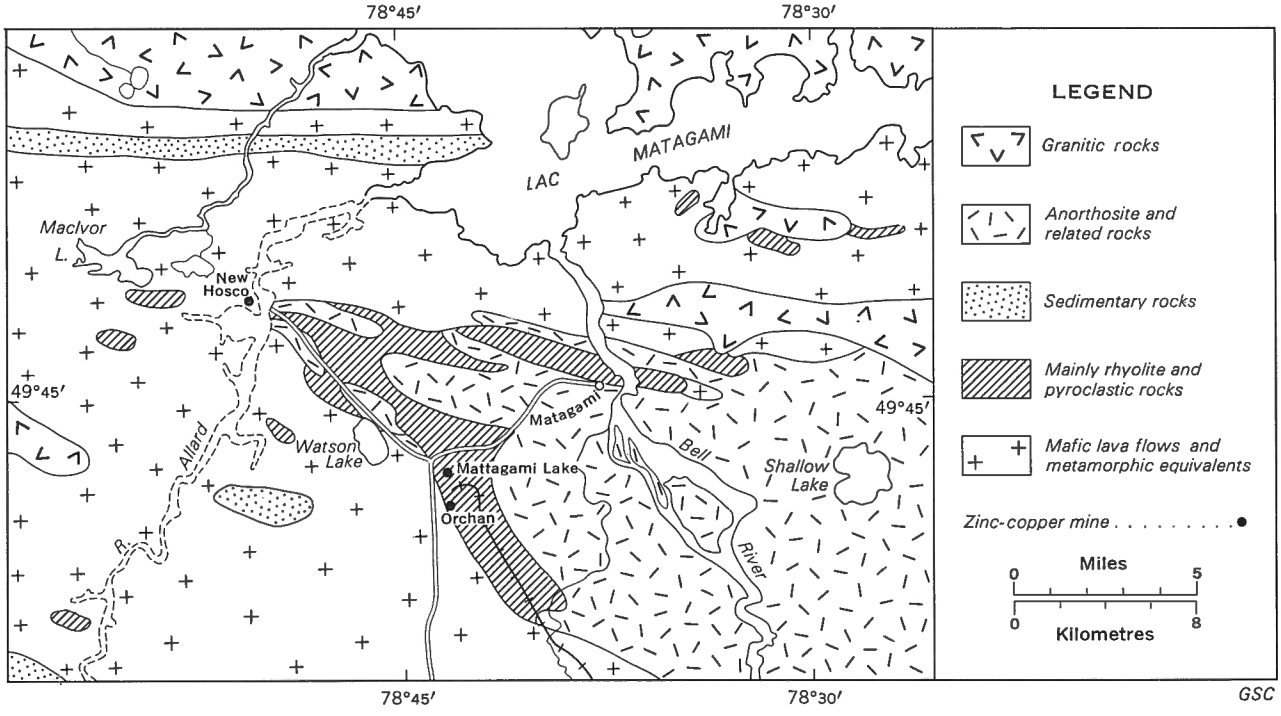


FIGURE V-24. Mines and prospects of Matagami area, Quebec (Latulippe, 1966).

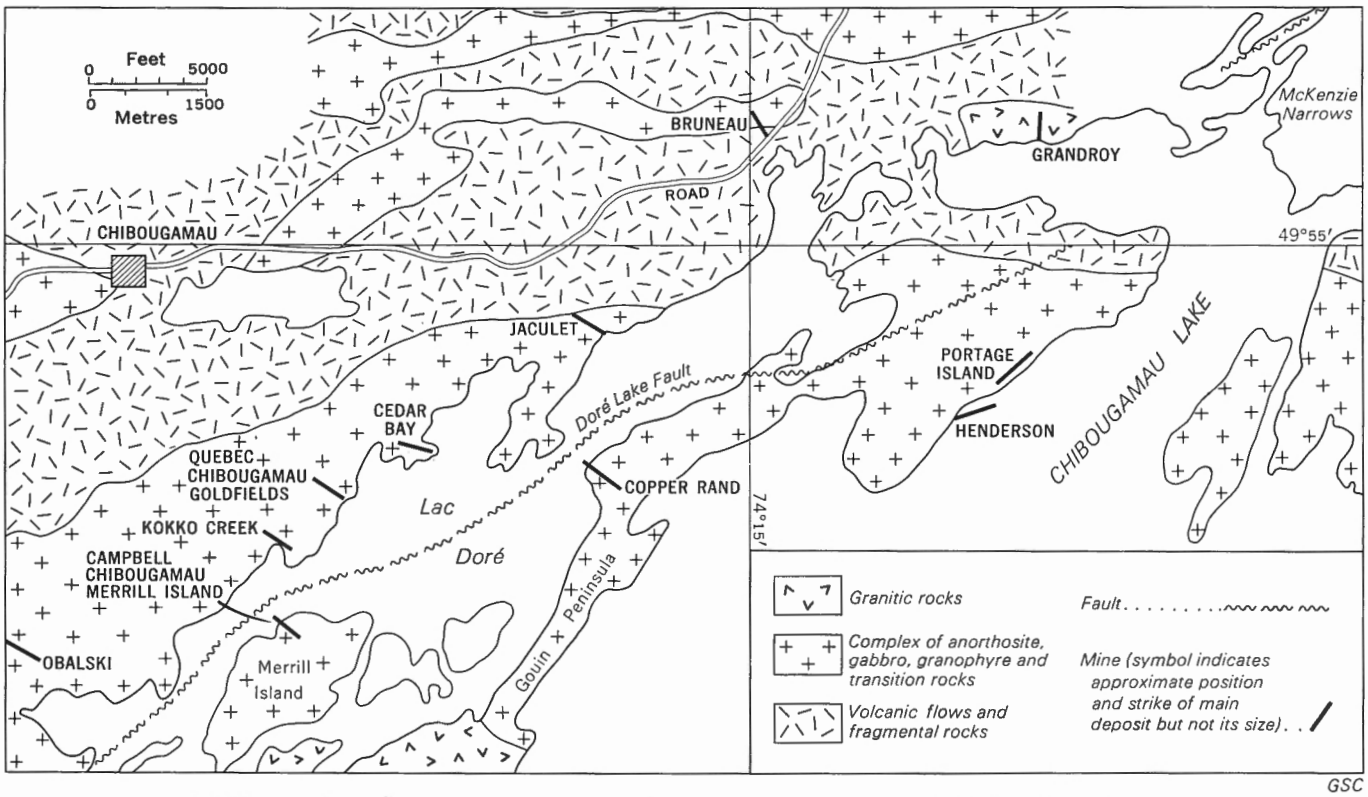


FIGURE V-25. Generalized geology and copper–gold mines of Chibougamau region, Quebec (by A. H. Lang).

schists. These are overlain by gneisses, schist, and iron-formation. A narrow belt of metavolcanics overlies the sedimentary rocks. Bodies of meta-gabbro intrude the volcanics but are not known to cut the sediments. Dykes and sills of microgranodiorite and dykes of pegmatite intrude the strata, and the youngest rocks of the area are diabase dykes. The area is crossed in various directions by faults; three northerly trending faults are near the orebodies but neither the emplacement nor the grade of ore seems related to them.

The orebodies are in foliated rocks, and conformable with them; all but one are in metasediments that are mainly iron-formation and associated garnet–amphibole schist. The orebodies are high-temperature types and of three kinds: disseminations, lode fissures containing cores of massive sulphides surrounded by disseminations, and a massive deposit. The most common minerals are pyrite, pyrrhotite, quartz, chalcopyrite, sphalerite, and galena. The relative amounts of sulphides and precious metals vary. Thus, some bodies are classed as copper–zinc–silver, some as copper–zinc–silver–lead, some as zinc–silver–lead, and one as copper–silver. The mineralization was later than the pegmatites and basic intrusive rocks and is probably older than the diabase dykes, and appears to have taken place during the later part of the period of granitic activity. Recent work has led to an explanation of the orebodies as being composed of series of generally coalescing lenses controlled by drag-folds later than the regional folding. Some lead isotope ratios have been interpreted as indicating that the age of the deposits is about 2,600 m.y. Others have suggested that early lead, about 3,250 m.y. old, was mixed with lead of about 2,250 m.y. to produce the intermediate age.

The main Geco orebody is divided by a north-striking fault and a diabase dyke 850 feet to the east. The west part is as much as 970 feet long and 50 feet wide, and extends to a depth of at least 1,000 feet. The central part is offset 200 feet to the north, and the east part extends an unknown distance beyond the dyke. The entire body rakes 35 degrees east. The west sector therefore represents the lowest part, and the east sector the highest. It has massive and disseminated phases. Massive ore is generally high in zinc, but copper and silver, as well as zinc, are common in the lower parts. Disseminated ore is mainly of value for copper and silver. One of the Willroy orebodies is in biotite–sillimanite–muscovite–quartz schist and the other three as well as the three Willecho bodies are in iron-formation.

Chibougamau Region, Quebec

Deposits of copper and gold were found and tested in Chibougamau region long ago because it was on a canoe route between Lake St. John and Hudson Bay; other deposits were found later, some by geophysical prospecting beneath lakes frozen in winter. After a road and later a railway were built production was begun at the Opemiska mine in 1953. The properties in the Chibougamau region (Fig. V-25) are now mostly operated by the Campbell

Chibougamau and the Copper Rand companies. Each has a concentrator at its main mine, to which it ships, or did ship, ore from outlying mines. Also, the Merrill Island company built a concentrator to treat ore mined from a part of its property not leased to Campbell Chibougamau; this plant has recently been used to treat custom ore from the Bruneau and Icon mines. Although much of the region is underlain by volcanic and granitic rocks the main feature is a differentiated sill complex of diorite, gabbro, and other rocks. The complex includes a large unit of altered anorthosite in which the principal deposits of the main Chibougamau camp are present. This unit is crossed by the Doré Lake fault, a northeasterly striking shear zone containing chlorite, sericite, quartz, and siderite. Duquette (1966) states that nine mines are within 4,000 feet of this fault and are mainly in shear zones striking west-northwest. They contain chlorite, sericite, carbonate, silica, sulphides, and gold. Siderite is common. The main economic mineral is chalcopyrite. The Portage and Henderson orebodies are in shear zones striking parallel with the apparent continuation of the Doré Lake fault.

The main *Campbell Chibougamau* (5b) property now contains a shear zone on the former Merrill Island claims and the underwater continuation to the Doré Lake fault (Fig. V-26). The zone contains porphyry dykes parallel with the shearing. The entire zone is 800 feet wide and 4,000 feet long. In it faulted sulphide bodies average 40 feet wide and are about 2,500 feet long. The main zone decreases below the 1,450-foot level and ends at the 2,000-foot level. Another body was found below the 2,700-foot level. The bodies contain about 33 per cent pyrrhotite, 10 per cent chalcopyrite, 5 per cent pyrite, 2 per cent sphalerite, 2 pounds of cobalt per ton, and free-milling gold. The lowest mine level is 3,750 feet below surface.

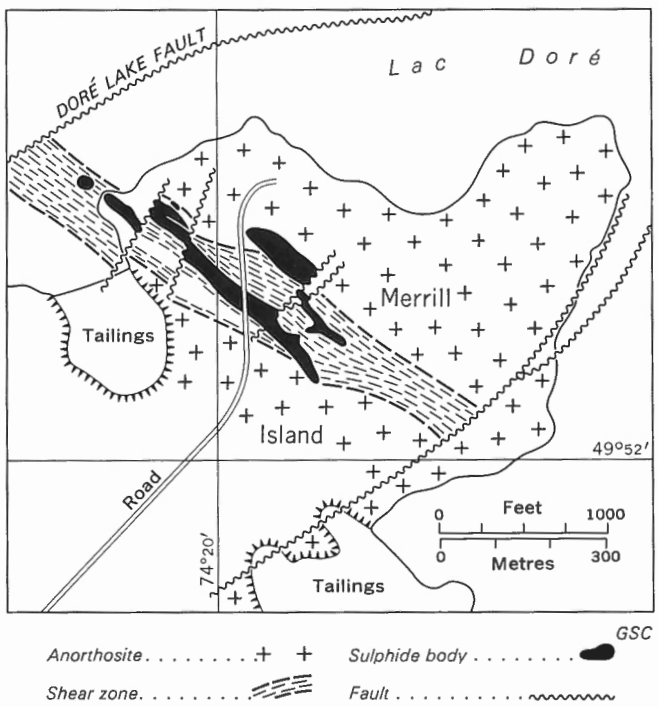


FIGURE V-26. Main Campbell Chibougamau–Merrill Island copper–gold deposit, Quebec (R. B. Graham).

The main *Copper Rand* (3b) mine lies in a northwesterly striking shear zone crossing Gouin Peninsula. The surface discovery at Machin Point is 1,000 feet from the probable position of the Doré Lake fault. The zone continues eastward into the *Bouzan* (3a) property. The zone of shearing and alteration is as wide as 2,000 feet and contains a zone of diorite and quartz porphyry dykes 400 feet in total width, parallel with the shearing. In the centre of the zone the orebodies are lenticular and occur at the sides of dykes. The lenses are 3 to 60 feet wide and have lengths of generally less than 500 feet. Most are within a total length of 2,500 feet. The mine is opened to a depth of 2,130 feet.

The *Icon* (1) mine, at the south end of Lake Mistassini, is a recent discovery in dolomitic strata of the Aphebian Mistassini Group. Diamond drilling showed three flat orebodies, all in the same stratigraphic horizon. The bodies are in a quartz–carbonate zone that may have been introduced along a fault parallel with the bedding. They are covered by 40 to 60 feet of overburden or a little rock. The bodies, 13 to 22 feet thick, contain disseminated and massive chalcopyrite, some recoverable silver, a little pyrite, and quartz and carbonate. They were stated in 1967 to contain 252,030 tons of open-pit ore averaging 5.94 per cent copper, and 504,370 tons averaging 2.4 per cent copper to be mined from an adit. In 1967, ore was trucked to the Merrill Island concentrator at a daily rate of 600 tons.

The *Opemiska* (6) deposits differ from typical Chibougamau ones in that they are mainly veins, occur in a gabbro sill rather than in anorthosite, and have quartz–calcite–chlorite gangue without reported siderite or sericite. The main veins are in shear zones in faults crossing the nose of an anticline in the sill. Parts of some deposits appear to be stockworks, and some wall-rock contains disseminations. The main orebodies are chalcopyrite, pyrite, and magnetite, with minor pyrrhotite, galena, sphalerite, molybdenite, and scheelite. Ore carries 5 per cent copper with about 0.05 ounce of gold and 0.5 ounce of silver per ton. The most important vein is a few inches to 25 feet wide. It was mined for a length of 2,500 feet, and to a depth of 2,400 feet; it is being deepened an additional 1,000 feet.

Other deposits in Superior Province. The *Zenmac (Zenith)* (33) mine, 10 miles northwest of Schreiber, Ontario, was worked in a small way from 1895 to 1901. Production with a 100-ton-per-day mill was begun in 1966. The main deposit is a fracture zone in gabbro that has irregular lenses 2 to 10 feet wide containing massive sphalerite with minor chalcopyrite and pyrrhotite. The zone has been traced 600 feet. In 1966, more than 13

million pounds of zinc, 126,000 pounds of copper, and 38,000 pounds of cadmium were produced. The nature of the deposit was reported to be such that an accurate estimate of ore reserves was not possible.

The *Temagami* (94) mine is on an island in Timagami Lake underlain by Archean basic to acidic volcanic rocks of Keewatin type intruded by a meta-diorite sill. Occurrences of sphalerite, chalcopyrite, and pyrite on the island had been known for some time but after diamond drilling and a geophysical survey two high-grade massive sulphide bodies were indicated. Open-pit mining was begun in 1955, the ore being shipped by barge to the railroad at Timagami. A shaft was sunk in 1956 and a 150-ton mill was built in 1958. The mine contains three types of ore: low-grade pyritic ore carrying copper, nickel, and cobalt; 'open-pit' high-grade massive chalcopyrite ore; and vein and replacement chalcopyrite ore. The orebodies lie in meta-diorite and rhyolite breccia and tuff, and are distributed near the south contact of the sill. They are surrounded by zones of chloritization or rock containing sericite, carbonate, and tourmaline. In 1966, ore reserves were estimated to consist of 97,884 tons averaging 6.73 per cent copper and 16,600 tons averaging 3.74 per cent copper. Production in 1966 totalled 6.8 million pounds of copper, 1,007 ounces of gold, and 17,438 ounces of silver.

Although deposits at *Mamainse Point,* 60 miles north of Sault Ste. Marie, Ontario, are not ranked with the main copper mines of the Canadian Shield they are of unusual interest because they contain chalcocite and native copper, have some of the characteristics of the Keweenawan deposits of Michigan, and were the object of some of the earliest Canadian attempts at mining. Experimental mining of copper was tried in 1770 by missionaries and an explorer near Sault Ste. Marie, probably at these deposits or near them. The orebodies are calcite–quartz veins carrying chalcocite and minor bornite, chalcopyrite, and native copper that occur in fault zones in interbedded Keweenawan basalt and conglomerate. In 1936, a piece of almost pure native copper weighing 147 pounds was found, but the native copper content of the deposits is generally minor. At the *Coppercorp* (68) property four main copper-bearing zones were located in 1949–52 and another was found later. Production was begun in 1965 and the mine is developed to a depth of 500 feet. The zone now providing most ore is reported to average 8 feet wide and 5,400 feet long. In 1965, reserves were reportedly 1,540,000 tons averaging 2.1 per cent copper.

The *North Coldstream* (25) mine, 65 miles west of Port Arthur, Ontario, produced from 1903 to 1917 and in 1957 to 1967. Sulphide orebodies lie in a lenticular chert-like body in basic and acidic schists. Whether the chert represents an original sediment or silicified schists is uncertain. The brittle chert appears to have fractured favourably for the introduction of pyrite, chalcopyrite, and a little pyrrhotite. These minerals are present in fracture fillings, or as masses and disseminations that form lenticular orebodies carrying an average of 1.95 per cent copper.

Copper Deposits of Southern Province

The *Bruce* (71) mine on the north shore of Lake Huron was the first copper mine in the Canadian Shield and one of the earliest producers in North America. Production was intermittent from 1846 to 1921, the ore being shipped to smelters in Wales, the United States, and finally Sudbury. Quartz–chalcopyrite veins in diabase sills in Aphebian Huronian sediments varied from stringers to 50-foot veins. The main vein, traced for at least 2,000 feet, had an average width of 5 feet and was worked to a depth of 450 feet. Early records were lost but the maximum tenor appears to have been about 3.5 per cent copper.

Copper–Zinc Deposits of Churchill Province

The Flin Flon and Snow Lake regions of Manitoba and Saskatchewan contain many zinc–copper deposits and in some copper predominates. With regard to the production of copper and zinc in recent years, this part of Churchill Province is second only to the Timmins–Chibougamau belt in Superior Province. Mining was begun in Saskatchewan at the Mandy mine in 1917. Its ore was taken by sled 40 miles and then by barge 130 miles to the railroad at The Pas, Manitoba, for shipment to the smelter at Trail, British Columbia. A larger deposit was discovered in 1915, a few miles away, at Flin Flon. It has been in continuous production since a smelter was completed in 1930 and is still the largest mine in the district. Ore and concentrates from mines in the Snow Lake district some 80 miles east of Flin Flon are shipped to the Flin Flon smelter. Four of the mines have been exhausted, four are still producing, and four others are being prepared. The large Sherritt–Gordon copper–zinc deposit at Sherridon, Manitoba, 40 miles northeast of Flin Flon, was found in 1922. Concentrates were shipped to the Flin Flon smelter in 1931–32 and from 1937 to 1951, when the ore was exhausted. The company acquired the Lynn Lake nickel–copper deposit and moved much of the Sherridon operation to it, production beginning in 1953.

Flin Flon–Snow Lake Region, Manitoba and Saskatchewan

This region is underlain largely by the Archean Amisk Group of volcanic and sedimentary strata, the unconformably overlying Missi sedimentary group, and the Kisseynew granitic gneisses that are derived from the Amisk and Missi Groups and developed extensively north of the mining district. The strata are intruded by granitic rocks, mainly contemporaneous with the Hudsonian folding and metamorphism. Some geologists believe that the rocks of this southern part of Churchill Province were

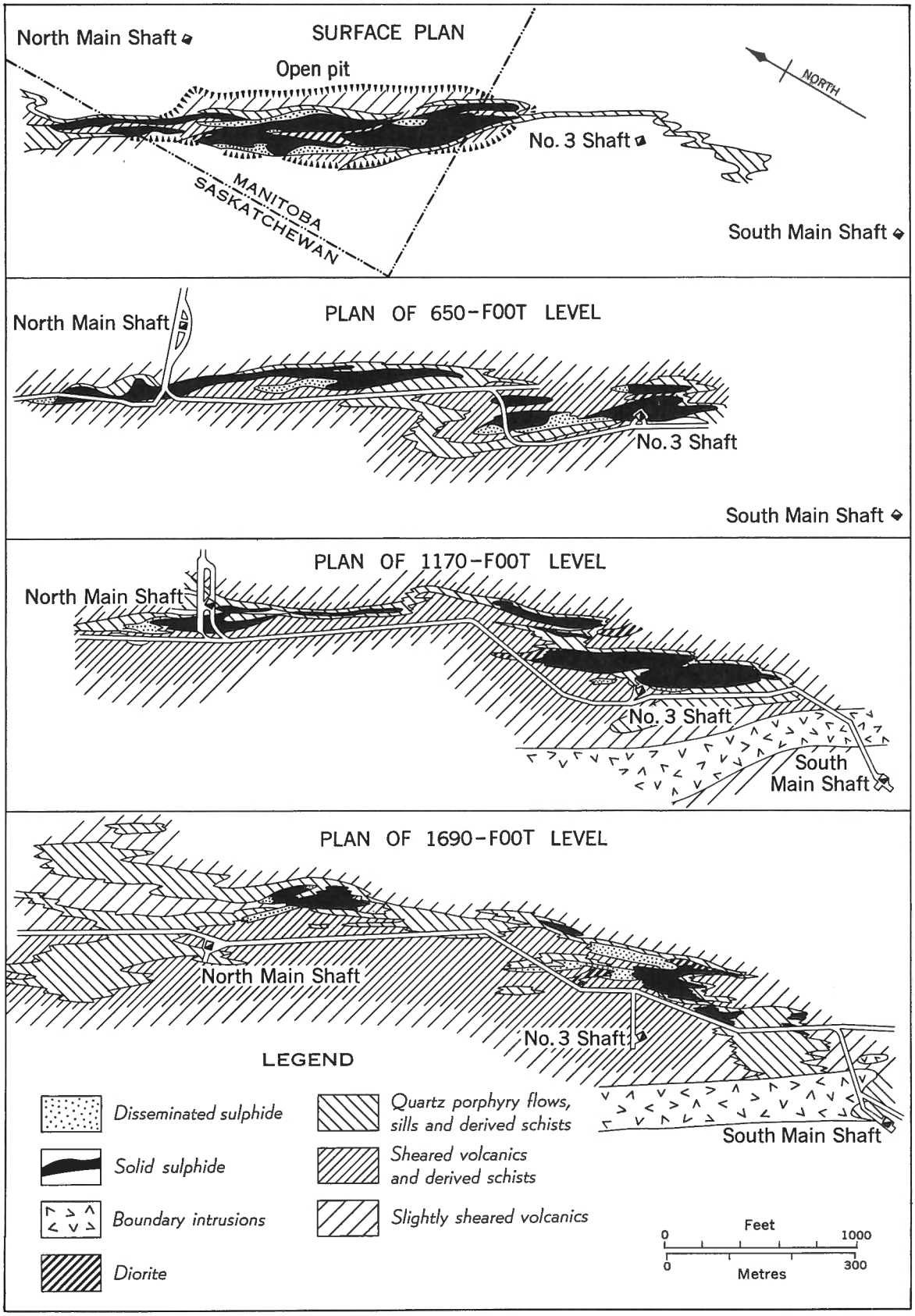


FIGURE V-27. Flin Flon mine, Manitoba (after the staff, Hudson Bay Mining and Smelting Company, Ltd.).

also subjected to the earlier Kenoran deformation related to the one in Superior Province. The orebodies are mainly in volcanic rocks of intermediate and basic composition or their metamorphosed equivalents. Most orebodies have both massive and disseminated zones, massive ore being the more abundant. Zoning is common in the massive deposits. Pyrite is the predominant sulphide mineral but sphalerite and chalcopyrite are the main ore minerals. Their ratios vary so that some deposits or parts of deposits are classed as zinc–copper ores and others as copper–zinc. Other sulphide minerals are present, and silver, gold, lead, cadmium, and selenium and tellurium are produced as byproducts. The orebodies are surrounded by narrow zones of alteration in which chloritization and carbonatization are characteristic of the more basic rocks whereas the more acidic ones are altered mainly by silicification, sericitization, and carbonatization. Ore deposits in the Flin Flon area have been found only in the Amisk rocks, which are cut by dykes and sills of feldspar porphyry that appear to be about the same age as the Boundary Intrusions. Lead isotopes indicate an age of 1,735 $\pm$ 160 m.y. for the galena in the orebodies.

The entire *Flin Flon* (3) deposit comprised six lenticular bodies in a shear zone having quartz porphyry as its hanging-wall (Fig. V-27). The porphyry lies along the contact between overlying competent flows and underlying less competent pyroclastic rocks and flow breccia. The general zone was 1,100 feet long by 200 feet wide, and is being worked to a depth of 3,700 feet. Complete production figures for this mine have not been published, but from 1930 to 1963 a total of about 53,200,000 tons of ore were mined. This averaged 4.57 per cent zinc, 2.23 per cent copper, and 1.30 ounces of silver and 0.074 ounce of gold per ton.

The nearby *Coronation* (10) deposit in Saskatchewan was mined from 1960 to 1965. During the life of the mine it yielded 1,412,861 tons of ore averaging 4.25 per cent copper, 0.20 per cent zinc, and 0.150 ounce of silver and 0.060 ounce of gold per ton. The deposit, 900 feet long and 120 feet wide, was mined to a depth of about 950 feet. It lay along a fault in basic lava flows and tuffs, recrystallized and altered to the staurolite–almandine grade of metamorphism. Although much smaller than the Flin Flon deposit it is of special interest and was selected for numerous research studies on ore genesis.

Other deposits in the Flin Flon region differ only in detail from the two described above. The *Birch Lake* (9), *Cuprus* (5), and *North Star* (6b) deposits have been mined out. The *Schist Lake* (4b) zinc–copper deposit is in production and the *Flexspar* copper deposit is being developed. In the Snow Lake area the *Chisel Lake* (8) zinc–lead–copper deposit and the *Stall Lake* (10) copper–zinc deposits are being mined. The *Osborne Lake* and *Anderson Lake* copper deposits and the *Dickstone* copper–zinc deposit are being prepared for production. These all occur in Amisk volcanic and sedimentary formations with the exception of the Osborne mine which is located in Kisseynew gneisses.

The *Sherritt–Gordon* (2) mine, at Sherridon, Manitoba, was operated in 1931–32 and from 1937 to 1951, concentrates being shipped to the Flin Flon smelter. During its life the mine produced 133,122 tons of copper, about 74,500 tons of zinc, 101,026 ounces of gold, and 3,218,324 ounces of silver. The ore lay along the contact between gneissoid quartzite and hornblende gneiss of the Kisseynew complex, on the southwest limb of an overturned syncline. The pyrrhotite–chalcopyrite–sphalerite ore apparently replaced a pegmatite sill that followed the contact, and formed two long, shallow lenses that are separated by a gap of 3,600 feet. The gap coincides with a cross-anticline and it is presumed that the two lenses were originally one body, the ore in the gap having been removed by erosion. The northwestern lens had a vertical extent of 500 to 800 feet; the southeastern about 250 feet. The combined length of both lenses, including the gap, was nearly 16,000 feet.

Other deposits. In the southern part of Churchill Province, west of Flin Flon and 25 miles north of LaRonge, Saskatchewan, the *Anglo–Rouyn* (6) mines copper deposit was found in 1915. It was brought into production in 1966 because of improved accessibility by road. Three *en echelon* ore zones lie in a strong shear zone at a contact between arkose and greywacke. The chalcopyrite lenses have a total of 2 million tons of ore averaging 2.4 per cent copper. Concentrates from a 900-ton-per-day plant are shipped to Flin Flon. At *Rottenstone Lake*, 90 miles north of LaRonge, production was begun in 1965 at a rate of 125 tons a day. The ore, reported to carry about 2 per cent copper and 2 per cent nickel, consists of sulphide minerals in a small mass of ultrabasic rock occurring at a 'buckle' in gneiss of Kisseynew type.

At *Hanson Lake* (7), Saskatchewan, 60 miles west of Flin Flon, production was begun in 1967. A 350-ton-per-day plant produced a zinc and a lead–copper–silver concentrate. Mill heads were reported to average 7.35 per cent zinc, 4.15 per cent lead, 0.40 per cent copper, and 4 ounces of silver per ton. The ore is in lenses in a shear zone crossing recrystallized tuff between dacite flows near a sill of quartz porphyry.

Many occurrences containing copper have been found in other parts of Churchill Province, mainly in Saskatchewan and Manitoba, at the East Arm of Great Slave Lake, and in parts of Ungava and Labrador. Some of these may respond to further testing or improved transport facilities. These regions include many belts containing rocks and structures favourable in a general way for prospecting, because they show similarities to those in which ore has been found. The Aphebian rocks of the Labrador Geosyncline and the Cape Smith–Wakeham Bay belt contain several deposits of copper, zinc, and other minerals, besides those described under nickel. For example, one deposit near Attikamagen Lake is reported

to be 660 feet long and 13 feet wide, and to have an indicated average of 6.75 per cent zinc, 1.4 per cent copper, with minor lead, gold, and silver.

Copper Deposits of Bear and Slave Provinces

No zinc and relatively little copper have been produced in these provinces. Copper was produced as a byproduct at the Eldorado uranium mine. In 1965 the *Echo Bay* (5) property east of Eldorado began producing silver and copper, 140 tons of ore per day being sent to the Eldorado concentrator. The ore averaged 2.09 per cent copper and yielded 1,617,882 pounds of copper in 1966. The deposit is described in the section on silver. Native copper from the Coppermine area had been used by natives for tools, weapons, and trade long ago, and was the object of Hearne's famous journey from Churchill in 1771. The deposits have been investigated at various times. Increased prices for copper, general improvement in northern transport and development, and interesting results from drilling have contributed to recent activity. Native copper occurrences having some of the characteristics of those of the Coppermine area were found many years ago at Bathurst Inlet. They are in strata of the Aphebian Goulborn Group overlying the Archean crystalline rocks of Slave Province. A fairly large copper–zinc deposit has been indicated in volcanics of the Yellowknife type at High Lake in the northern part of Slave Province, and a zinc–copper one at Indian Mountain Lake in its southern part.

The *Coppermine* deposits are scattered in a belt extending about 70 miles from the northeast end of Dismal Lake to Coppermine River. The rocks are basalt flows dipping 12 degrees north and forming the lower part of the Helikian Coppermine River Group. The flows are broken by numerous fractures and faults, the more prominent of which strike north to northeast, with a less prominent group striking north to northwest. The deposits are either native copper and chalcocite amygdules that are commonly associated with individual flow tops, or quartz–carbonate veins and breccia zones in the fractures and faults. Some of the veins and breccia occurrences seem to be more important than the amygdaloidal ones so far as prospecting and testing have been carried out, but less is known about the amygdaloidal deposits. Chalcocite is the main mineral in most veins and breccia zones, but in some, bornite or chalcopyrite predominates. One particular flow has high copper content where drilled in several of the deposits. Mineralized zones as much as 175 feet wide and 1,000 feet long were found. One zone is estimated to contain 62,000 tons averaging 8.78 per cent copper. In 1967, more than 10,000 claims were staked. Diamond drilling of certain deposits was done to greater depths than previously, and one 55-foot section was reported to have averaged 6.4 per cent copper. A preliminary estimate of a zone 800 feet long reportedly indicated 10,000 tons per vertical foot, averaging 3 per cent copper. The genesis of the Coppermine deposits remains to be established. Most of the chalcocite and native copper occurrences of the world are considered supergene. The Coppermine deposits have some interesting analogies with the native copper deposits of Michigan in Keweenawan strata, but the latter are characteristically native copper, little chalcocite being reported.

Zinc–Lead–Copper Deposits of Grenville Province

The New Calumet and Tétreault mines in the southern part of Grenville Province have been worked mainly for zinc. They produced smaller amounts of other metals including copper, but no mines have been worked mainly for copper. Occurrences of chalcopyrite have been reported from various places mainly in the Bancroft area of Ontario and near Portneuf, Quebec, but this may be only because prospecting and geological studies have been more extensive there than in many other parts of the province. Occurrences of zinc have been reported from various parts, mainly in the Ottawa valley and near Portneuf, Quebec. The apparent concentration of zinc, and also of lead, in the southwestern part of Grenville Province may be significant geologically because in that part of the province limestone and related rocks are abundant. Furthermore, the deposits in Grenville Province are not far from the overlapping Ordovician carbonates of the St. Lawrence Platform, in which a few small veins containing sphalerite and galena have been found. This suggests that some zinc–lead deposits in Grenville Province may be post-Ordovician in age although the few isotopic age determinations so far available for these deposits do not provide conclusive evidence of this.

The *New Calumet* (41) mine, Quebec, is at present the only producer of zinc in Grenville Province. The orebody is in a sheet-like mass of biotite–hornblende gneiss that strikes northerly and dips 35 to 40 degrees east. The gneissic body ranges in thickness from a few tens of feet to about 300 feet, and may be a bed of altered impure limestone. The presence of gahnite suggests that the rock contained zinc before being metamorphosed. The principal ore minerals are sphalerite with some galena. They are accompanied by small amounts of chalcopyrite, native gold, and silver minerals. Part of the ore is massive or nearly massive sulphide; other parts consist of sulphides disseminated through a gangue of rock minerals, mainly altered feldspar and pyroxene. A less common type of ore is a dark calcite–tremolite rock carrying crystals of auriferous galena and veined by stringers of calcite with coarse auriferous galena. The property was discovered in 1893 but production did not begin until 1943. To the end of September, 1966, it had yielded more than 417 million pounds of zinc, 117 million pounds of lead, 5 million pounds of copper, 9 million ounces of silver, and 52,000 ounces of gold. Ore reserves then were reported to average 8.64 per cent zinc, 2.58 per cent lead, and 4.88 ounces of silver per ton.

Considerable zinc and lead were obtained from the *Tétreault* (69) mine near Portneuf, Quebec. Masses of sphalerite and galena were found along the footwall of a band of altered Grenville limestones. They varied from less than a foot to more than 50 feet in width and consisted of sphalerite with minor amounts of other sulphides, or of a fine-grained mixture of sphalerite, galena, pyrrhotite, pyrite, and minor amounts of chalcopyrite. Average ore contained about 9 per cent zinc, 3 per cent lead, 0.1 per cent copper, 0.09 ounce of gold, and 8.3 ounces of silver per ton. Mining was carried on from 1912 to 1929, 1934 to 1937, and 1948 to 1955. To the end of 1954 the mine had produced about 83,000 tons of zinc, 61,000 tons of lead, several million ounces of silver, and several thousand ounces of gold.

Lead

The Canadian Shield has supplied surprisingly little lead compared to its output of other metals. An important question that awaits more rock analyses for an answer is whether the rocks themselves have an abnormally low lead content or whether they have normal amounts that rarely became concentrated as deposits. Of the total Canadian production of lead in 1966 only 1.6 per cent was derived from the Shield. This was mainly from the *New Calumet* (41) and *Coniagas* (8) mines of Quebec, worked primarily for zinc, and the rest was a byproduct of various mines. The New Calumet mine is described in the previous section. Mining of zinc–lead ore was begun at *Hanson Lake* (7), Saskatchewan, in 1967. Some lead concentrate was produced from ore from the *Chisel Lake* (8) zinc mine in Manitoba, whose ore carried 0.4 per cent lead, and the *Geco* (36) zinc–copper mine in Ontario which produced 840 tons of lead in 1966. Small to moderate amounts of lead were produced several years ago from a few mines in which lead predominated. These are described briefly below, together with occurrences on Hudson Bay that are of a different type and as yet unproductive.

About 25,000 tons of lead were produced from 1914 to 1932 from the *Kingdon* (100) mine near Galetta, Ontario. A fault, striking north 65 degrees west, cuts through limestone and other rocks of the Grenville Group. The fissure vein was filled with galena and calcite with minor amounts of other minerals. It had an average width of 5 feet, and was mined for more than 2,700 feet along strike and to a depth of more than 1,300 feet.

The *Jardun* (69) mine near Sault Ste. Marie, Ontario, was mined at intervals between 1878 and 1904. Production was resumed in 1954 and in that year it produced about 1,300 tons of lead, 550 tons of zinc, as well as 23,000 ounces of silver. The orebody consisted of sphalerite, galena, pyrite, and chalcopyrite within a band of Archean greenstone schist bounded by granite. In 1956, ore reserves were 304,000 tons averaging 3.14 per cent lead, 3.52 per cent zinc, and 1.65 ounces of silver per ton. The mine was closed in 1957.

At least 2,000 tons of ore are reported to have been taken from the *Frontenac* mine north of Kingston, Ontario, from 1875 to 1880. The galena had a low silver content and occurred in a calcite vein in sedimentary gneiss. From 1903 to 1916 more than 2.6 million pounds of lead were obtained from a similar deposit at the *Hollandia* mine north of Madoc, and small shipments are understood to have been made from other veins in these parts of Grenville Province.

The *Wright* mine on Lake Timiskaming near Ville Marie, Quebec, is of historical interest because Indians showed the occurrences to a French exploring party in 1686 and it was indicated on a map published in France in 1744. The argentiferous galena in a brecciated zone was mined in 1890–91. Diamond drilling, in 1947, was reported to have indicated 25,000 tons containing 5.9 per cent lead and 1.3 per cent zinc.

Deposits of sulphide minerals along the east coast of Hudson Bay near Richmond Gulf and Great Whale River were trenched and drilled in 1947–48 and found to be mainly disseminations of pyrite occurring as massive lenses and veins in concretionary cherty limestone and dolomite of the Aphebian Manitounuk Group. Some galena and sphalerite are present and a little chalcopyrite. Assays are reported to indicate a little gold and silver. One of the localities, 6 miles south of Little Whale River, contains a main mineralized zone 8 to 20 feet thick and another 5 to 10 feet thick. Quartz veins and pegmatite containing galena have been found at Bathurst Inlet, District of Mackenzie.

PRECIOUS METALS

Gold

Gold is found throughout the Shield and has been mined in every province or territory in Canada in which it is extensively exposed. Although considerable gold is recovered as a byproduct in the mining of copper, zinc, and nickel deposits (the Horne copper mine, for example, which is one of the largest gold producers in Quebec) the greater part of the production of gold has come from six principal gold camps: Porcupine, Kirkland Lake–Larder Lake, northwestern Quebec, Red Lake, Yellowknife, and Little Long Lac. The first three camps account for about three quarters of the total production. They are grouped in a relatively small area that extends a hundred miles

east and west of the Ontario–Quebec boundary. The other camps are in northwestern Ontario and in the Northwest Territories. All but one of the principal camps and virtually all the isolated mines are in Superior Province. The other camp, the Yellowknife, is in Slave Province. Gold deposits in Churchill and Grenville Provinces are few and have produced very little. The distribution of gold occurrences is less affected by economic or other non-geological factors than is the distribution of prospects for any other metal. Economic gold production has been feasible at one time in almost every part of the Shield and prospecting has been correspondingly widespread and thorough.

The first discoveries of gold in the Shield were made about 1866 in Grenville Province close to the early settlements in southeastern Ontario. The veins were erratic and small. Discoveries in Superior Province in the Lake of the Woods area of northwestern Ontario followed on completion of the Canadian Pacific Railway in 1886. They raised great expectations and spurred considerable promotional activity at the turn of the century. More than 100,000 ounces of gold are reported to have been produced between 1897 and 1903. More than half this amount appears to have come from the *Wendigo* (16) mine.

The Cobalt silver rush set the stage for the discovery of the major camps of the Shield and establishment of gold mining as a permanent industry. Prospecting spread out from Cobalt into the Larder Lake area where in 1906 gold was discovered. Additional discoveries were made in the rush that followed although thirty years were to elapse before significant production began. In the Porcupine area the first discoveries were made in 1909. Production began the following year and has continued to the present. In the Kirkland Lake camp gold was first discovered in 1911. Shipment of high-grade ore began in 1913 but the establishment of mines, beset by hardships and setbacks, was not achieved until 1917. The economic recession of 1930 to 1940, which gave gold mining and prospecting the double stimulus of an increase in the price of gold and a simultaneous decrease in the cost of both labour and supplies, was the principal factor that led to the development of the remaining camps, although in all but the Yellowknife camp, gold had been discovered earlier.

During the early prospecting activity in northwestern Ontario gold had been discovered in 1897 at Red Lake. Further discoveries were made during 1922–25 but it was not until 1930 with the Howey mine that production began. In the Little Long Lac area the first prospecting was for iron. Gold-bearing quartz veins were discovered in iron-formation in 1917. Renewed activity beginning with additional discoveries in 1931 culminated in production in 1938. Although gold had been discovered as early as 1923 in northwestern Quebec, the large low-grade deposits typical of the camp had to wait many years before development. Production began in 1935. Gold was first discovered in the Yellowknife area in sediments but the deposits were all small. The discovery of gold in the volcanics during mapping by the Geological Survey of Canada in 1935 led to the development of the principal deposits. Production began in 1938.

The depression brought gold to first place in value among the mineral products of Canada, taking over from coal in 1931; it maintained this position for more than twenty years until it fell behind petroleum in 1952. During the second world war, particularly after the entry of the United States in 1941, gold mining and exploration decreased. The increased activity that followed the war was short lived and with continued inflation and the fixed price of gold, gold mining has become less profitable and exploration correspondingly less attractive.

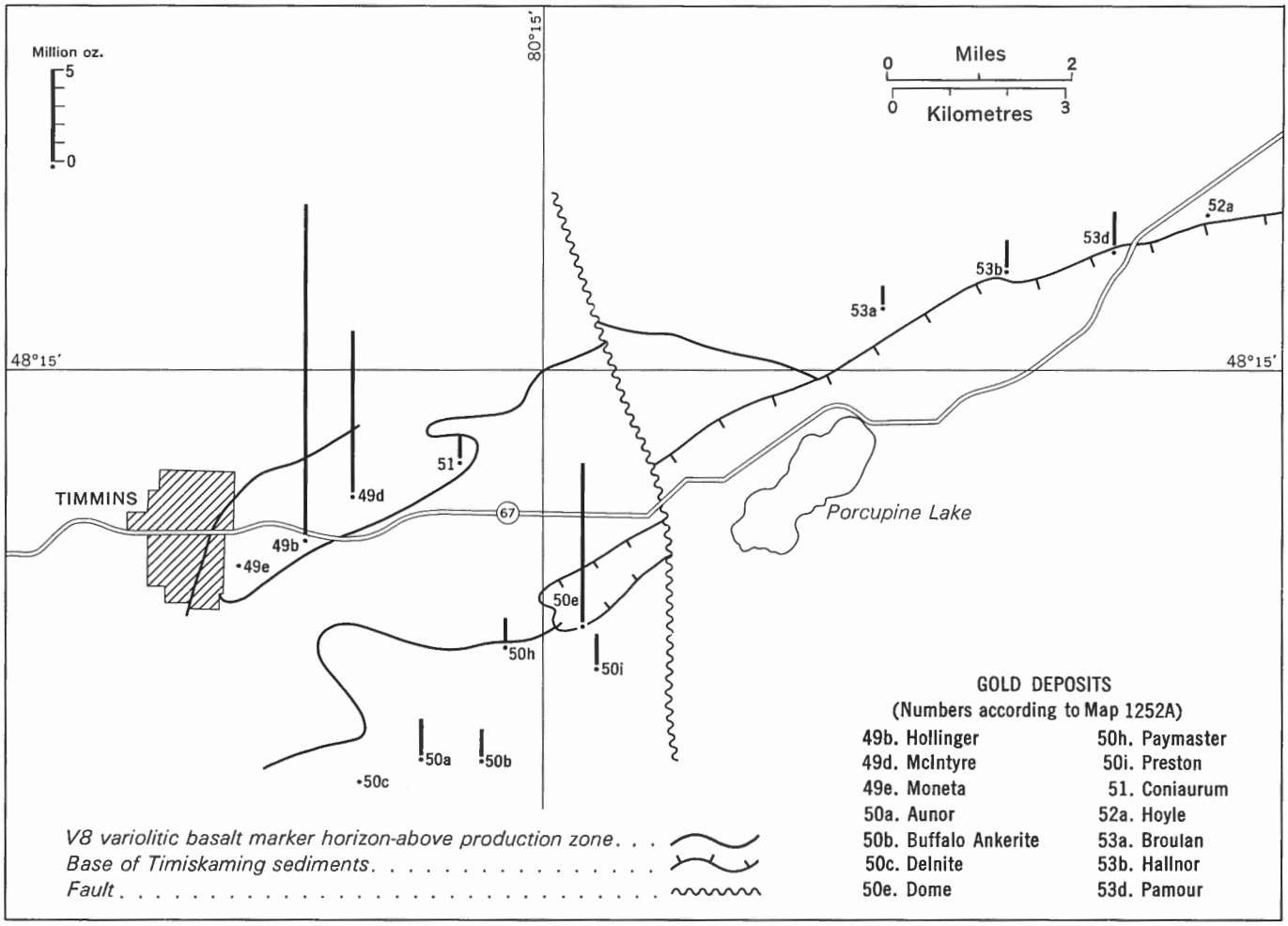
In the Canadian Shield gold deposits show a close relationship to the belts of Archean volcanic rocks. Most of the deposits occur in basic lavas, tuffs, and sediments closely associated with or derived from them, or with small intrusive bodies, principally acidic types, that intrude them. The association with volcanics is even more marked where total quantity of gold rather than number of deposits is considered. The three principal camps, Porcupine, Kirkland Lake–Larder Lake, and northwestern Quebec, lie in the extensive complex of volcanic rocks of the Abitibi belt. The volcanic area containing the Red Lake camp forms part of an interrupted chain of similar areas extending easterly from Rice Lake in Manitoba some 250 miles to Pickle Lake in Ontario forming the Uchi belt. Each area contains one or more deposits, and there are twenty-two mines altogether. Little Long Lac and Yellowknife camps are in similar but smaller belts. Outside the principal camps the scattered distribution of mines reflects the paucity and limited extent of the remaining volcanic areas. Most of these are in Superior Province. The rocks in all appear to be Archean (pre-Kenoran Orogeny). This appears true even of areas in the southeastern part of Churchill Province.

In some camps the distribution of mines appears to be partly stratigraphically controlled. For example, the most productive parts of the Porcupine camp lie in or close to No. 95 flow or its equivalents. A similarity in the stratigraphic sequence of volcanics and sediments of the Red Lake and Porcupine camps has been noted by Ferguson (1966). However, the formation of orebodies and the distribution of ore shoots within them is clearly controlled by faults or other local structural features. In some camps, such as Kirkland Lake–Larder Lake, faults exert a dominant influence on the camp as a whole. Of the two controls the structural was given more attention during the period of most active exploitation of the mines. Recognition of the structural controls within a camp provided a very effective tool to predict and locate ore shoots. It is only recently that the stratigraphic component in gold ore distribution has been widely recognized and its implications considered.

Superior Province

Porcupine camp, Ontario (Fig. V-28). The oldest rocks are the Archean Keewatin volcanics that form a layered sequence estimated to have a maximum thickness of 10,500 feet. The basal Tisdale Group consists of a conformable succession of pillowed, flow-structured, and massive meta-basalts. Several persistent stratigraphic units can be recognized, in particular a group of variolitic and fragmental flows in the upper part. Fine-grained, cherty, or carbonaceous sediments are present locally between flows and are particularly well developed at certain stratigraphic positions. Some 60 to 300 feet of sediments of this type overlie the Tisdale Group and are, in turn, overlain by as much as 2,000 feet of acid pyroclastic volcanics, mainly latite breccia, comprising the Krist Group. Interbedded with the pyroclastics are banded chert and carbonaceous slate, and above them is a sequence of argillite, black slate, and minor greywacke of the Hoyle Group variously estimated at 3,000 to 7,000 feet thick. Minor discordances occur within the volcanics and between the volcanics and the predominantly sedimentary part of the sequence. However, for a predominantly volcanic assemblage the general conformability of the lower part is remarkable and indicates uniform accumulation of flood basalts in a subaqueous or submarine environment. Overlying the Keewatin rocks with marked angular discordance is the Timiskaming Group, 1,500 feet thick, and consisting principally of slate, greywacke, quartzite, and conglomerate. The conglomerate is present at and near the base of the group and also near the base of the lowest quartzite.

Both acidic and basic intrusions are present. Most important from the point of view of the gold deposits are porphyry stocks that intrude the Tisdale volcanics, but which are considered by some to be equivalent in age to the volcanics of the Krist Group. The gold deposits of the Hollinger and McIntyre mines are grouped in the volcanics around these intrusions, although few auriferous veins penetrate far into the porphyry. The porphyry masses are pipe-like, in their lower parts at least, and differ slightly one from another in composition or texture. There are seven such masses at the McIntyre mine for example and, although they approach one another at depth, separated in some instances only by thin screens of schistose lava, they tend to maintain separate identities.



GSC

FIGURE V-28. Gold deposits of the Porcupine camp, Ontario (by D. R. E. Whitmore).

The ore comprised the quartz veins and the adjacent mineralized schist. Many of the veins formed complex sheeted zones or *en echelon* swarms and were commonly contorted. Thus the orebodies were complex units defined largely by assays. Mineralized schist predominated over quartz. The quartz veins contained ankerite, sulphides, tourmaline, minor albite, and scheelite. The adjoining schist was silicified, carbonatized, and contained disseminated pyrite. The sulphide minerals in the veins were pyrite, pyrrhotite, sphalerite, chalcopyrite, and galena. Locally arsenopyrite occurred in some abundance, as for example in the deeper parts of the Hollinger and Delnite mines. For the most part gold appeared in native form. Tellurides occurred in the Hollinger, McIntyre, and some other properties but were not characteristic of the camp. A few orebodies, as at the Moneta, consisted predominantly of sulphides, apparently a replacement in a volcanic breccia.

Volcanic rocks of the Porcupine area form an easterly plunging syncline. In the eastern part of the syncline, centred somewhat south of its axial portion, is a discordant trough of Timiskaming sediments with basic and ultrabasic intrusions. Mines of the camp fall into three groups: those along the north limb of the syncline in the volcanics, those along the south limb, and those near the unconformity at the base of the Timiskaming Group.

Hollinger (49b), the largest mine of this group and of the camp as a whole is developed towards the western end of a complex anticlinal arch on the north limb of the main syncline which altogether extends some 3 miles easterly from Moneta to Coniaurum. Ore is contained in a complex system of veins in a zone 6,000 feet long and 4,000 feet wide forming a flat arc that strikes N60°E and plunges 55°E, approximately conformable to the attitude of the bordering Pearl Lake porphyry. Orebodies included: (1) well-defined continuous quartz veins locally branching, pinching, and swelling; (2) sinuous veins either replacing drag-folds or themselves folded; (3) tabular vein zones consisting of parallel stringers of quartz in pyritized lava; and (4) *en echelon* vein zones. Many veins show a complex history of fracturing with later introduction or redeposition of quartz with gold.

McIntyre (49d), another large producer, adjoins the Hollinger to the northeast and shares the same vein system. Many of the veins lay beneath the pitching porphyry mass, not reaching the surface. Ore in the *Coniaurum* (51) mine east of McIntyre is in veins that occur mainly in dacite flows on the flanks of a porphyry mass, or in deep re-entrants within it. In the upper levels the veins extended 600 to 700 feet from the porphyry. This distance decreased with depth. Ore at the *Moneta Porcupine* (49e), adjoining Hollinger at the west end of the camp, was largely a heavy sulphide replacement consisting of as much as 13% pyrite in brecciated pillowed andesite and in andesite with conspicuous leucoxene. Ore from the two flows showed consistent differences both in grade and texture.

Of the group on the south limb, the *Dome* (50e) was comparable in size with the Hollinger and McIntyre. It lay athwart the unconformity separating the Keewatin volcanics from the Timiskaming Group. The latter formed a narrow synclinorium extending obliquely from the main syncline that plunges northeast. The original orebodies, more than thirty in number, lay in the sediments. They consisted of quartz and mineralized wallrock. Most appeared to be associated with conglomeratic facies. Some were very large, as long as 600 feet, and many were very irregular in outline. Some orebodies crosscut the bedding but most did not. A few showed a relationship to embayments in the sediment-volcanic contact. Later orebodies were developed in veins in the volcanics below the unconformity. The veins formed as related groups controlled by an anticlinal nose in the greenstone. One well-developed type consisted of an *en echelon* swarm within a dacite flow.

Preston (50i) mine, southwest of the Dome, is developed on a drag-fold on the south limb of the main syncline. Two quartz–feldspar porphyry stocks comprise the main host for the orebodies which are of three kinds: (1) stockworks of quartz–carbonate stringers with pyrite and minor arsenopyrite in the porphyry, (2) single quartz and quartz–tourmaline veins in the porphyry and extending into the volcanics, and (3) shatter zones in cross-faults. *Paymaster* (50h) lay west of Preston on the south limb proper. Much of the ore lay close to a tuffaceous flow top and was zoned relative to a quartz porphyry. The west end of the veins was barren. An intermediate section showed refractured quartz–carbonate veins with gold. The inner portion showed late, fine-grained, well-mineralized quartz.

Other mines on the south limb included the *Buffalo Ankerite* (50b), which lay along both sides of a synclinal cross-fold. The orebodies were in quartz–ankerite–tourmaline veins and in pyritized dacite. The *Aunor* (50a) and *Delnite* (50c) mines were developed on brittle flow rocks between zones of chlorite or talc chlorite schist. At the Aunor both parallel and branching veins were present in pillowed andesite. At Delnite the veins formed a ladder-like series. In all three mines plunge of the orebodies was to the west. To the east orebodies at the *Broulan* (53a), *Hallnor* (53b), *Pamour* (53d), and *Hoyle* (52a) are developed on both sides of the unconformity at the base of the Timiskaming sediments. Dunbar (1948) notes that the orebodies occur in the general vicinity of the No. 95 flow where it was eroded.

Kirkland Lake–Larder Lake camp, Ontario (Fig. V-29). The mines of the camp lie in an east-west trending trough of Archean sediments, tuffs, and trachytes set in older basic volcanics. The rocks dip steeply and face south towards a complex and persistent zone of schist, the Larder Lake break, which forms the south boundary of the trough. The break, thought to be a thrust fault, is interrupted by Kenoran acidic intrusions and is therefore an early struc-

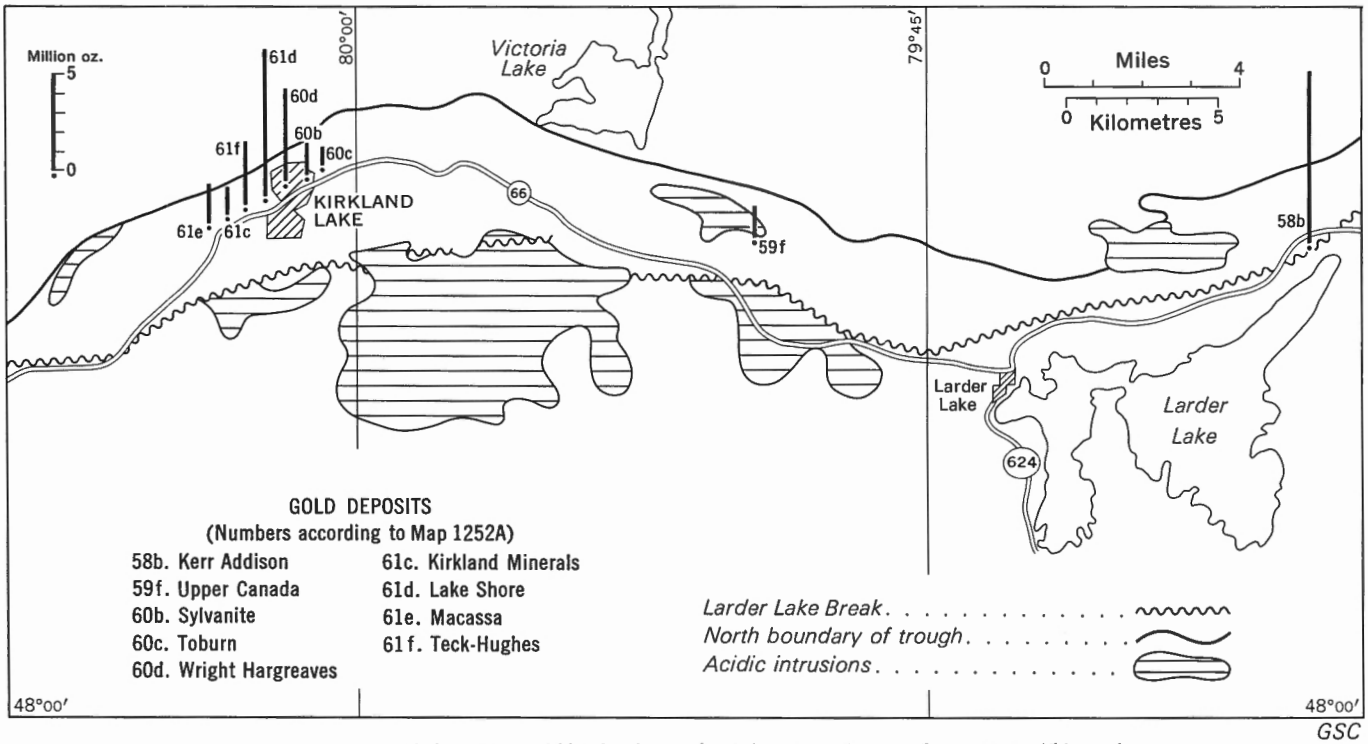


FIGURE V-29. Gold deposits of Kirkland Lake–Larder Lake camp, Ontario (by D. R. E. Whitmore).

ture that may be related in some way to the formation and filling of the initial trough itself. In many ways it is analogous to the Bouzan–Cadillac break of Quebec; the two are also in alignment. Principal mines of the region fall into three groups. To the west is Kirkland Lake proper a continuous series of seven mines lying just north of the axial portion of the trough. In the centre of the trough 9 miles to the east is Upper Canada Mines. At the south boundary on the Larder Lake break is Kerr Addison.

At Kirkland Lake seven mines—the *Macassa* (61e), *Kirkland Minerals* (61c), *Teck-Hughes* (61f), *Lake Shore* (61d), *Wright Hargreaves* (60d), *Sylvanite* (60b), and *Toburn* (60c)—are developed on veins in a composite fault zone comprising the Kirkland Lake fault, or main break, two important branch faults in the middle of the camp, and a number of smaller splays. The zone cuts sediments and a complex of syenitic intrusions elongated parallel with the sediments and pitching west. Ore is found in all rocks traversed by the fault system and consists of quartz-filled fissure veins, breccias, and gash veins tributary to the fault fissures.

Quartz with some calcite is the principal gangue although it does not form a large part of the ore as milled. Gold is native in gangue and in fractures in the wall-rock. Calaverite and petzite as well as other tellurides are relatively common. Some gold occurs with fine pyrite that constitutes as much as 2 per cent of the ore. The longest orebody, shared by the Teck-Hughes, Lake Shore, and Wright Hargreaves mines, was 6,400 feet long horizontally of which 2,600 feet was on the main break and 3,800 feet lay on the north vein fault. Average stoping width was 6 feet, although a few stopes were as much as 100 feet wide. Largest shoots were vertical in long section although others pitched west in conformity with the trace of geological contacts on the break.

Upper Canada (59f) mine occurs on a subsidiary break cutting tuff, trachyte, trachyte porphyry, and a syenite porphyry. Some evidence suggests that the intrusions are partly contemporaneous with the tuff. The break consists of branching and subparallel fractures, some filled with gouge. Orebodies are complex silicified zones with tourmaline, calcite, and quartz of several ages. Gold and tellurides occur in the quartz and with pyrite in the wall-rock.

The *Kerr Addison* (58b) is the largest single producer in the camp. Orebodies lie in and just south of a zone of carbonatized volcanics 600 feet wide that forms part of the Larder Lake break. They include irregular lenticular masses of quartz stringers within the break and silicified flows with auriferous pyrite immediately south of it.

Northwestern Quebec (Fig. V-30). Eight of the ten principal gold producers of Quebec lie within an east-west trending area 16 miles long and 3 miles wide centred on Lac DeMontigny. The west end of the area containing four of the mines—*Canadian Malartic* (32c), *Barnat* (32a), *East Malartic* (32d), and *Malartic Gold Fields* (33a)—is intersected by a persistent narrow west-northwesterly trending zone of schist, the Cadillac break. For some 20 miles west of the intersection numerous gold prospects occur along the Cadillac break, but only one medium-sized producer, the *O'Brien* (25b), and a few small ones,

the *Lapa Cadillac* (26b), *West Malartic* (26c), and *Pandora* (25c), are productive. East of the intersection there is little gold showing along the Cadillac break. The gold-bearing area extends east and north of the break, centred on Bourlamaque batholith and several smaller granodiorite stocks around or within which the deposits are clustered. These include the remaining four larger deposits—*Lamaque* (34b), *Sigma* (34d), *Sullivan Consolidated* (31b), and *Siscoe* (31a)—as well as several smaller ones such as *Perron* (36b) and *Bevcon* (37).

Three of the four mines at the intersection are grouped around the western end of an elongated peridotite mass 7 miles long that borders the Cadillac break. *Barnat* and *East Malartic* orebodies lie in small syenite porphyry and diorite intrusions close to the peridotite contact. *Canadian Malartic* is more distant in sediments and porphyry south of the break. The fourth mine, *Malartic Gold Fields,* lies in schist re-entrants along the northern contact of the peridotite, the orebodies having formed principally in small diorite pipes and dykes within the schist. The orebodies in all four deposits form a coherent group similar to one another and unlike those east of the Cadillac break. Although gold has been found in all rocks present along the break the orebodies are restricted to greywacke, diorite, and porphyry. They occur in zones of shattering, silicification, and pyritization. Small intrusive pipes or dykes surrounded by schist may be entirely converted to ore. Shattering in the larger intrusions and greywacke is more localized, but commonly occurs at the contacts. Drag-folds control the position and shape of some orebodies.

In the shattered zones the quartz veins are barren; the gold is in the altered wall-rock and associated with pyrite, virtually the only sulphide present. The amount of pyrite, which may reach 20% by volume, varies with the host rock, and in a gross way the grade of the ore varies similarly. Diorite contains the most pyrite and orebodies within it average 0.17 to 0.23 ounce per ton. Porphyry contains only a quarter as much pyrite, and ore averages 0.11 to 0.17 ounce per ton. Greywacke contains somewhat less pyrite, the ore averaging 0.09 to 0.14 ounce per ton. Silver content of the ore also varies with the host rock. The highest Au:Ag ratio is 1:1; these orebodies occur in the greywacke.

The deposits of the eastern part of the camp are more scattered. They consist of well-defined quartz veins with spectacular amounts of visible gold. Arsenopyrite as well as pyrite is commonly present.

Lamaque, the largest single gold producer in Quebec is developed on a small, oval, concentrically zoned intrusion in tuff. Granodiorite forms the centre and quartz–diorite the periphery. The intrusion plunges steeply northeast and is cut by a series of faults dipping gently southwest within which the veins are associated. The *Sigma* mine comprises a group of south-dipping veins that forms a shatter zone along the north side of a shear zone in volcanic rocks. *Siscoe* mine lies at the western extremity of the Bourlamaque batholith. Most of the ore came from quartz veins within the intrusion, the balance from a zone of shearing along its south margin. The *Sullivan Consolidated* deposit consists of quartz veins filling fractures cutting the western nose of the Bourlamaque batholith. Sulphides in the veins include chalcopyrite, sphalerite, and

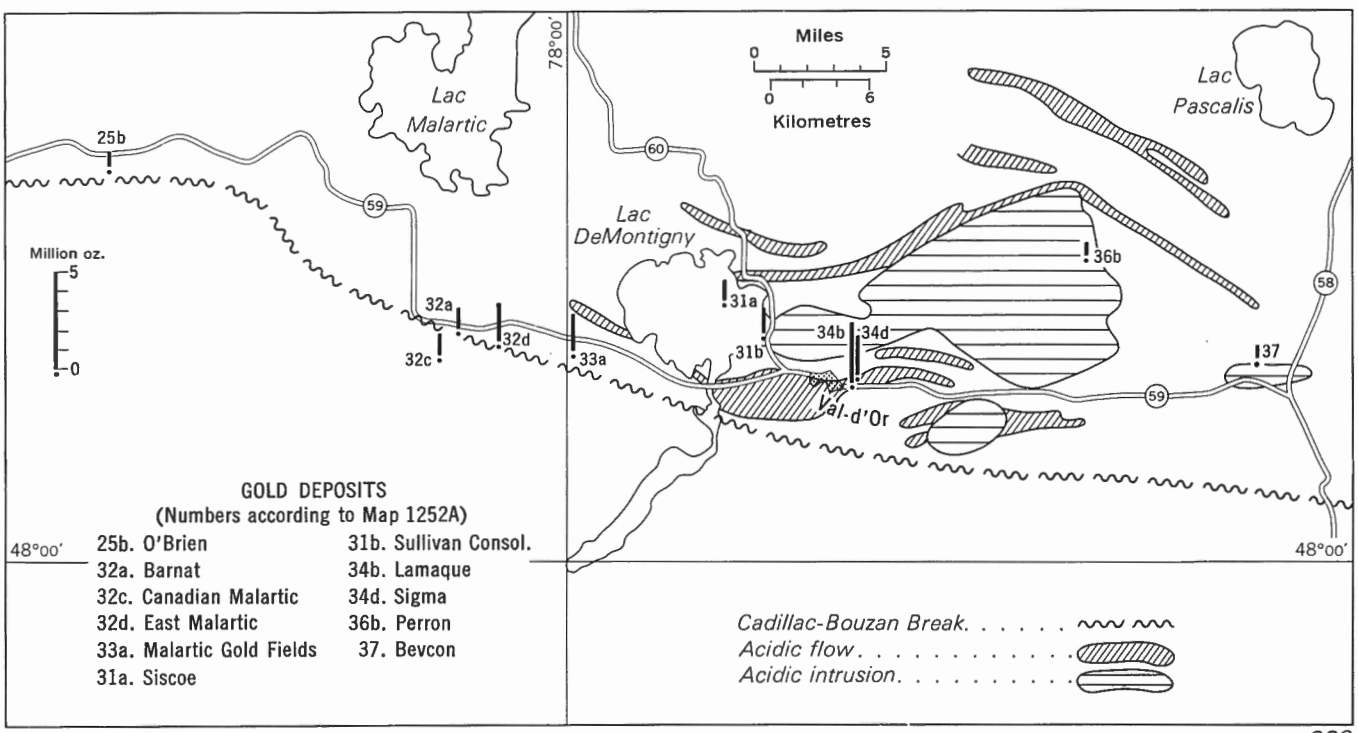


FIGURE V-30. Gold deposits of northwestern Quebec (by D. R. E. Whitmore).

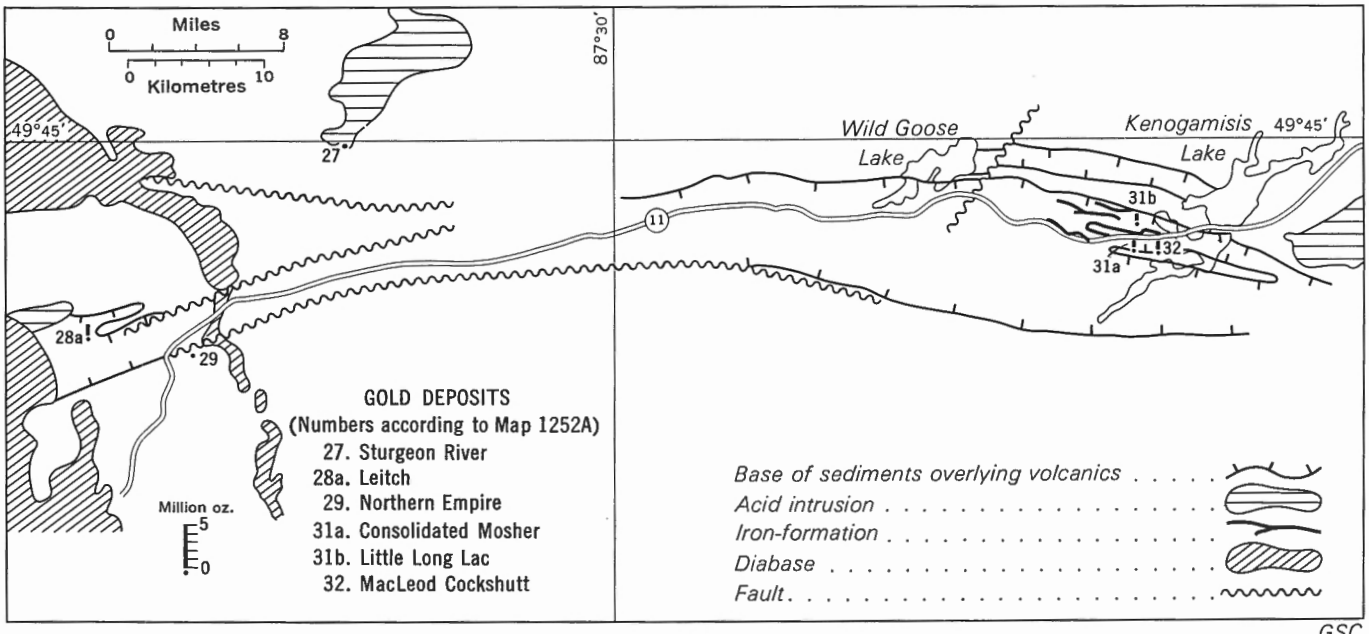


FIGURE V-31. Gold deposits of Little Long Lac camp, Ontario (by D. R. E. Whitmore).

galena. On the eastern edge of the Bourlamaque batholith, the *Perron* consists of quartz veins in flat fractures subsidiary to a steeply dipping barren shear zone cutting into the granodiorite. At *Bevcon,* auriferous quartz veins with pyrite and minor bismuth lie in a sheared and fractured zone between the north side of a small granodiorite body and a porphyry dyke.

Little Long Lac camp, Ontario (Fig. V-31). Extending eastward for 110 miles from the flat-lying, lower Proterozoic sediments and diabase sills of Nipigon basin is a belt of greenstone and poorly sorted clastic sediments about 20 miles wide. Rock types are asymmetrically arranged in it with greenstones predominating in the northern part, sediments and mica schists in the southern. Gold deposits occur in the central and western part of the area. Three of the four largest mines, *Consolidated Mosher* (Hardrock) (31a), *MacLeod-Cockshutt* (32), and *Little Long Lac* (31b) lie at the eastern limit of the camp. Gold is found in two principal types of orebodies and a number of subtypes. All types are present in the most easterly mines; those to the west have less variety. One principal type consists of sulphide replacement masses, with as much as 35% quartz, in iron-formation and greywacke. Pyrite, and arsenopyrite are the principal sulphides present in all orebodies. The other type consists of quartz stringer zones in the sediments. Some stringer zones are remarkably persistent although the individual veins may be narrow.

In the west the mines are more scattered and are developed on simple quartz–gold veins that contain no other minerals. The veins cut greywacke at the *Leitch* (28a), greenstone at the *Northern Empire* (29), and greenstone and granodiorite at *Sturgeon River* (27) mine.

Red Lake camp, Ontario (Fig. V-32). In the Red Lake area of northwestern Ontario twelve mines have produced gold since 1930. The deposits occur in an area of complexly folded greenstone, 35 miles long by 15 miles wide, whose long axis trends somewhat north of east. Two of the largest mines, *Campbell* (9b) and *Dickenson* (9d), and one of the smallest, *H. G. Young* (8), share the principal vein system of the camp, accounting for almost half the gold produced. The system is developed in andesite lavas

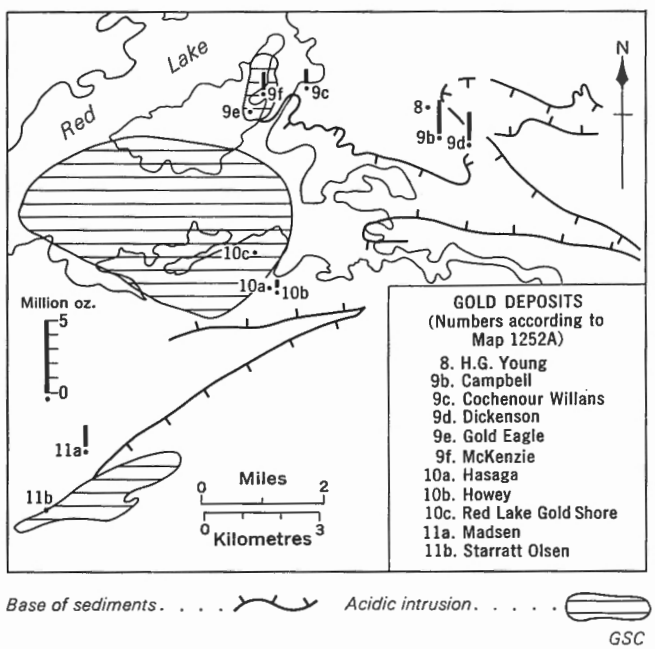


FIGURE V-32. Gold deposits of Red Lake camp, Ontario (by D. R. E. Whitmore).

adjacent to a persistent band of rhyolite tuff. The main lode, a sheeted swarm of quartz veins, extends a considerable distance from the tuff but a number of other small veins lie along or close to the tuff. The second largest producer, the *Madsen* (11a) mine, is also associated with tuff. The ore occurs in a series of *en echelon* lenses of silicified rock with feldspar porphyry, possibly originally tuffaceous sediment, that lies between quartz porphyry and talc schist. The presence of garnets indicates a higher metamorphic grade than elsewhere in the camp. Other deposits with similar associations are the *Cochenour* (9c) and nearby *McMarmac* (9g) in the northeastern part of the camp, and the *Starratt Olsen* (11b) at the southwestern extremity of the greenstone belt.

The remaining deposits are more closely related to intrusions. In contrast to the deposits of the volcanic association previously described, which are typically either cherty quartz and carbonate replacement zones or shear zones with heavy sulphides, the deposits with the intrusive rocks are well-defined quartz veins and stringers with relatively little sulphide mineralization. Three are related to the main granite stock of the camp. At the *McKenzie* (9f) the quartz veins occur in fractures and shear zones along a thrust fault cutting the stock and dipping 35°W. Veins at the *Gold Eagle* (9e) crosscut a narrow granitized sedimentary inclusion in the granite. Orebodies at *Red Lake Gold Shore* (10c) are pipe-like masses of quartz within the stock. Except for the McKenzie all are small.

The *Howey* (10b) and *Hasaga* (10a) mines were developed on an intricate stockwork of quartz stringers in a narrow part of a quartz porphyry dyke cutting a tongue of andesite breccia that projected into the main granite stock.

The deposits associated with the intrusions are generally lower grade than those closely associated with the tuff. They are also characterized by tellurides that are absent in the deposits in the tuff, but on the other hand they contain little arsenopyrite or pyrrhotite which are abundant in most tuff deposits.

Several of the volcanic segments forming part of the Uchi belt that extends east and west of Red Lake contain more than one mine. With the exception of those in the Bissett area in Manitoba and Pickle Lake in Ontario the mines have been small and short lived.

In southeastern Manitoba, the *San Antonio* (18b) and *Forty-four* (18a) mines are developed on a thick part of a diabase sill that dips 45° northeasterly within the tuff of the Rice Lake Group. The veins occur in two complementary sets, ending near the margins of the sill. One is a set of shear zones across the strike of the sill and dips 60° northwest. The other, a set of fracture zones subparallel with the strike of the sill, dips vertically. The veins consist of quartz and carbonate with minor chlorite and albite. Pyrite and minor chalcopyrite occur in the quartz with native gold. The San Antonio mine occurs in the Rice Lake volcanics northwest of an elliptical stock of quartz diorite 9 miles long. Southeast of the stock are three much smaller producers. *Ogama-Rockland* (19) consists of quartz veins with sparse pyrite and chalcopyrite in shear zones cutting quartz diorite near the margin of the mass. *Central Manitoba* (20) occurs in the volcanics and consists of a group of quartz veins, some *en echelon*, in a zone cutting across andesite and gabbro between two barren carbonatized shear zones. At the *Gunnar* (21) mine, quartz veins occur in basalt and andesite near the end of an irregular granite dyke.

One hundred and sixty miles east of Red Lake, in Ontario, in an isolated segment of the Uchi belt, are two adjoining mines. On the west the *Central Patricia* (4) orebody lies in a band of iron-formation, the sulphides filling cross-fractures. The cross-fractures contain pyrrhotite, arsenopyrite, and chlorite and are from a quarter inch to ten inches wide. Gold appears related to the chlorite. At the *Pickle Crow* (3) mine to the east are two main producing veins, one crosscutting the iron-formation at a low angle and the other cutting a quartz porphyry. The two veins are parallel with each other both in strike and dip. They consist of quartz and gold, with tourmaline and very minor pyrite, pyrrhotite, and chalcopyrite.

Slave Province

Yellowknife camp (Fig. IV-12). More than 80 per cent of the gold produced in the Northwest Territories comes from a small area of volcanics of the Archean Yellowknife Group on the west side of Yellowknife Bay. The volcanic rocks form a steeply dipping homoclinal sequence facing southeasterly, bounded on the west by a large granite mass. Small stocks of granite intrude the volcanics. The principal orebodies lie within shear zones transecting the volcanics at high angles. They consist of chlorite schist and may be as much as 200 feet wide. The zones form a single system of related shears, now separated into two parts by a large northerly trending post-mineralization fault, the West Bay fault. Horizontal displacement on the fault in the order of several miles is west side south. In vertical section the shear zones show complex curving outlines, as if folded, and it has been suggested they represent early thrust faults subsequently deformed. However, the schist composing them shows in many places a remarkable consistency in orientation regardless of the attitude of the zone and would appear to have developed after folding.

The most important orebodies consist of as much as 90 per cent auriferous quartz intimately mixed with sericite schist, and carbonate within the shear zones. Fine pyrite is the principal sulphide with arsenopyrite, and minor sphalerite, chalcopyrite, sulpharsenides, and sulphantimonides. The orebodies appear somewhat related to sharp changes in attitude of the shear zones. This type of orebody is characteristic of the *Giant* (8) mine east of the West Bay fault and of the deeper levels of the *Con* (9a) mine west of it. In the early days of the camp at the *Con Rycon* (9c), and *Negus* (9b) mines relatively narrow shear zones subsidiary to a main shear contained a somewhat different, higher grade ore. The minor shears

were more chloritic, with narrow, fractured, and vuggy quartz veins that contained sulphosalts and in places spectacular quantities of visible gold.

Other deposits are numerous in the sediments overlying the volcanics, but very few even approach mineable dimensions. For the most part these are simple bedded veins or saddle reefs as at the *Consolidated Discovery* (6), *Thompson-Lundmark* (7), and the *Ptarmigan*. The Consolidated Discovery deposit lies in argillaceous and quartzose greywacke near tuffaceous volcanics on a nose of amphibolitized pillow lava. The original vein lies conformably within the sediments in a steeply northward plunging and westerly pitching fold, which becomes broader with depth. Additional ore lies in a shatter zone within the pillow lava. Pyrrhotite is the chief metallic mineral with minor pyrite, galena, arsenopyrite, and chalcopyrite.

Churchill Province

Volcanic areas are fewer in Churchill than in Superior Province. Although gold occurrences are found in all areas, only that extending between Flin Flon and Herb Lake, Manitoba, has mines. In the Flin Flon–Herb Lake belt the structure is complex. The belt is divided by cross-folds and igneous intrusions into many segments. Three mines occur in separate isolated segments. The largest, the *Nor-Acme* (9), produced more than half a million ounces of gold from 1949 to 1958. The orebody occurred along a thrust fault that dips 45°N separating feldspathic sediments and volcanics of the footwall and coarse mafic pyroclastic rocks of the hanging-wall. The ore consisted of disseminated gold in silicified and carbonatized country rocks also containing arsenopyrite, pyrrhotite, and pyrite. Vein quartz did not exceed 5 per cent. Ore shoots where the mineralized zone thickened to mineable dimensions, were controlled by rolls in the fault plane that pitched northeasterly at 30 degrees.

At *Laguna* (11) and *Gurney* (7) the gold occurs in quartz veins. The vein at Laguna exceeded 2,100 feet in length and paralleled the long axis of a quartz–feldspar body. It contained abundant arsenopyrite with minor galena, sphalerite, pyrite, and gold and, at depth, pyrrhotite and chalcopyrite. At Gurney the veins occurred in shear zones in tuff and contained pyrite, pyrrhotite, chalcopyrite, sphalerite, and galena. Both these mines were much smaller than Nor-Acme, the Laguna about one-tenth smaller and the Gurney one-twentieth.

At the *Box* (5) mine on the north shore of Lake Athabasca, a granitic sill or a granitized sedimentary layer, up to 120 feet thick, in Tazin sedimentary gneisses, contains a network of auriferous quartz veins with pyrite, galena, sphalerite, and free gold. The overall grade was less than 0.05 ounce per ton.

Grenville Province

Scattered gold occurrences are found throughout southeastern Ontario northeasterly of Havelock. In a general way the area corresponds with the southern limit of a complex chain of small volcanic belts stretching across this part of Grenville Province. Only a few of the deposits are associated with the volcanics directly, although most occur in diorite that may in part be derived from volcanics. Surface enrichment in the weathered zone was a common feature. The deposits were all small. At *Cordova* (112) veins of quartz and carbonate containing pyrite lay in shear zones cutting diorite of variable composition and texture at the margin of a large mass of gabbro. At *Deloro* (111a) shear zones cut diorite, possibly metamorphosed lava, near the margin of a granitic stock. Veins consisted of quartz and carbonate with arsenopyrite and native gold.

Silver

In 1967 some 25 million ounces of silver were produced from the Canadian Shield, equivalent to 64 per cent of the silver production of Canada. The principal types of deposits from which silver is obtained and their production in 1967 follow:

1. *Skarn deposits,* mined essentially for their copper, zinc, and lead contents. Silver is a byproduct. These deposits produced 1.2 per cent of the total silver obtained from the Shield.

2. *Massive Ni–Cu sulphide deposits* (Sudbury type), mined essentially for their nickel, copper, and cobalt contents, with silver as a byproduct. Some 6.5 per cent of the total silver from the Shield was obtained from these deposits.

3. *Massive Cu–Zn sulphide deposits* (Flin Flon type) mined essentially for their copper and zinc contents, with silver as a byproduct. About 61 per cent of the total silver obtained from the Shield came from these deposits.

4. *Gold–quartz deposits,* mined essentially for their gold content. Silver is a byproduct. These deposits produced 1.6 per cent of the total silver obtained from the Shield.

5. *Native silver–Ni–Co–arsenide deposits.* Some of these, such as the ones at Port Radium, were mined essentially for uranium with silver as a byproduct. Others, such as those in the Cobalt, South Lorrain, and Gowganda areas, are mined essentially for the silver with cobalt, copper, and arsenic as byproducts. This type of deposit provided 26.7 per cent of the silver obtained from the Shield.

6. *Sulphide vein deposits,* containing pyrite, sphalerite, chalcopyrite, and argentiferous galena. Most of these have been worked in the Thunder Bay district of Ontario mainly for their silver content. There was no silver production from deposits of this type in 1967.

The four types of deposits that yield silver as a byproduct are described in more detail in other sections and

need only be considered briefly here; those mined essentially for silver are described in more detail below. A comprehensive report on silver, containing a metallogenic map on which are recorded the location of the principal silver deposits and prospects in the Shield, has been published by Boyle (1968).

Deposits Yielding Silver as a Byproduct

Skarn Deposits

Most deposits in this category occur in Grenville Province and are associated with the common skarn minerals that are developed in limestones, dolomites, or calcareous schists. The form of the deposits is irregular and bunchy, but many are tabular, and follow certain beds that have been selectively altered.

The deposits usually contain a characteristic suite of relatively high-temperature Ca–Mg–Fe–silicate and carbonate minerals such as garnet, epidote, vesuvianite, diopside, tremolite, scapolite, wollastonite, and crystalline calcite and dolomite. The sulphide minerals are generally chalcopyrite, bornite, pyrite, pyrrhotite, galena, sphalerite, and molybdenite. Magnetite and hematite occur in many of the deposits. The silver is characteristically associated with galena in most deposits, although in a few deposits of this type it may be associated with copper sulphides or native gold.

The *Tétreault* (69) deposit, 45 miles west of Quebec City, produced several million ounces of silver before production ceased in 1954. The silver was associated principally with galena and probably occurred mainly in tetrahedrite. Electrum and native silver have also been reported and accounted for part of the silver and gold content of the ore. The grade was 7 to 10 per cent Zn, 1 to 3 per cent Pb, 0.09 ounce Au, and 5 to 9 ounces Ag per ton. The orebodies were irregular lenses and tabular bodies of disseminated and massive sphalerite, galena, pyrrhotite, and chalcopyrite in silicified limestone and metasomatized paragneiss consisting of tremolite, diopside, wilsonite, scapolite, talc, and other Ca–Mg–Fe silicates and carbonate minerals.

The *New Calumet* (41) deposit on the Ottawa River is a tabular body of massive and disseminated sulphides in a highly altered impure limestone now essentially composed of pyroxene, tremolite, feldspar, chlorite, garnet, and calcite. The principal sulphides are pyrite, pyrrhotite, sphalerite, and galena. The silver content ranges between 5 and 10 ounces to the ton, and there is about 0.05 ounce gold to the ton. The silver is closely associated with the galena and is present in various silver-bearing minerals, including tetrahedrite, acanthite, and polybasite. The gold is mainly in the native form associated with all the sulphides and as minute grains in the silicate gangue. One variety of ore, characterized by fine-grained pyrite and pyrrhotite in coarse calcite with feldspar and tremolite, is relatively rich in native gold. This native gold appears to have been introduced late in the mineralization history of the deposit. It carries minor amounts of silver.

Massive Ni–Cu Sulphide Deposits (Sudbury Type)

These deposits contain pentlandite, pyrrhotite, and chalcopyrite within or associated with basic rocks of the gabbro clan. They are described in the section on nickel. Most silver is apparently in the chalcopyrite in many of the deposits and is recovered in refining.

Massive Cu–Zn Sulphide Deposits (Flin Flon Type)

These deposits are mined essentially for their copper and zinc but supply by far the largest amount of silver won from the Canadian Shield. They are described in the section on copper–zinc. The deposits are generally large lenses or irregular bodies consisting of pyrite, pyrrhotite, chalcopyrite, sphalerite, and minor galena. Much of the silver is in chalcopyrite or in a variety of accessory minerals such as native gold, electrum, native silver, galena, tetrahedrite, tennantite, and other sulphosalts. One of the largest massive sulphide bodies with a huge potential for silver production in the Canadian Shield is the recently discovered *Kidd Creek* (48) deposit near Timmins, Ontario. This deposit is said to contain some 60 million tons of ore averaging 4.85 ounces Ag per ton, 1.33 per cent Cu, and 7.08 per cent Zn. The total content of silver is, therefore, of the order of 300 million ounces, an amount approaching that produced from the whole of the Cobalt camp. The sulphide minerals are mainly pyrite, chalcopyrite, and sphalerite. In addition there are small amounts of bornite, covellite, digenite, stromeyerite, pyrrhotite, marcasite, galena, and arsenopyrite. The silver occurs mainly native in small inclusions in the pyrite.

Gold–Quartz Deposits

Most of these deposits occur in shear zones, faults, and fractures in the Archean greenstone belts of the Canadian Shield or in shear zones, fractures, drag-folds, and saddle reefs in the greywackes and slates associated with the greenstones. The silver is mainly in native gold in a gangue of quartz and carbonates. Some silver is present in tetrahedrite and other sulphosalts in some deposits and in tellurides in others. One or two deposits of this type contain a little of their silver as the native metal.

Deposits Mined Essentially for Silver

Deposits mined essentially for their silver content include the native silver–Ni–Co–arsenide deposits of the Cobalt, South Lorrain, Elk Lake, and Gowganda areas of Ontario, and those in the Great Bear Lake area, District of Mackenzie. The deposits of the Thunder Bay district of Ontario are mainly of the sulphide vein type, although the

orebody of the principal producing mine, *Silver Islet*, was of the native silver type.

Cobalt District, Ontario (96–98). Principal mines operating in the Cobalt area, in 1967, were Glen Lake, Hiho, Silverfields, Silver Town, Silver Regent, Agnico, Sudbury Contact, Deer Horn, and Langis. The silver deposits were discovered in 1903 during the building of the Timiskaming and Northern Ontario Railway, and since that time the area has yielded a large part of the Canadian silver production and some cobalt, copper, nickel, arsenic, and bismuth. The bonanza years lasted from 1904 until about 1930, after which there was a gradual decline. Today the camp is experiencing a rejuvenation. Cumulative production of silver to the end of 1967 was about 500 million ounces. In addition, some 25 million pounds of cobalt, 3 million pounds of nickel, and 1.5 million pounds of copper have been extracted. The nature of the veins at a typical deposit in the Cobalt area is shown on Figures V-33 and -34. The host rocks are the steeply dipping Keewatin basic and intermediate lavas (greenstones) and interflow sediments, the unconformably overlying flat to gently dipping Coleman Formation of the Cobalt Group that consists of conglomerate, greywacke, and quartzite, and the Nipissing diabase sill, which cuts across all these rocks at a low angle.

Most of the high-grade deposits are in the Cobalt Group and the Nipissing diabase. The deposits are in short veins, although some have been traced for a thousand feet or more along their strike and dip. A typical orebody consists of one or two anastamosing veins, sometimes more, varying from a few tenths of an inch to a foot or more in width. Some deposits occur in extensive faults or in fractures parallel with or related to the faults. Examples of this feature were the orebodies in the Cobalt Lake fault and those in subsidiary and parallel fractures associated with this structure. The regional faults, in general, are not the controlling features in localizing orebodies (Thomson, 1957). The fissures and fractures are largely parallel with the contact between flows and sediments in the Keewatin, or are related to basins and arches in the Nipissing diabase, the upper and the lower contacts of the intrusion, and the basal contact of the Cobalt Group, particularly where the sediments fill small basins in the underlying Keewatin greenstones.

The veins have a complex mineral assemblage. The principal gangue is white or pink dolomite, with calcite, quartz, and hematite. The metallic minerals include native

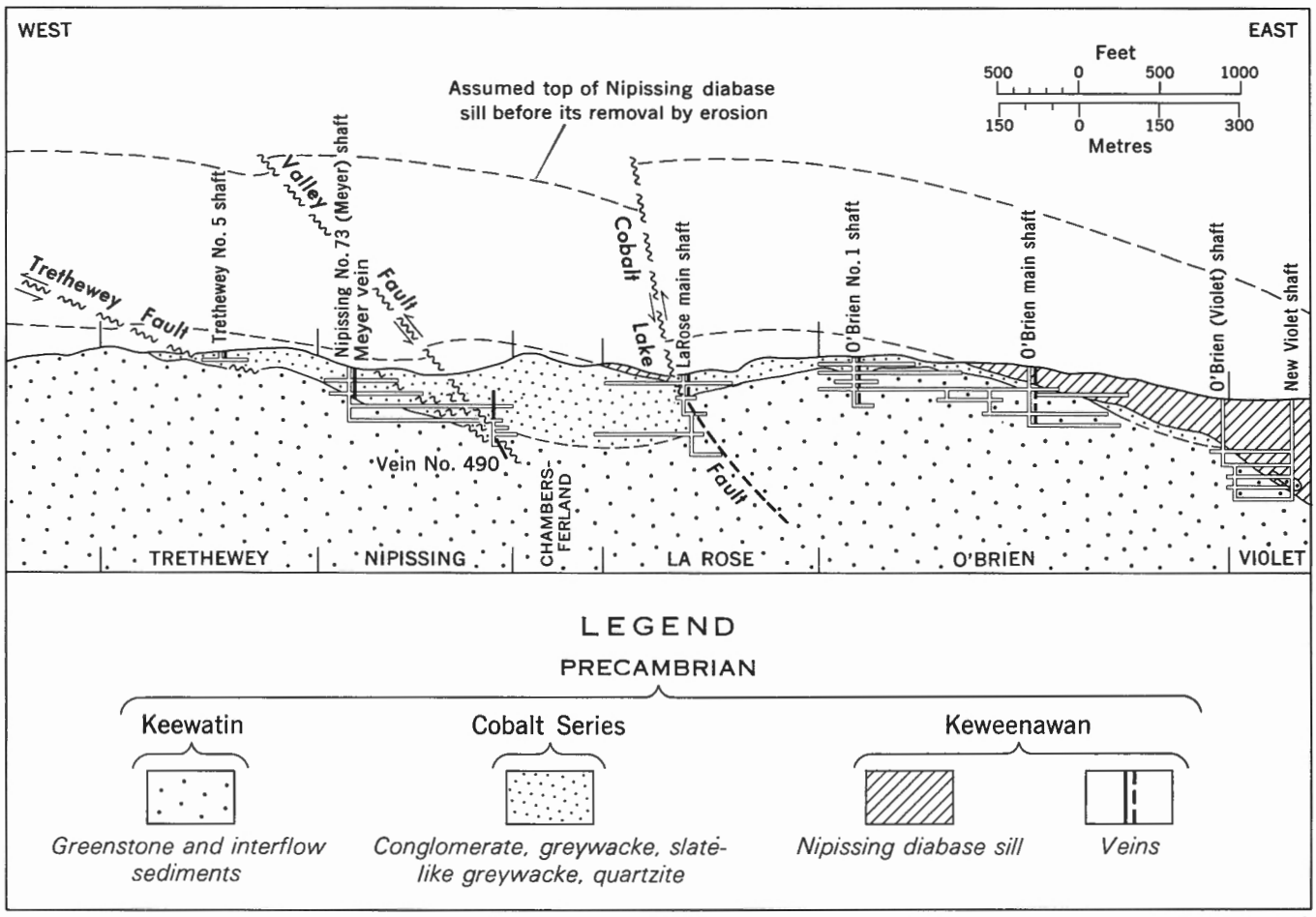


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FIGURE V-33. Section showing generalized geological features of Cobalt area, Ontario (after Knight, Miller, Thompson, and others).

silver, dyscrasite, argentite, smaltite–chloanthite, loellingite, safflorite, rammelsbergite, skutterudite, arsenopyrite, gersdorffite, cobaltite, glaucodot, niccolite, breithauptite, chalcopyrite, galena, pyrite, pyrrhotite, sphalerite, tetrahedrite, native bismuth, and minor amounts of pentlandite, polybasite, stephanite, ruby silver, and stromeyerite. Native silver and the Co–Ni arsenides are the most abundant and characteristic minerals. Secondary minerals include erythrite, annabergite, scorodite, manganese oxides (wad), and limonite. The native silver occurs as specks, spangles, and masses, some of which are huge slabs weighing as much as 1,600 pounds. Both the native silver and dyscrasite contain mercury. The silver may be intimately intergrown with the Ni–Co arsenides, dolomite, calcite, and other minerals or it may be alone, often without any gangue minerals, in tiny fractures in the wall-rocks. The Ni–Co arsenides, calcite, and other minerals are intimately intergrown, and there has been much controversy on their paragenesis and also on their relationship to the native silver. Many of the arsenides exhibit a 'concretionary' or 'tubercular' texture that suggested to Lindgren and others crystallization from a gel. There are probably no silver veins without cobalt and nickel arsenides, but there are veins with abundant cobalt minerals that have little or no silver.

Many veins in the diabase are marked by narrow alteration zones as much as an inch wide that have good bilateral symmetry and appear as a narrow dark band immediately adjacent to the vein followed outward by a light grey, bleached band. The light band consists of severely sericitized feldspar, epidote, clinozoisite, magnesium-rich carbonate, leucoxene, biotite, pyrite, apatite, and quartz, all alteration products of the diabase, which when fresh contains sodic labradorite, augite, some secondary hornblende and biotite, titaniferous magnetite, pyrite, biotite, apatite, and interstitial quartz. The dark band is gradational with the light one and consists of pennine chlorite, minor epidote, oligoclase, carbonate, leucoxene, quartz, pyrite, and apatite. The ophitic texture of the diabase, retained in the light band, is destroyed in the dark one. Some of the veins in the Cobalt sediments are also marked by a faint zone of wall-rock alteration (Boyle, 1968). The zone is rarely more than an inch or so wide and is essentially a grey bleached area that grades imperceptibly into the green sedimentary rocks. In places there is evidence of a darker chloritic zone adjacent to the veins. The bleached zone consists of a matted aggregate of clastic grains of quartz, highly altered zoisitized feldspar, abundant carbonate, and much pennine chlorite. The darker zone is due to greater amounts of chlorite. The fresh sediment contains clastic grains of quartz and feldspar in a matrix of chloritic material and minor carbonate. The texture in the bleached zones is not much different from that in the relatively fresh sediment.

South Lorrain, Ontario. These deposits are essentially mineralogically similar to those at Cobalt. Some veins occur in the Nipissing diabase near the upper contact of the sill; others occur in the Keewatin rocks above the sill. One of the veins in the Keeley mine (Wood's vein) was deeply oxidized to a depth of 600 feet or more. Much of the silver in this oxidized vein was apparently of secondary origin.

Gowganda, Ontario. The principal producing mine in this area is the Siscoe. The deposits are similar to those at Cobalt and South Lorrain and their geological setting is much the same. The same suite of Ni–Co

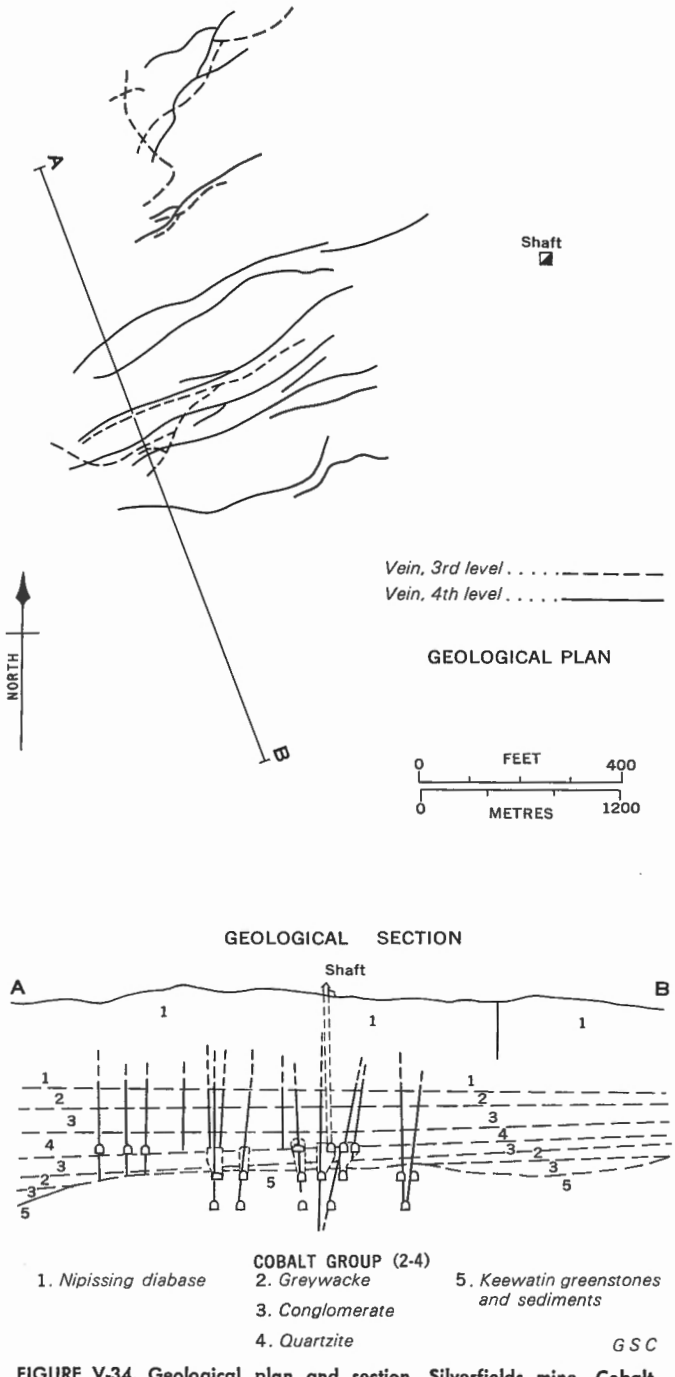


FIGURE V-34. Geological plan and section, Silverfields mine, Cobalt, Ontario (by R. W. Boyle).

arsenides is present as well as the other common metallic minerals. The principal economic mineral is native silver. There is some bismuthinite, and hematite is present in a few of the veins. The principal gangue is calcite, but some veins contain more quartz than calcite. Epidote and axinite occur and are generally earlier than the metallic minerals. Most of the veins in the Gowganda area are fissure-fillings, and are localized mainly in a Nipissing diabase sill, 300 to 400 feet from its upper contact. A few productive veins occur in the Keewatin greenstones and Cobalt conglomerates.

Origin of deposits. The origin of the Cobalt deposits and those in nearby areas has been debated for many years and still remains an enigma. Summaries of the various early theories may be found in Knight (1924) and Moore (1934). Most early investigators considered the diabase to be the source of the silver, the veins originating during differentiation of the sills. Miller (1913) studied the Cobalt deposits for many years, at times favouring the theory that the metals were leached out of the Keewatin greenstones and other associated rocks, but at other times he thought that the diabase and the ores came from one and the same magma.

Whitman (1920, 1928) took a different view of the origin of the veins at Cobalt from that of most investigators. He substantiated his theory at least in part by laboratory experiments. He concluded that:

1. The cobalt–nickel–silver ores of the Cobalt district were derived by diffusion and metasomatic fixation from the sill of Nipissing diabase with which they are associated.

2. They were deposited essentially in their present form below the groundwater level, not by circulating solutions, but by migrant ions traversing virtually stagnant or slowly circulating aqueous solutions that coursed all parts of the diabase. Deposition took place in such favourable seats of deposition as were accessible in the neighbourhood of the margins of the diabase.

3. Mineral-laden water filled all the pores, joints, and fractures of the diabase and its neighbouring rocks and through it moved the ions of vein material, at a temperature considerably below the melting point of the diabase, and at a pressure of several thousand feet of hydrostatic head.

Boyle (1968) enumerated a number of features that appear to bear directly on the problem, as follows:

1. The deposits, in addition to silver, are greatly enriched in cobalt, nickel, bismuth, arsenic, and sulphur, and a source for all these elements must be found before any adequate theory can be formulated.

2. The deposits are structurally controlled in fractures or faults that cut all the rocks in the area except perhaps the late diabase dykes. The Nipissing diabase sills were completely crystallized before the veins were formed, and accordingly the sills could not produce the constituents in the veins by the classical differentiation mechanism.

3. Diabase is lacking in many areas where native silver deposits identical in mineralization to those at Cobalt are present, as for example Jáchymov, Czechoslovakia, and Kongsberg, Norway. Diabase therefore, may not be the key to the source of all the elements in the veins.

4. Many veins at Cobalt cut deeply into the Keewatin rocks, occurring in both greenstones and interbedded sediments. Some veins are present in the Cobalt conglomerate and Nipissing diabase where they overlie the Keewatin sedimentary layers.

5. Many of the Keewatin greenstones, and the graphitic sediments especially, contain abundant sulphide minerals, mainly pyrite and pyrrhotite. The graphitic sediments also contain small amounts of galena, sphalerite, and chalcopyrite. These sulphides could be interpreted as having been introduced, but their presence in similar sediments in many parts of the Canadian Shield suggests that they were originally sedimentary. They have been recrystallized and probably were partly mobilized during regional metamorphism.

Trace element contents of the rocks at Cobalt are shown in Table V-10. The black pyritic sediments interbedded with the Keewatin greenstones have the greatest amounts of all the elements analyzed. This feature is not unique for the Cobalt district as is shown by the data from the Yellowknife area, District of Mackenzie. The greenstones, likewise, are relatively rich in most of the elements found in the veins at Cobalt. The Nipissing diabase, in areas beyond the effects of mineralization at the Cobalt camp, is remarkably low in the vein elements, except for nickel and cobalt. The Cobalt conglomerates and greywackes are, likewise, low in most of the vein elements. It would appear, then, that the only readily available sources of arsenic, antimony, and bismuth are the Keewatin sediments and greenstones. Because the Cobalt veins are so restricted in their lateral and vertical extent, their metallic constituents were probably derived from sources nearby. As the fractures dilated they drew in the more mobile constituents of the Keewatin greenstones and sediments, namely, Ni, Co, As, Ag, Bi, S, $MgCO_3$, and $CaCO_3$. The Nipissing diabase may have contributed some Ni and Co, and the heat associated with the intrusion of the sill may have had a stimulating effect in mobilizing the elements. The mode of migration of the vein constituents was probably by diffusion through a static water medium as suggested by Whitman. The Cobalt veins are considered to be essentially secretion veins, and their constituents to be late distillates from the country rocks.

Thunder Bay, Ontario. The silver deposits in the Thunder Bay district of western Ontario lie in two belts, the "grey argillite" or "Island belt", and the "black slate" or "Mainland belt" (Oja, 1967). The Island belt includes the islands of diabase in western Lake Superior. The Mainland belt is roughly 4 miles wide and extends northeast and southwest of the city of Port Arthur. The older rocks are steeply dipping metamorphosed Archean volcanic

rocks, gneisses, and schists intruded or granitized to granite, granite gneiss, and migmatite during the Kenoran Orogeny. Unconformably overlying these rocks are the Aphebian Gunflint and Rove Formations of the Animikie Group, which are in turn overlain, apparently unconformably, by the Helikian Sibley sediments. All these formations are intruded by Keweenawan diabase sills and dykes and severely faulted, chiefly northeast and northwest.

The principal deposit in the Island belt was the *Silver Islet* mine which produced, from 1868 to 1884, more than $3 million in silver. It was the first silver mine to be worked in Canada. The problems with which the miners had to deal on such a small islet of rock some 80 feet in diameter, exposed to the fury of Lake Superior, are legendary.

The steeply dipping vein system was mainly in a diabase dyke, 250 feet wide, that cuts through relatively flat lying Animikie argillites, greywacke, and slates. The main vein averaged 8 feet wide and produced to a depth of some 1,300 feet, below which it became too low grade to be worked economically at that time. The gangue was mainly calcite and manganiferous dolomite with a little quartz. Rhodochrosite and barite occurred in places. The metallic minerals included abundant native silver, argentite, galena, sphalerite, chalcopyrite, pyrite, marcasite, tetrahedrite, domeykite, niccolite, cobaltite, smaltite, breithaup-

TABLE V-10 *Analyses of rocks from Cobalt and South Lorrain, Ontario (Boyle, 1968) (all values in parts per million except where noted)*

Description of Samples	Cu	Pb	Zn	Ni	Co	Bi	Ag	As	Sb	S	Au
Composite sample of Keewatin greenstone, South Lorrain area	330	<5	95	130	67	NF	0.30	15	2.5	1,700	ND
Composite sample of Keewatin greenstone, Cobalt area	120	<5	215	105	69	NF	0.19	2	0.5	775	ND
Composite sample of Cobalt conglomerate and greywacke, Cobalt area	20	<5	20	60	25	NF	<0.05	<2	<1	180	ND
Nipissing diabase: average values computed from 36 samples from drill hole through diabase, Henwood twp., Cobalt area	83	<2	70	128	40	<0.05	0.11	<2	<1	590	ND
Composite sample of Keewatin black pyritiferous slate, schist, and greywacke, Cobalt area	420	75	2,400	160	74	0.83	0.83	140	12.5	~6%	ND
Sulphides (mainly pyrite with some pyrrhotite, chalcopyrite, etc.) separated from above sample	750	200	4,400	570	217	NF	5.7	1,150	46	ND	0.128
Sulphides (mainly pyrite and pyrrhotite) separated from Keewatin greenstone, South Lorrain area	ND	ND	ND	1,500	3,000	NF	7	ND	ND	ND	ND
Composite sample of black pyritiferous slate, schist, greywacke, tuff, etc., Yellowknife area (range in ppm)	<5–400	<5–20	10–1,600	10–110	<5–110	ND	0.33–1.6	<5–15	<1–1.0	~5%	0.01–0.07

Gold analyses by Mines Branch, Ottawa. Other analyses by Geochemical laboratories, Geological Survey of Canada.
NF = not found; ND = not determined.

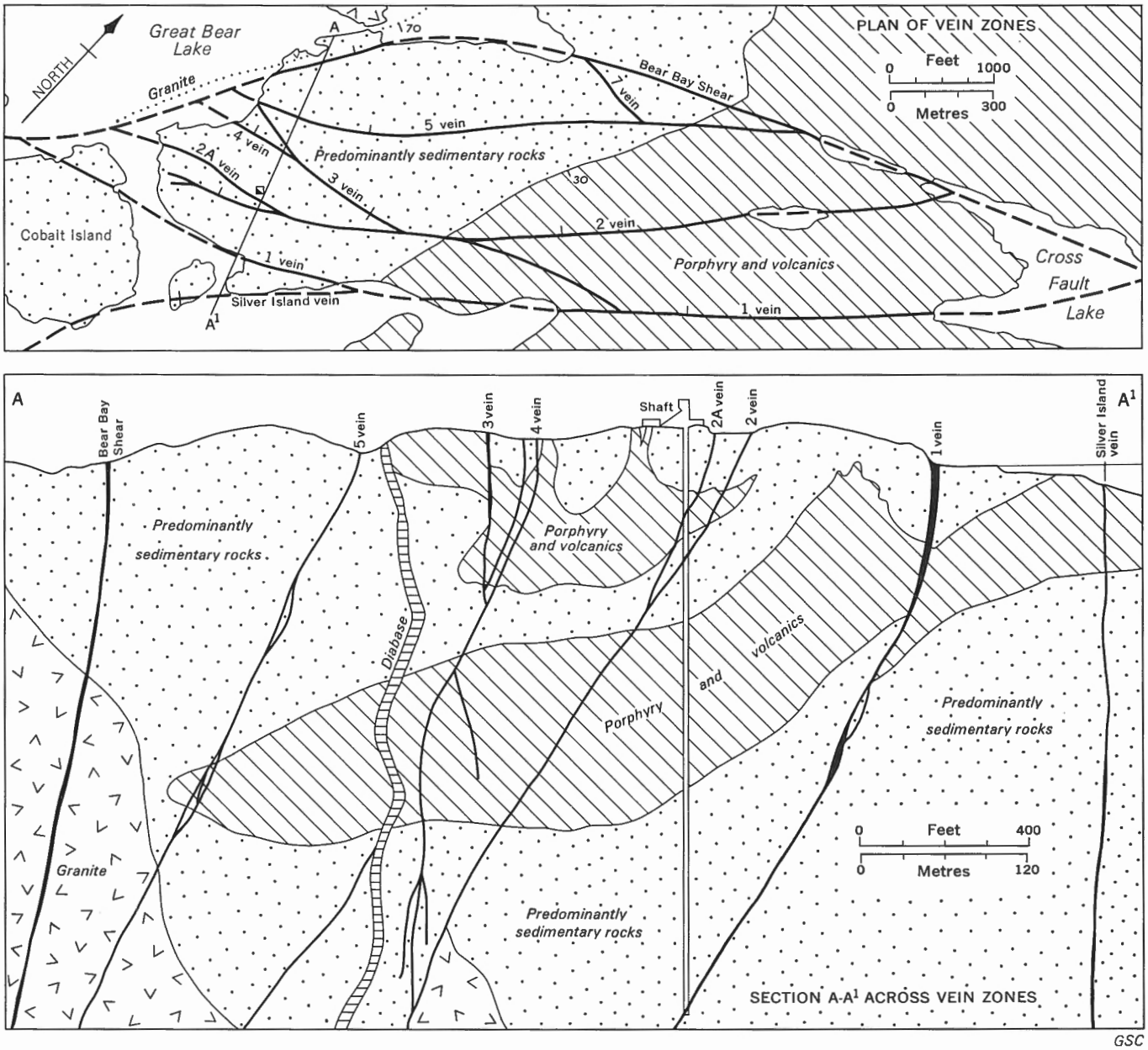


FIGURE V-35. Generalized plan and section of geology in the vicinity of Port Radium mine, Port Radium, Northwest Territories (Campbell, 1957).

tite(?), millerite(?), and other poorly differentiated arsenic, antimony, cobalt, nickel, and silver minerals. Graphite (or carbonaceous matter) occurred in quantity in parts of the vein. Methane and hydrogen sulphide were encountered in relatively large quantities in some parts of the vein, and flows of salt water were common. The principal secondary minerals were quartz, calcite, barite, marcasite, nickel bloom (annabergite), cobalt bloom (erythrite), pyrolusite, native silver, and chlorargyrite. The shoots occurred as fabulously rich bonanzas along the vein system and were separated by relatively lean stretches.

The diabase dyke, which the vein cuts nearly at right angles, acted only as a structural site for the mineralization. It probably played little part in the mineralization of the vein, although it may have supplied some of the gangue elements and perhaps some cobalt and nickel. The sediments seem to be the most likely source of the metals, arsenic, sulphur, and the abundant graphite.

The prospects and mines in the Mainland belt include the Beaver, Silver Mountain, Badger and Porcupine, Rabbit Mountain, Shuniah, Jarvis, Thunder Bay, 3A and Beck, and several others. Each of the first three have produced about half a million dollars in silver. The country rocks are mainly flat lying, black, pyritic slates and argillites, greywacke, and phyllite of the Rove Formation that

have been intruded by Keweenawan diabase sills and broken by several normal faults giving rise to numerous buttes, mesas, and cuestas formed by the sills. The deposits are veins, lodes, and irregular stockworks in the steeply dipping faults and subsidiary fractures, which strike northeast or north to northwest. Most known veins occur in the Rove sediments, but some extend into the diabase sills and at least one, the Shuniah vein, cuts the basement complex. The gangue minerals are mainly calcite, white, smoky, and amethystine quartz, fluorite, barite, and some witherite. Metallic minerals include native silver, argentite, sphalerite, galena, pyrite, pyrrhotite, and chalcopyrite. Oja (1967) noted that most geophysical methods failed to locate the veins in the Mainland belt. Geochemical methods using zinc and silver as indicators in the humus and soil were found to be useful in tracing the veins.

Great Bear Lake area, District of Mackenzie. A number of silver deposits and occurrences are known in a belt that stretches along the eastern shore of Great Bear Lake southward to Hottah Lake. This belt includes the now exhausted Eldorado uranium mine at Port Radium. Additional description of the deposit is given in the section on Uranium.

The *Eldorado* (3) mine obtained its ore from veins and stockworks in faults, shear zones, and fractures that cut all the consolidated rocks of the area (Fig. V-35). Most productive deposits, however, occurred in the Echo Bay Group, generally in the lower part, but some were in the granodiorite and other granitic rocks. The principal gangue was quartz and hematite, but there was some carbonate, usually calcite. In some veins the principal economic mineral was pitchblende; in others native silver. Other metallic minerals included a great variety of Co–Ni arsenides and sulphides, chalcopyrite, argentite, galena, molybdenite, pyrite, and native bismuth. Secondary minerals included yellow and orange uranium ochres, azurite, malachite, erythrite, annabergite, and various manganese oxides. Campbell (1957) recognized five stages of mineralization in the Port Radium mine as follows: (1) hematite, quartz; (2) pitchblende, quartz, hematite; (3) quartz, Co–Ni arsenides and sulphides; (4) copper sulphides, chlorite, dolomite; (5) carbonate, silver, bismuth, chalcopyrite. This sequence is somewhat similar to that at Jáchymov, Czechoslovakia, although there are differences. The sequence of quartz and hematite followed by pitchblende and the Ni–Co arsenides, in turn with silver and bismuth as a late phase, is typical of deposits of this type. The principal alterations in the Port Radium veins were argillic, chloritization, hematitization, and carbonatization. There was also some silicification, sulphidization, and the development of apatite in some veins.

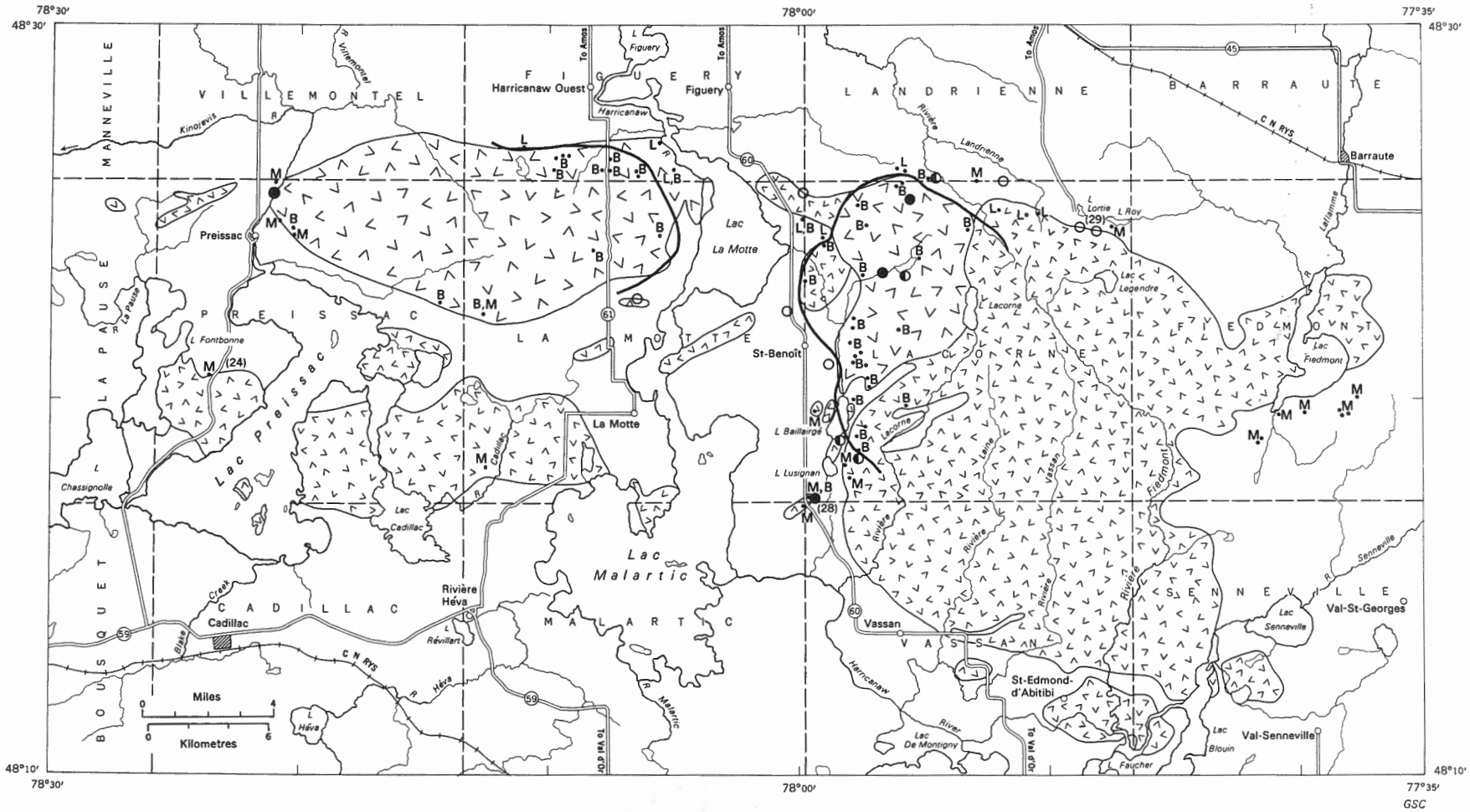
The *Echo Bay* (2) mine is producing silver about a mile northeast of Port Radium. The deposits are in irregular veins in fractured and shear zones in tuff, dacite, feldspar porphyry, and fine-grained banded sediments of the Echo Bay Group (Thorpe, 1966). Four principal northeast-striking veins, ranging from a few hundred feet to more than 1,000 feet in length, have been explored. Most dip steeply north and vary in width from a few inches to five feet. The principal gangue is manganiferous carbonate and quartz. Metallic minerals include the Ni–Co arsenides and sulphides, pyrite, argentiferous galena, sphalerite, chalcopyrite, native silver, acanthite (argentite), and stromeyerite. Several ages of mineralization are evident. In 1965, production was 1,408,246 ounces of silver from 35,609 tons of ore. Some copper and lead are also extracted.

A number of other silver veins in the Great Bear Lake area have received some attention in the past, but none has produced any quantity of ore. The *Camsell River* silver deposits on Camsell River (Lord, 1951) are veins and stockworks in fracture and shear zones in andesitic tuffs and agglomerates, diabase, and quartz–feldspar porphyry. The economic mineral is native silver in a gangue of quartz and carbonate. Ni–Co arsenides and sulphides, native bismuth, and bismuthinite are present in a number of veins, with some pyrite, chalcopyrite, sphalerite, and galena. Hematite and minor amounts of pitchblende are reported. Veins on the north shore of *Contact Lake* are in fractured and sheared zones in granodiorite, probably of Aphebian age (Furnival, 1939; Lord, 1951). The gangue is quartz and carbonate, and the minerals of interest are native silver and pitchblende. Other metallic minerals include the Ni–Co arsenides and sulphides, arsenopyrite, the common base metal sulphides, native bismuth, bismuthinite, pearceite, stromeyerite, argentite, and hessite.

Platinum Group

The metals of the platinum group include platinum, palladium, iridium, osmium, rhodium, and ruthenium. In Canada, these metals are produced solely as byproducts from the treatment of nickel–copper ores. Most come from the mines at Sudbury, Ontario, and Thompson, Manitoba. Minor production is obtained from nickel–copper concentrates shipped to Sudbury from Gordon Lake, Ontario, and Belleterre, Quebec.

Platinum and some rhodium are present in the mineral sperrylite, a platinum diarsenide that is scattered sparsely through the sulphides. Rare palladium bismuthides have been identified in Sudbury ores. Elements of the platinum group are held in solid solution in minute amounts in various copper–nickel arsenides, which are generally more abundant in the copper-rich sections of the ore zones. The average content of these metals in Sudbury ores is 0.025 ounce per ton, with platinum making up about half the total. In 1929, the last year for which separate records are reported, the yield was 12,474 ounces



Main areas containing muscovite granite

Other granitoid rocks of the Preissac-Lacorne batholith

Zone boundary

Location of pegmatite, aplite, or vein .....•
Mainly beryllium occurrence .....●
Main lithium occurrence .....○
Combined beryllium - lithium occurrence .....◐
Minor occurrences of beryllium, lithium .....B, L
Molybdenite .....M

PROPERTIES DESCRIBED IN TEXT

(24) Preissac Molybdenite
(28) Molybdenite Corporation (Lacorne)
(29) Quebec Lithium

FIGURE V-36. Regional zoning of lithium and beryllium occurrences, Preissac–Lacorne area, Quebec (by R. Mulligan).

platinum, 12,231 ounces palladium, 3,037 ounces rhodium, 1,376 ounces ruthenium, 497 ounces iridium, and no osmium. In 1965 a method was developed for recovering osmium. Since 1920, production essentially has been equally divided between platinum and the remainder of the group; for 1966 it was 396,059 troy ounces. Cumulative Canadian production was 11,187,500 troy ounces to the end of 1966.

MINOR METALS

Arsenic

Arsenopyrite is commonly associated with silver–cobalt ores and Canada's total production of refined white arsenic originates from this source. In 1966, 350 tons of arsenic oxide were produced as a byproduct from silver–cobalt ores at Cobalt, Ontario. In western Quebec, at the O'Brien mine and the Beattie–Duquesne mine, prior to 1966, crude white arsenic was recovered from the treatment of gold ores.

Beryllium

Beryllium occurrences are widespread in Slave, Superior, and Grenville Provinces of the Canadian Shield, and are also found in Churchill and Nain Provinces. Most occurrences are in granitic pegmatite dykes, and the beryllium is in the form of beryl or, rarely, chrysoberyl and other minerals. Most of the richer berylliferous dykes of Slave and Superior Provinces are associated with lithium minerals, as parts of either a regional or an internal zoned sequence. The most important of these areas are the *Preissac–Lacorne* district, Quebec (Fig. V-36), the *Cat Lake–Winnipeg River* district, Manitoba, and the *Yellowknife–Beaulieu River* district, District of Mackenzie. However, beryllium occurs much more extensively than lithium, although it has not been found in comparable concentrations. Some berylliferous pegmatites are associated with molybdenite, as at *Lacorne* (28), Quebec, where a small amount of byproduct beryl represents the only current production in Canada. Beryl is also found in molybdenite ore at the old Height of Land mine nearby.

The few non-pegmatitic beryllium occurrences of the Shield include the unique alkali-metasomatic deposit at *Seal Lake,* Labrador. Fine-grained barylite and eudidymite occur in zones of soda-rich paragneiss associated with syenitic intrusions. Some zones average 0.35 to 0.40 per cent BeO. Niobium is also present, together with minor zinc and rare-earth minerals. Most other beryllium occurrences are related to granitic intrusions. The berylliferous pegmatites, except for the complex lithium–beryllium ones, are as a rule unzoned or crudely zoned, although the beryl is commonly concentrated in or near quartz-rich central areas. Beryl is commonly associated with muscovite and the platy albite cleavelandite. In regionally zoned districts the berylliferous pegmatites are generally closer to the related granitic bodies than are the simple spodumene-bearing dykes. In the complex lithium–beryllium pegmatites the beryl is chiefly in the outer intermediate or wall zones, although some late alkali beryl is also present in the innermost lithium-rich and other zones.

No deposits are being mined primarily for beryllium at present, but beryl is recovered from the waste-sorting belt at Lacorne molybdenite mine. Some years ago, one or more shipments of beryl, totalling perhaps 50 tons with an average beryl content of 0.25 per cent, were made from the dykes in Lyndoch township, near Qaudeville, Ontario. Small amounts of beryl have been taken from several pegmatites in the Winnipeg River district, Manitoba. Some beryl was recovered from the Moose and Best Bet, and probably also the Peg Tantalum properties in the Yellowknife–Beaulieu district, Northwest Territories.

Bismuth

Bismuth, used in pharmaceuticals and chemical compounds, is recovered mainly as a byproduct from the *Preissac* (24b) and *Lacorne* (28) molybdenum deposits in Quebec. It occurs as the native metal or as the sulphide, bismuthinite. Small amounts of bismuth are also recovered from silver–cobalt ores. The deposits occur in quartz veins and in intrusive rocks of granitic or intermediate composition that contain varying amounts of feldspar, mica, bismuth, bismuthinite, molybdenite, and iron sulphides.

Cadmium

In 1966, a quarter of the Canadian cadmium production came from the Canadian Shield. The element occurs as a minor constituent of sphalerite, commonly ranging between 0.1 and 0.3 per cent but it can be as high as 0.7 per cent. Its concentration generally varies inversely as the iron content of the sphalerite. Cadmium is recovered from zinc concentrates of the Flin Flon area of Manitoba and Saskatchewan, the Manitouwadge area of Ontario, and the Matagami and Noranda areas of Quebec. The main uses of cadmium are in the plating of steel and the manufacture of pigment.

Caesium

Caesium deposits are rare, and the pollucite deposit at *Chemalloy Minerals,* Winnipeg River district, Manitoba, is the largest and richest known (Fig. V-37). It contains

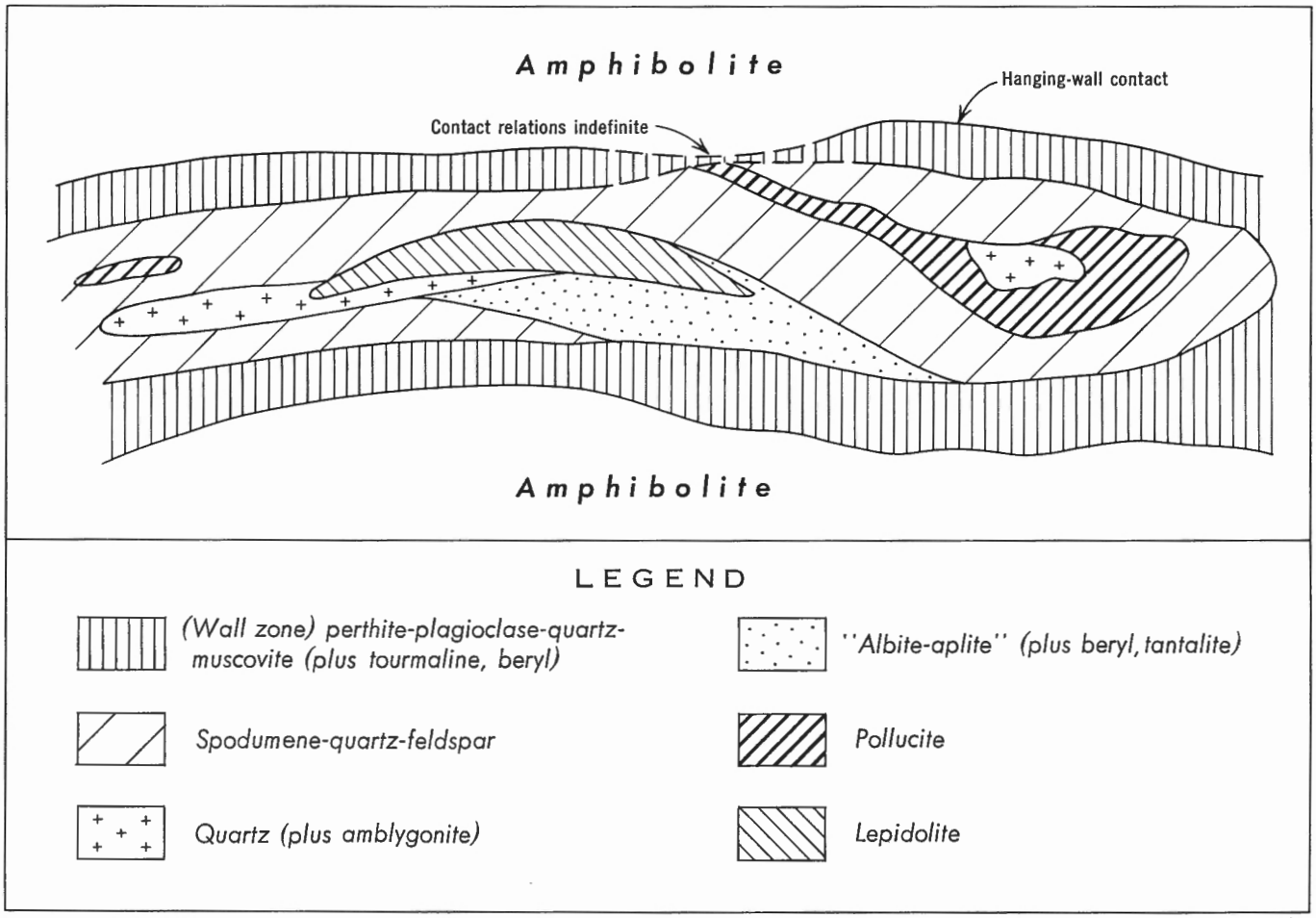


FIGURE V-37. Diagrammatic composite vertical section of Chemalloy Minerals Limited pegmatite (by R. Mulligan).

about 300,000 recoverable tons averaging more than 20 per cent Cs_2O. The pollucite occurs in the inner units of a highly complex pegmatite, and is associated with spodumene and lepidolite. Some pollucite is present in a complex pegmatite in the Lacorne area and a little has been reported from spodumene-bearing pegmatites elsewhere.

Lithium

Lithium occurrences are widespread in Archean metasedimentary and metavolcanic belts of Superior and Slave Provinces. The chief districts are near Lacorne, Quebec, Nipigon, Ontario, Winnipeg River, Manitoba, and Yellowknife, District of Mackenzie. Among other major fields are the East Braintree–West Hawk Lake district of Manitoba, the Lac la Croix, Root Lake, Dryden, and Zig–Zag Lake areas of Ontario, and the Lac Expanse and Assinica Lake areas of Quebec. Except for the Herb Lake deposit in Churchill Province near the boundary with Superior Province, lithium occurrences are rare in the younger Precambrian provinces. All the lithium deposits are in moderate to highly metamorphosed rocks near their contacts with granitic intrusions. The main lithium-bearing mineral is spodumene, although amblygonite and lepidolite are important constituents of some complex pegmatites in the Winnipeg River and Yellowknife districts. Petalite is rare; a little occurs in some pegmatites of the Winnipeg River area. Many deposits contain several million tons of ore grading 1.2 per cent or more Li_2O.

The lithium occurrences are mainly in unzoned pegmatite dykes. These contain spodumene crystals a few inches long oriented perpendicular to the walls and uniformly distributed between them. Other occurrences are well-zoned bodies that have tourmalinized wall zones, beryl-bearing feldspar–quartz–muscovite intermediate zones, and spodumene-rich inner zones. The latter locally contain quartz–amblygonite "cores" and lepidolite assemblages. In these bodies the spodumene is commonly coarser grained and random in orientation. Most zoned pegmatites are gently dipping sheets, whereas most unzoned ones are steeply dipping. A regional zoning is apparent in several districts, notably Lacorne (Fig. V-36) and Yellowknife. In these districts simple unzoned spodumene pegmatites lie farther from the central parts of the associated granitic plutons than the lithium-free berylliferous pegmatites and the complexly zoned lithium–beryllium pegmatites.

The *Quebec Lithium* (29) mine in the Preissac–Lacorne district, Quebec, is in an unzoned spodumene pegmatite. A belt 8,000 feet long and 2,000 feet wide contains numerous subparallel dykes as much as 2,000 feet long and 100 feet thick with a moderate to steep dip. The dykes pass from amphibolitized greenstone at the surface into granodiorite at depth, and are remarkably homogeneous in grade and texture. Reserves in the shaft area alone have been estimated at 15 million tons of ore averaging more than 1.2 per cent Li_2O. Mining and milling, at about 1,000 tons per day, began in 1955. In 1960, a chemicals plant was erected at the property to produce lithium carbonate and hydroxide. In recent years it has produced more than a million tons per year of 'contained lithia' valued at over $1.1 million.

In the Winnipeg River district of southeastern Manitoba, several pegmatites are of the complexly zoned type. The *Chemalloy Minerals* mine is developed in a virtually flat lying, domical lenticular sheet some 2,500 feet long, 1,500 feet wide, and 250 feet thick in the central explored area (Fig. V-37). An inner intermediate assemblage of spodumene–quartz–feldspar is overlain and underlain by a wall-zone of perthite–plagioclase–quartz–muscovite with tourmaline and minor beryl. The spodumene zone, in turn, virtually encloses core-pods of quartz with minor amblygonite, lenses rich in lepidolite and the caesium mineral pollucite, and an aplitic albite unit that carries late beryl and wodginite, a rare tin–tantalum mineral. A modern surface plant was completed in 1960 and extensive development work was done but only minor shipments, mostly for testing purposes, were made before the property was shut down. Reserves have been estimated at about 5 million tons averaging 2 per cent Li_2O. This reserve is mostly the spodumene ore but includes some lepidolite and the pollucite ore.

Magnesium, Magnesia

Defence requirements and the transportation industry have created a very great demand for magnesium metal, particularly for alloying with aluminum in aircraft construction and for smelting of other metals. The deposit at Haley, Ontario, near Renfrew, which came into production in 1942, is Canada's only magnesium and metallic calcium producer. Open-pit mining is carried out in a dolomite of exceptional purity, part of the Grenville Group, bounded on one side by quartzite and on the other by paragneiss. The dolomite is calcined to magnesium by the Pidgeon ferrosilicon process and then refined. Production has been fairly steady at 6,000 to 9,000 tons per year since 1955 and in 1966 was 6,786 tons.

Magnesia, derived from the minerals magnesite and brucite, is used mainly in the manufacture of basic refractories. Canadian production comes only from the Shield, and currently only from within the Grenville sediments. Dolomitic magnesite is mined at *Kilmar* (48), Quebec. At *Wakefield* (44), Quebec, a brucitic limestone was quarried. Its composition suggests that it was formed through thermal metamorphism of a pure dolomite. Some of the material mined was shipped to Arvida, Quebec, for conversion to magnesium chloride and thence to magnesium metal, but most was processed at Wakefield, and magnesia recovered. Magnesium hydroxide, used in the manufacture of pulp and paper by the Magnefite process was also produced at Wakefield. Magnesite and brucite production averaged $3,000,000 to $3,500,000 per year since 1955, and in 1966 was $3,948,599.

Molybdenum

Molybdenum occurs in molybdenite in pegmatites, quartz veins, disseminations in granite and syenite, skarns, and sulphide deposits, especially of the porphyry copper type. It is used mainly in the manufacture of steel. Although minor deposits are known in many places in the Canadian Shield, most production has been from deposits in the Preissac–Lacorne district, Quebec, in Superior Province. The deposits are in pegmatites and quartz veins in and near granitic bodies related to the Preissac–Lacorne batholith. They represent the outermost zone of a regionally zoned sequence of pegmatites also bearing lithium and beryllium (Fig. V-36). At the *Lacorne* (28) mine, molybdenite-bearing pegmatitic and quartz veins of slightly different ages form a complex network in mica schist cut by numerous granite dykes, near a stock or sill of granodiorite. The grade of ore mined has varied from 0.5 to 0.25 per cent MoS_2. Bismuth is an important byproduct, and the pegmatitic veins contain a little beryl. The Lacorne mine was the only important Canadian producer from 1944 until about 1963. On Indian Peninsula the deposits are similar pegmatitic and quartz veins in and related chiefly to muscovitic granite phases of a southern tongue of the batholith. They produced intermittently during World War II. The *Preissac* (24b) molybdenite property commenced production in 1965. It includes the Old Indian molybdenite mine, where early production was from quartz veins contained between two faults in the granite.

In Grenville Province molybdenum occurrences are numerous and of various types but significant production, mostly during World War I, was confined to a few deposits in Renfrew county, Ontario, and near Quyon and Maniwaki in Quebec. The deposits are in a belt containing crystalline limestone cut by syenitic and other intrusions. The most characteristic types are pyroxenite skarns with pyrite, pyrrhotite, and molybdenite. The Hunt and Spain mines in Renfrew county, Ontario, and deposits near Maniwaki, Quebec, were dominantly of this type. At the Moss mine, near Quyon, the molybdenite occurred chiefly as disseminations in a syenite complex, associated with feldspar, quartz, fluorite, magnetite, and sulphides. The Moss mine was at one time, during World War I, the world's leading molybdenum producer. It was reopened during World War II and produced about 150

tons per day of ore, grading about 0.2 per cent MoS_2, until 1944.

Selenium and Tellurium

Selenium and tellurium are recovered together as by-products in the electrolytic refining of copper from the nickel–copper ores produced at Sudbury, Ontario, and Thompson, Manitoba, and from the copper ores of the Noranda, Quebec, and Flin Flon, Manitoba areas. All Canadian selenium and tellurium came from the Shield until Gaspé Copper Mines Limited began production in 1955.

Tantalum and Niobium

Pegmatitic tantalite–columbite deposits were formerly the main source of both niobium (columbium) and tantalum but alkaline complexes now yield most of the niobium. Columbite–tantalite is a minor constituent of most lithium and beryllium pegmatites in the Canadian Shield but occurs in significant amount in only a few dykes, chiefly those of the main lithium–beryllium districts. The only recorded production was from the Yellowknife–Beaulieu district. Near *Ross Lake* beryl-liferous feldspar–quartz–muscovite pegmatites yielded 3,750 pounds of tantalite–columbite concentrates from 940 tons of rock quarried in 1946 and 1947. Near Hearne channel zoned lithium–beryllium pegmatites yielded 2,600 pounds of tantalite–columbite concentrate in 1946 and 1947, and 39,100 pounds from 18,928 tons milled in 1953 and 1954. In the Winnipeg River district, the zoned *Chemalloy Minerals* pegmatite at Bernic Lake contains columbite–tantalite and the rare tin tantalite, wodginite, in an interior albitite zone.

The largest reserves of niobium are associated with alkaline igneous complexes. These are more or less circular bodies composed of various rocks including sodium-rich and nepheline-rich syenites, ijolites, jacupirangites, carbonate, and metasomatized carbonate rocks. The deposits are in the complexes and the fenitized wall-rocks. Niobium-bearing minerals include pyrochlore, uranium pyrochlore, betafite, columbite, ilmenorutile, perovskite, and niocalite. Accessory minerals include apatite, magnetite, biotite, soda pyroxene, soda hornblende, pyrite and pyrrhotite. The largest deposits are in a Cretaceous, Monteregian Intrusion at Oka, Quebec. Others occur in central Superior Province extending north of Lake Superior through Kapuskasing. At *Oka* (49), Quebec, production of niobium concentrates began in 1961. Shipments of pyrochlore concentrates in 1965 amounted to 2.3 million pounds of contained columbium pentoxide, worth $2,350,000. Reserves are estimated at 62.7 million tons averaging 0.4 per cent Cb_2O_5.

Tin

No economic tin deposits have been found in the Canadian Shield. In the Yellowknife–Beaulieu district a few pegmatites near Sproule Lake and Hearne Channel carry appreciable concentrations. The tantalum–columbium concentrates from the Best Bet property were reported to contain 13.4 per cent tin. In the Winnipeg River district, Manitoba, attempts were made to mine pegmatites near Bernic and Shatford Lakes for tin. At the *Chemalloy Minerals* property near Bernic Lake, the albite–aplite zone of a complex pegmatite (Fig. V-37) contains the rare tin–tantalum mineral wodginite, as well as a little cassiterite. At Red Sucker Lake in northeast Manitoba an albitite dyke contains cassiterite.

Tungsten

The principal ore of tungsten in the Shield is scheelite, the calcium tungstate. It is found chiefly in gold ores and also in quartz veins in the Yellowknife–Beaulieu area of the Northwest Territories. No deposit worth mining for tungsten alone has yet been found, but to fill war requirements, the relatively small amounts occurring in gold ores were separated. At the Hollinger property in the Porcupine area, Ontario, where perhaps the largest bodies of scheelite are found, a mill was erected that treated 135 tons of ore daily. Ore from the other mines in the district was shipped to this mill. Some tungsten was obtained in the past from a deposit on Outpost Islands, Great Slave Lake. The ore is a gold–copper–tungsten–tin complex, and difficulty was encountered in producing a satisfactory tungsten concentrate. The tungsten mineral there is mainly ferberite, a tungstate of iron, and the ore is said to run about 1 per cent tungsten.

INDUSTRIAL MINERALS

Apatite

Apatite, the tri-calcium phosphate, was mined at one time in considerable quantity for the manufacture of superphosphate fertilizer. About the beginning of the century, however, most of the mines closed. The little apatite produced since then has been a byproduct of mica mining and used mainly for the manufacture of phosphorus. Production in 1941, about 2,500 tons, was the largest for many years; none has been produced since 1951. Almost all the apatite came from Cantley, Quebec, north of Ottawa, or from near Kingston, Ontario. It was found in irregular pockets in a pyroxenite formed by alteration of crystalline limestones of the Grenville Group.

Asbestos

A small amount of asbestos was produced from the Munro mine near Matheson, Ontario, from 1950 to 1964 and since then from the Hedman mine. Production from the former mine ranged from 25,000 to 30,000 tons per year, and from the latter, was 1,700 tons in 1966. The Munro deposit was in a vertically dipping, differentiated mafic sill, about 1,000 feet thick, intrusive into Keewatin volcanic rocks. Serpentinized dunite and peridotite form the southern half of the sill in the main ore zone at the surface, but are enveloped in a sheath of talc–carbonate rock that thickens with depth. Veins of asbestos were contained in the core of the ultramafic sill or in adjacent serpentinized peridotite, associated with a medium to light green, granular serpentine. Fibre was as much as an inch long, mostly rather harsh, although flexible. Crossfibre occurred in one- and two-fibre veins which formed a rectangular pattern.

Small bodies of serpentine that carry veins of asbestos in many places are common in the Grenville Group north and east of Ottawa. Opinions differ as to whether these represent intrusions of peridotite or metamorphosed crystalline limestone. The fibre is of high quality, low in iron and usually entirely free from magnetite. Attempts have been made from time to time to mine the deposits, but they are too small. Large bodies of actinolite were mined near Actinolite village, Ontario, intermittently between 1883 and 1927. The mineral appears to occur in a basalt or other greenstone that has been altered by the intrusion of granite gneiss. As dolomite or ferruginous carbonate are present in other parts of the belt it is possible however, that the rock is an altered crystalline limestone.

Elsewhere in the Shield, small deposits of chrysotile asbestos are known but none has proved profitable although several have yielded small quantities, especially those in northern Ontario from the Abitibi serpentine belt. A small amount was produced from an unusual deposit in Bannockburn township, Ontario. There, veinlets of asbestos occur in a shattered zone of rhyolite at its contact with a highly serpentinized body of peridotite. Extensive development work has been taking place at the Asbestos Hill deposits in the Ungava district of northern Quebec, estimated to have reserves of 20 million tons. Several showings of crocidolite, the blue, sodic and ferruginous asbestos, have been found near Knob Lake, Quebec, and are also being explored.

Corundum

Corundum is a constituent of the nepheline syenites of Grenville Province in southeastern Ontario. The corundum is chiefly a component of reddish alkali syenites, and of their pegmatitic facies, occurring in dykes as much as 18 feet wide. In places corundum forms as much as 10 per cent of the rock mass, but the average content of the rock milled was between 5 and 6 per cent. Mining of corundum began about 1900, and production reached a maximum in 1906. Between 1900 and 1921, 19,524 tons of the mineral were shipped. No commercial shipments have been made since 1946 when treatment of old tailings at the Craigmont property ceased. During the two years of operation about 2,600 tons of concentrate were shipped to the United States.

Feldspar

Feldspar has been replaced in the last ten years by nepheline syenite for use in the glass industry. Feldspar is now used as flux and in the manufacture of ceramics. Feldspar is mined from pegmatite dykes, and from the same general areas as apatite and mica. Canadian production has come from the Kingston and Bancroft areas of Ontario and the Gatineau and Buckingham areas of Quebec. The sole Canadian producer in 1966 was the Derry mine near Buckingham, Quebec. Hand-cobbed feldspar is produced from a large dyke. Production from the Baié Johan Beetz area and byproduct production from spodumene mining in the Val d'Or area, both in Quebec, ended in 1959. Production of feldspar, since a high of 55,000 tons in 1948, has averaged roughly 20,000 tons per year until 1959, and about 10,000 tons per year since then. Production in 1966 was 10,924 tons.

Fluorite, Barite, Celestite

These minerals occur in veins of Paleozoic age in southeastern Ontario and adjacent parts of Quebec. The veins are in Precambrian crystalline rocks and overlying flat-lying Paleozoic strata. The proportions of the principal minerals vary widely from one vein to another, and even in different parts of the same vein. Those that carry fluorite in commercial quantities are found near Madoc and Cardiff, Ontario. The veins near Madoc occur in two groups. Those of the Moira Lake group centre around a northwesterly trending fault that has been traced more than 5 miles. Most of them lie in the fault fissure, but some are in parallel subsidiary fractures. The Lee–Miller group lies a mile or two west of the Moira Lake group, and forms a string of deposits trending somewhat more northerly. Their linear arrangement suggests that they also lie along a fault. The vein materials are concentrated in lenses a few feet to 200 feet long, and 2 to 17 feet wide. The fluorite is partly massive and partly forms crystals in numerous vugs. Production from these veins has been mainly during times of war. From 1906 to 1920 nearly 20,000 tons of fluorite were mined and from 1940 to 1945, 38,671 tons were recovered. A few hundred tons of fluorite have been shipped annually in recent years from veins in the Madoc area. Production ceased in 1961 after an estimated 120,000 tons had been mined.

Garnet

Garnetiferous schists and gneisses are developed in various parts of the Shield, but very little garnet is mined as it does not meet the rather exacting specifications demanded for industrial use. In 1942, 17 tons were mined, all from the River Valley property 40 miles northwest of North Bay, and in 1944 only 3 tons were mined.

Graphite

Flakes of graphite form a common constituent of the crystalline limestones of the Grenville Group. In places these rocks have been acted on by silica-bearing solutions, presumably derived from granitic or other intrusions, and converted to a varying extent to silicate minerals such as feldspar, diopside, scapolite, and mica. Calc-silicate bodies also carry important amounts of graphite and some pyrite and quartz in places. These minerals appear to have been among the last to form, as they cut through and seemingly replace the silicates. In addition to this relationship, there appears also to have been a structural control, because the graphite bodies are much wider and richer at the crests of sharp folds than farther out along the limbs. The most important graphite deposit was the Black Donald mine, about 25 miles southwest of Renfrew, Ontario. This body was discovered about 1889, and was in operation, with interruptions, from 1895 to 1954. The deposit is a bed-like mass, 10 to 30 feet thick, that has been sharply folded into an asymmetrical syncline, plunging northeast at an angle of about 20 degrees. Available data seem to suggest that, at the northeast end of the workings, the north limb of the syncline rolls over into an anticline. The grade of the ore was unusually high. Much of the ore mined in the earlier days was 70 to 85 per cent graphite, and it has been estimated that the body as a whole averaged 55 to 65 per cent, although the last two years of salvage operation produced only about 15 per cent graphite. In the ten years from 1944 to 1953 about 23,000 tons of graphite were produced.

Kaolin

Kaolin deposits are known in various areas of Quebec and northern Ontario, but their small size and the presence of impurities preventing suitable beneficiation have generally hindered their development. The kaolin deposit near St. Remi, Quebec, is on a north-trending ridge about half a mile wide, composed of nearly vertical dipping beds of Grenville quartzite and garnet gneiss striking north-northwest. The eastern side of the ridge is massive and unbroken, but the west side, throughout a zone about 1,000 feet wide, is shattered to a friable condition by faulting. The shattered zone has been traced at least 7,000 feet. In it are vein-like masses of kaolin ranging in width from a few feet to more than 100 feet. Within them are fragments and disseminated grains of quartzite. The kaolin formed by replacement of the shattered quartzite presumably through the agency of ascending hot liquids or vapours. The average kaolin content of the whole shattered zone has been estimated to be about 11 per cent. Nearly 25,000 tons were mined in 1942, but none since 1946. A by-product is washed silica sand suitable for glass-making. Several other occurrences of kaolin have been discovered in Quebec. One, near Point Comfort on Thirty-one-mile Lake, carries high-grade china clay. Other deposits are near Brebeuf on Lake Labelle, and near Chateau Richer, but are not of sufficient size and uniformity to warrant development.

Kyanite

Kyanite is fairly common in the sedimentary gneisses in many parts of the Canadian Shield. Extensive occurrences are near Mattawa and Sudbury, Ontario, and in Labrador, Manitoba, and the Northwest Territories. At Mattawa selected areas contain 15 per cent kyanite or more. East of Sudbury the kyanite occurs in a gneiss with lenses of sillimanite. An estimated 2 million tons of kyanite grading 18 per cent are present, from which a mullite aggregate, useful to the metallurgical and glass industries, could be produced.

Mica

Phlogopite has been mined from the same areas and types of deposits as described for apatite. Closing of the mine at Cantley, Quebec in 1966 brought to an end its long production of amber mica. Scrap phlogopite is shipped from large waste dumps near Kingston, and other past producers in southeastern Ontario and southwestern Quebec. All the muscovite of the Shield has come from Ontario. It occurs in pegmatites, apparently where they are especially coarse grained, such as the Eau Claire deposit near Mattawa, Ontario, which produced minor amounts from 1956 to 1960, the Purdy mine near North Bay, a famous producer during World War II, and the Parry Sound district. Total Canadian mica production in 1966 was 270 tons.

Nepheline Syenite

The large deposit of nepheline syenite at Blue Mountain, near Peterborough, Ontario, provides the only Canadian production. Three quarters of the production is exported. The Blue Mountain syenite body is a pear-shaped mass about 5 miles long with a maximum width of 1½ miles. It contains about 50% soda feldspar, about 20 to 25% of both nepheline and potash feldspar, and small quantities of easily removed iron-bearing minerals. Production has risen steadily from 146,068 tons in 1955 to 366,696 tons in 1966. The Blue Mountain Intrusion is one of a string of nepheline syenite bodies in Grenville Province of southeastern Ontario. These bodies are

developed close to contacts between granite and crystalline limestones, and have commonly been regarded as differentiates of the granite. However, detailed studies, undertaken in connection with the mining, strongly indicate that they are due to the granitization of calcareous sedimentary rocks of the Grenville Group. The nepheline-rich parts are essentially concordant with the enclosing gneisses, and show a banding that possibly represents bedding parallel and continuous with that of the gneisses into which they may grade.

Silica

Silica is mined from the Canadian Shield both as quartz and quartzite. The smelters at Sudbury and Flin Flon have used low-grade copper or gold ores as flux, the metal content being a valuable byproduct. Many pegmatite dykes mined for feldspar contain segregated masses of almost pure quartz, and although much of this has been discarded as waste, some is sold to smelters or manufacturers of ferrosilicon, as from River Valley near North Bay and Verona north of Kingston, both in Ontario. The upper third of the Lorrain Formation of the Aphebian Cobalt Group is a very pure quartzite. It is quarried near Sudbury for smelter flux, near Sault Ste. Marie, Ontario, for manufacture of silica brick, and at Killarney, Ontario, for the production of ferrosilicon.

Sulphur

Sulphur in the form of sulphuric acid is produced during base-metal mining and smelting operations from smelter gases or byproduct pyrite or pyrrhotite. Although large pyrite deposits are known, it is unlikely that they will be used in view of the very large volumes mined as a byproduct. Production is effected at smelters where sulphuric acid can be marketed nearby, often to industries integrated with the producer. Pyrite deposits utilized on the Shield are at the Noranda, Quemont, and Normetal mines in Quebec. Smelter gas sulphuric acid is produced from Quebec and Ontario ores processed at Arvida and Valleyfield, Quebec, and Copper Cliff and Port Maitland, Ontario. Elemental sulphur is derived from refining of nickel sulphide matte from Sudbury, Ontario, and Thompson, Manitoba, and was also recovered until 1959 at Noranda. Sulphur is recovered as ammonia sulphate byproduct from the nickel ores at Lynn Lake, Manitoba.

Talc

All the talc produced from the Shield has been from deposits near Madoc, Ontario. The Henderson and Conley mines (110a) have operated since 1937, and several low-quality grades of ground talc have been recovered. Production in 1966 was 15,100 tons. White, foliated talc forms a nearly vertical tabular mass 25 to 75 feet wide and at least 1,100 feet long. It lies between beds of dolomitic limestone into which it grades sharply. Within the talc body can be seen traces of the bedding planes of the original dolomitic limestone from which it was derived. It is probable that the action of solutions from a nearby granite led to the formation of the talc, and that an intermediate stage in the alteration was the conversion of the limestone to tremolite.

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VI. Geology of Southeastern Canada

Introduction 228

Precambrian 230

Paleozoic 235

Mesozoic and Cenozoic 296

Selected References 302



INTRODUCTION

Southeastern Canada lies southeast of the Canadian Shield and is composed of the St. Lawrence Platform, Appalachian Orogen, and Atlantic Continental Shelf (Fig. VI-1). These three major geological tectonic elements comprise a combined area of about 475,000 square miles, of which nearly three quarters is covered by sea.

St. Lawrence Platform borders the Canadian Shield. It is a narrow, northeast extension of the stable interior of North America bounded on the southeast by the Appalachian Orogen. The platform is underlain by relatively thin, nearly flat lying orthoquartzite, carbonate, and shale of early and middle Paleozoic age that lie unconformably on the crystalline rocks of the Canadian Shield. It is the oldest settled part of Canada, and therefore has long been an agricultural and highly industrialized region. The platform has an area of about 60,000 square miles and forms a remnant of a more extensive cratonic cover that was once continuous with the Hudson Platform to the north. St. Lawrence Platform is separated into three parts by crystalline rocks of the Canadian Shield: (1) southwestern Ontario, forming the marginal areas of Michigan Basin and Allegheny Trough; (2) Ottawa–Quebec Lowland (Quebec Basin and Ottawa Embayment); and (3) Anticosti–Mingan Islands and the coastal plain of western Newfoundland and southeastern coast of Labrador (Anticosti Basin).

The peninsula of southwestern Ontario is separated from the Ottawa–Quebec Lowland by the Precambrian rocks of the Frontenac Arch. The peninsula itself is underlain by a southwest-trending Precambrian basement high—the Algonquin Arch—from which Cambrian, Ordovician, Silurian, and Devonian rocks dip northwesterly into Michigan Basin and southeasterly into Allegheny Trough. The Paleozoic cover has a maximum thickness of nearly 5,000 feet near southern Lake Huron and beneath Lake Erie.

In the central or Ottawa–Quebec Lowland division of the platform the lower Paleozoic rocks are locally in fault contact with the Precambrian, which rises as an escarpment above the surface of the lowland. In Quebec the lowland is bordered on the southeast by thrusts referred to as "Logan's Line". Upper Cambrian and Upper Ordovician rocks vary in thickness from nearly 3,000 feet in Ottawa Embayment to about 9,500 feet in the central part of Quebec Basin.

The eastern division of the platform comprises Anticosti Basin, the Paleozoic rocks of which rise above the waters of the Gulf of St. Lawrence on Anticosti and

[1] The accounts of the geology of St. Lawrence Platform were prepared by B. V. Sanford, the Carboniferous and Permian systems by D. G. Kelley, and the Precambrian by H. Williams. In the remaining Paleozoic sections H. Williams is responsible for the geology of Newfoundland and W. H. Poole for that of the mainland. W. H. Poole is also responsible for the Mesozoic and Cenozoic sections and general coordination. When acknowledging, reference should be made to individual author and page number.



VI

Geology of Southeastern Canada

W.H. Poole, B.V. Sanford, H. Williams, and D.G. Kelley[1]

Gypsum in the Codroy Group, St. George's Basin, Newfoundland.

Mingan Islands and underlie narrow strips along the north shore of the Gulf of St. Lawrence and northwestern Newfoundland. The rocks range from Cambrian to Lower Devonian and are about 12,000 feet thick in the central part of Anticosti Basin.

On the Canadian Shield north of St. Lawrence Platform are several outliers of Paleozoic rocks. Most are small with only a few feet of flat-lying strata; others are preserved in tilted fault blocks and contain relatively thick successions.

The Appalachian Orogen extends from Newfoundland 2,000 miles southwestward along the Atlantic seaboard to Alabama, and northeastward to the edge of the Continental Shelf. The system was possibly once continuous with the Caledonian mountain chain on the eastern side of the north Atlantic in Ireland and Britain. To the southeast the orogen extends beneath the probably flat lying cover of upper Mesozoic and Cenozoic strata. The Appalachian Orogen in Southeastern Canada is a glacial region of uplands, highlands, lowlands, valleys, and fiords. The protolith of the orogen was an orthogeosyncline characterized by great mobility and relatively thick eugeosynclinal lower and middle Paleozoic sedimentary and volcanic deposits. The northwestern part is locally miogeosynclinal. Appalachian Orogen is divided into two parts: (1) a fold belt produced by the Ordovician Taconian Orogeny and the Devonian Acadian Orogeny; (2) an epieugeosyncline of late Paleozoic strata, part of which was mildly deformed by the Maritime Disturbance.

The Appalachian Geosyncline was dominated during the Cambrian and Ordovician by four facies belts: miogeosynclinal sediments in the northwest; a thick eugeosynclinal sequence in the centre; a relatively thin sequence of shale and minor limestone to the southeast; and an immensely thick greywacke–slate sequence still farther southeast. These rocks were locally deformed and intruded during the Ordovician Taconian Orogeny. In west Newfoundland two large klippen were transported westward from their place of deposition within the western part of the geosyncline onto the eastern edge of St. Lawrence Platform. The transported sequences include ultramafic Middle Ordovician plutons.

Silurian and Devonian deposition was restricted to several troughs separated by geanticlinal uplifts. Shallow-water marine sediments, conglomerate, volcanic rocks, and fluviatile terrestrial sandstones accumulated. During the Middle and Upper Devonian nearly the entire geosyncline was deformed, partly metamorphosed, intruded by granite batholiths, uplifted, and eroded during the Acadian Orogeny.

Subsequently, the fold belt acted as a platform upon which an epieugeosyncline with Upper Devonian to Permian deposits extended in a narrow zone from the Bay of Fundy to White Bay, Newfoundland. Deformation within the epieugeosyncline was restricted mainly to faulting and folding, local small intrusions, and incipient regional metamorphism. The deformed Carboniferous

rocks are in turn overlain by Upper Triassic terrestrial sedimentary and volcanic rocks deposited in a fault trough.

The Atlantic Continental Shelf comprises the slightly submerged area that extends eastward to the edge of the continental block. Its stratigraphy and structure are not known, but geophysical and drillhole data suggest that it consists mainly of relatively undisturbed Cretaceous and Cenozoic strata similar to those of the Atlantic Coastal Plain resting on a deformed Paleozoic basement.

PRECAMBRIAN

Tectonic Summary

The Precambrian of Southeastern Canada includes a variety of rocks with contrasting depositional, structural, and metamorphic histories. They are grouped into the four categories: (1) crystalline Helikian and older rocks of the eastern craton that underlie the Paleozoic strata of St. Lawrence Platform; (2) generally unmetamorphosed sedimentary and volcanic rocks, possibly early Hadrynian, that were deformed and locally intruded during the Hadrynian Avalonian Orogeny; (3) thick late Hadrynian sedimentary and volcanic rocks; (4) thin platformal quartzite, of latest Hadrynian and/or earliest Cambrian age, that unconformably and conformably overlies Hadrynian rocks and conformably underlies fossiliferous Lower Cambrian shale. In addition, poorly dated rocks that may be in part Hadrynian form the oldest recognizable rocks in the northwestern Appalachian Region. They are described with Cambrian rocks.

A peneplain that developed on Helikian and older rocks of Grenville Province of the Canadian Shield dips southeasterly below Paleozoic strata of the St. Lawrence Platform and under the northwestern Appalachian Geosyncline. It may extend beneath the entire geosyncline as far southeast as the edge of the continent under the Atlantic Ocean, but there is no direct evidence of this. Parts of the craton adjacent to the Appalachian Geosyncline were more or less continuously depressed during the Paleozoic and formed Allegheny Trough, Quebec Basin, and Anticosti Basin (Fig. VI-1). To the west, the intracratonal Michigan Basin was a strongly negative element. The intervening Algonquin Arch consisted of a broad elongate structure that was depressed to a much lesser extent but maintained some degree of subsidence during the Paleozoic. Cambrian and Lower Ordovician sediments thin toward it, and as a result of slight uplift, are bevelled and overlapped by Middle Ordovician sediments.

Precambrian rocks of indefinite age, possibly early Hadrynian, occur within the Trinity Belt of the southeastern Appalachian Orogen in eastern Newfoundland, Cape Breton Island, and southern New Brunswick. They are the oldest rocks exposed in the orogen, and if they are younger than the Grenvillian Orogeny they represent the first strata to be formed on or adjacent to the flanks of the craton. The tectonic element in which they were deposited cannot be defined. The rocks in Newfoundland are dominantly volcanic assemblages, whereas argillite, limestone, and quartzite were deposited in Cape Breton and New Brunswick. In Newfoundland, volcanic rocks, mainly subaerial, and red beds, were deposited on a basement that is now covered. Locally intercalated conglomerates containing granite pebbles suggest some tectonic activity; on the other hand, the sedimentary assemblages of Cape Breton Island and southern New Brunswick suggest platformal deposition on a stable basement, now covered. A comparison of these strata with similar rocks of the Grenvillian Orogen shows that they could be Aphebian and represent a remnant of a much larger craton upon which the Appalachian Geosyncline evolved.

During the middle Hadrynian Avalonian Orogeny the early Hadrynian volcanic and sedimentary rocks in eastern Newfoundland were folded, weakly metamorphosed, intruded by granite, uplifted, and eroded (Fig. VI-2). Probable lower Hadrynian rocks in Cape Breton Island and southern New Brunswick may also have been deformed, but the structural relations are obscure.

During the late Hadrynian, thick sediments and volcanics were deposited in the Trinity Belt in what must have been a deep trough. The lowest assemblages are slate and greywacke of deep-water marine facies succeeded by volcanic rocks and an upper assemblage of arkose and red conglomerate—a typical shallow-water continental facies. The younger rocks were deposited throughout Trinity Trough, whereas the older are known only in eastern Newfoundland. The upper Hadrynian rocks of Newfoundland were deposited in elongate troughs bordered by uplifted lands formed of older volcanic, sedimentary, and granitic rocks. The assemblage records a progressive shallowing of the seas, a change to non-marine conditions, and an increasing supply of coarse detritus.

In latest Hadrynian or earliest Cambrian time (Fig. VI-4) the site of Trinity Trough became Avalon Platform. A relatively thin, pure quartzose sandstone was deposited over much of the platform in a shallow-water sea. Stable, low-lying land existed in easternmost Newfoundland and northwestern Cape Breton. About the same time the oldest rocks in the Paleozoic Notre Dame Trough, described with the Cambrian, were deposited along the northwestern edge of Appalachian Geosyncline in the Eastern Townships of Quebec and the Burlington Peninsula of Newfoundland. These rocks are mainly sandy and shaly facies with some volcanics and carbonate beds. On St. Lawrence Platform thick, coarse, red, clastic beds were deposited in Lake Superior Basin near Manitoulin Island. They were presumably derived from a Precambrian land area located in northern Michigan.

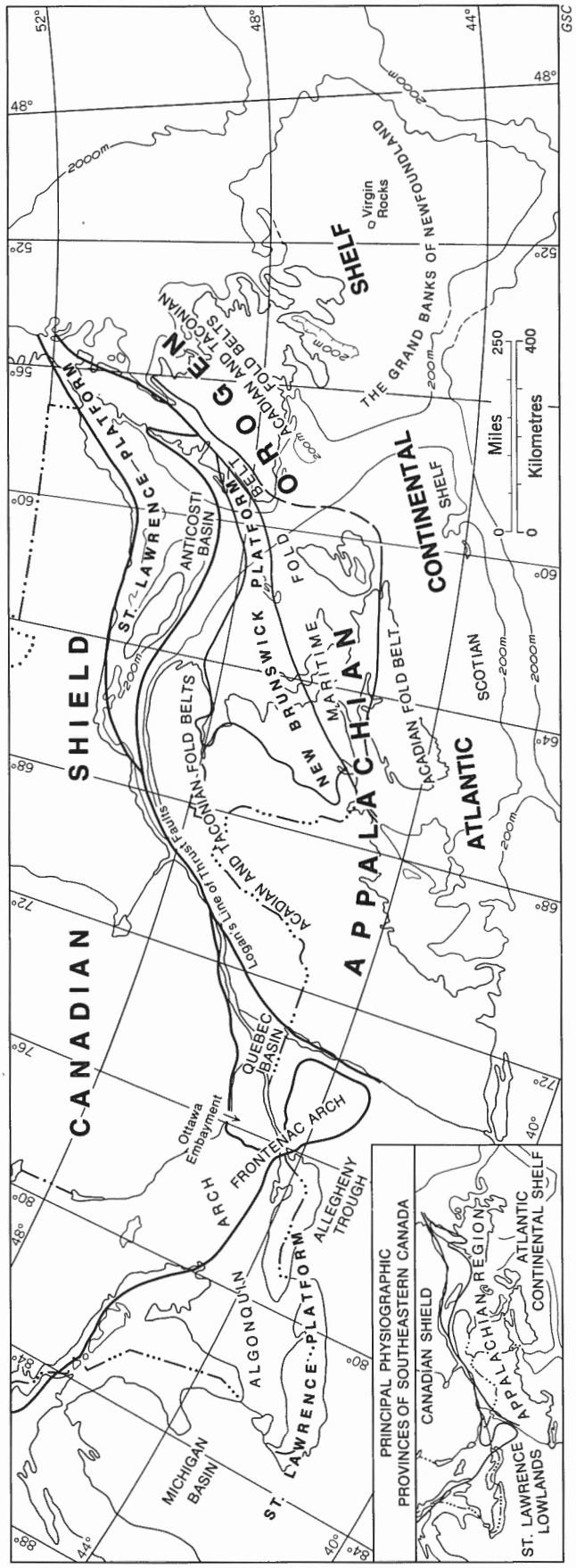


FIGURE VI-1 Principal geological elements of Southeastern Canada.

Helikian and Older

Helikian and older rocks of the Great Northern Inlier of Newfoundland and small inliers to the south are referred to as the Long Range and Indian Head Range Complexes (Clifford and Baird, 1962; Riley, 1962). They consist of schist and gneiss that have been tightly folded on northerly trending axes. The Long Range Complex was intruded by granites from which the biotite yielded K–Ar dates of 945 and 960 million years. Biotite from pegmatite and gneiss in the Indian Head Range Complex yielded dates of 830 and 900 million years.

The Long Range Complex consists of well-layered assemblages of psammitic schist and gneiss, quartzite, pelitic and semi-pelitic schist and gneiss, and hornblende schist and gneiss. Most are thought to be of sedimentary origin, and the assemblages of amphibole and plagioclase, with local occurrences of sillimanite, indicate regional metamorphism to almandine amphibolite facies. The Indian Head Range Complex occurs in two small inliers southwest of the Long Range Complex. It consists of mafic to silicic rocks that are largely of plutonic derivation and include anorthosite, pyroxenite, a variety of granitic gneisses, and granite pegmatite. The anorthosite is similar to those of the Grenville and Nain Provinces of the Canadian Shield where they have been interpreted as products of the Elsonian Orogeny. A large anorthosite body within Paleozoic schist and gneiss, 20 miles east of the Indian Head Range Complex, is also regarded as a Grenvillian inlier, but the relationship of the anorthosite to the Paleozoic rocks has been obscured by metamorphism and deformation. A northeast-trending swarm of diabase dykes cuts the Long Range Complex. These are believed to be feeders for the basalt flows that unconformably overlie gneiss and schist near the Strait of Belle Isle, and in turn are succeeded by Lower Cambrian strata. Mafic dykes, probably of equivalent age, also cut the Indian Head Range Complex.

Hadrynian

Early Hadrynian

In eastern Newfoundland the Harbour Main Group, probably early Hadrynian, comprises more than 6,000 feet of dominantly subaerial andesite, rhyolite, rhyolitic welded tuff, and intercalated red sandstone (McCartney, 1967; McCartney, *et al.*, 1966; Rose, 1952). In the lower half of the succession there is well-bedded green to grey siltstone and slate of probable marine origin. The volcanic rocks have a Rb–Sr date of 568±29 m.y. (Fairbairn, *et al.*, 1966). This date is equivalent to Middle Cambrian and much too young because the group is overlain unconformably by late Hadrynian and Lower Cambrian rocks.

The Love Cove Group (Jenness, 1963) is correlated lithologically with the Harbour Main Group and consists of 15,000 feet of alternating volcanic and sedimentary rocks. Silicic lavas and pyroclastic rocks of intermediate composition are present, and some basalt is probably represented by chlorite–epidote schists. Sedimentary rocks

are dominantly feldspathic greywacke with some metamorphosed conglomerate and slate. The Love Cove Group is everywhere more deformed and metamorphosed than nearby Hadrynian rocks, and pebbles of Love Cove chlorite schist have been reported in Hadrynian conglomerates. The Love Cove Group lies nearer the central Newfoundland axis of intense Paleozoic deformation than does the Harbour Main Group, and it has been locally feldspathized and granitized during the Paleozoic.

The Deer Park Pond and Southern Hills Formations of Burin Peninsula are dominantly volcanic assemblages that are possibly equivalent in part to the Harbour Main and Love Cove Groups. They are overlain by Hadrynian and Cambrian sedimentary rocks.

On Cape Breton Island the George River Group consists of quartzite and other clastic metasedimentary rocks, marble, quartzo-feldspathic schist and gneiss, amphibolite, and probable metavolcanic rocks (Weeks, 1954). In the type area two mappable units were recognized by early workers: a lower unit of quartzite, schist, and gneiss; and an upper unit of marble. The most distinctive rock types are marble and quartzite, and although abundant in the type area, they are subordinate to schist and gneiss in northern Cape Breton. Sedimentary features such as crossbedding, ripple-marks, and scour-and-fill structures are locally preserved and indicate shallow-water deposition. The George River Group is overlain with marked angular unconformity by Middle Cambrian strata, and Hadrynian rocks nearby conformably underlie the Cambrian.

The Green Head Group near Saint John, New Brunswick, resembles the George River Group in most respects and is composed mainly of limestone, quartzite, argillite, and limestone-conglomerate that locally grade into schist and gneiss (Alcock, 1938). Some quartzite is crossbedded, and certain limestone horizons contain the stromatolite *Archaeozoon acadiense*.

Avalonian Orogeny

During the middle Hadrynian, the Harbour Main Group in Newfoundland was deformed, weakly metamorphosed, intruded by the Holyrood granite, uplifted, and eroded during the Avalonian Orogeny. The Harbour Main Group is overlain with angular unconformity by Lower Cambrian strata and with local unconformity by the late Hadrynian Conception Group that locally contains clasts of granite like the Holyrood and hence is interpreted as younger than the granite (McCartney, 1967). The pre-orogenic rocks formed highlands and yielded detritus to narrow troughs and basins during the late Hadrynian. Avalonian metamorphism was low grade and produced slate in the Harbour Main and greenschist in the Love Cove Group. Folds are open and north-trending. The Holyrood granite consists dominantly of massive, coarse-grained, biotite quartz monzonite and granodiorite

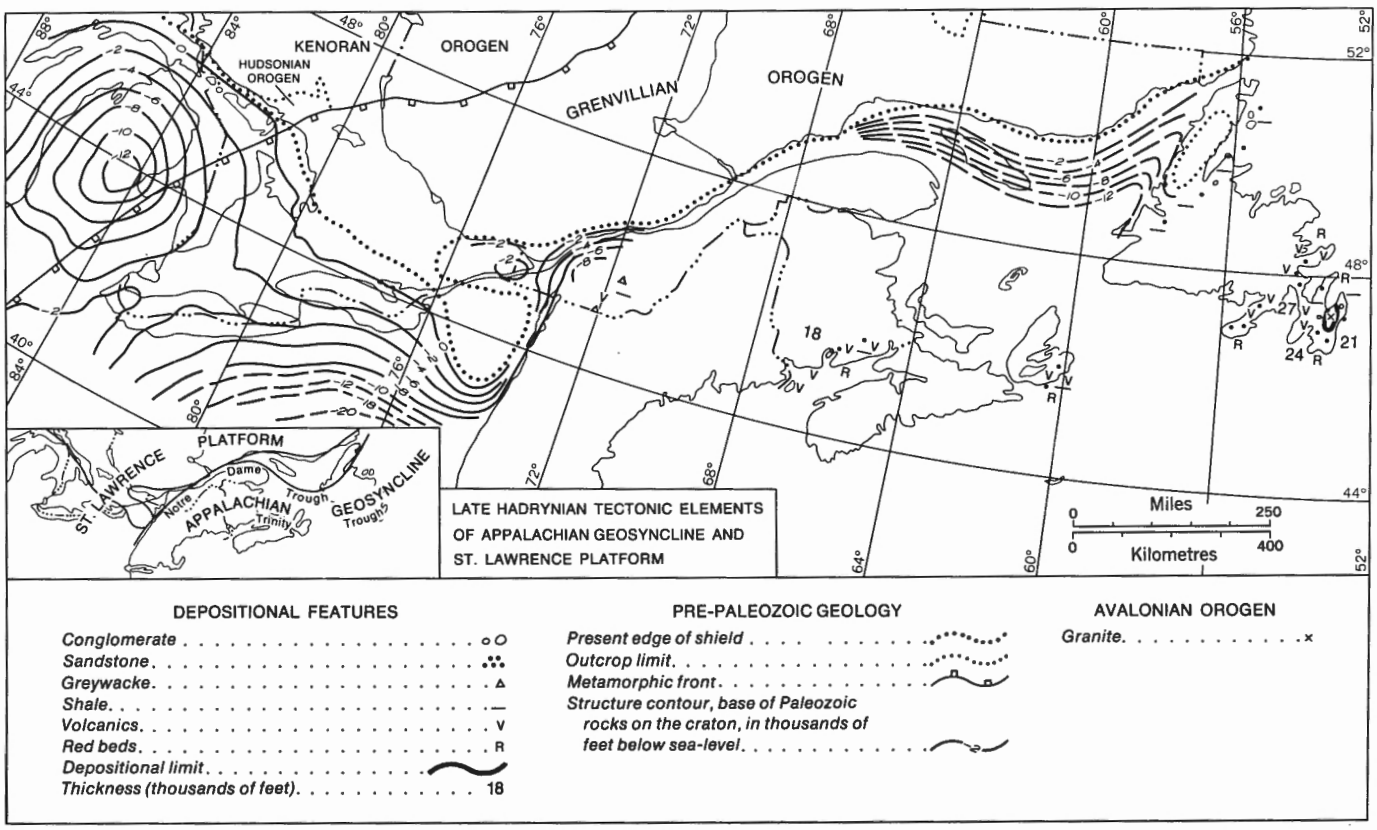


FIGURE VI-2. Late Hadrynian sedimentation, volcanism, and tectonism in Southeastern Canada, with features of the middle Hadrynian Avalonian Orogeny and the pre-Hadrynian basement surface of St. Lawrence Platform.

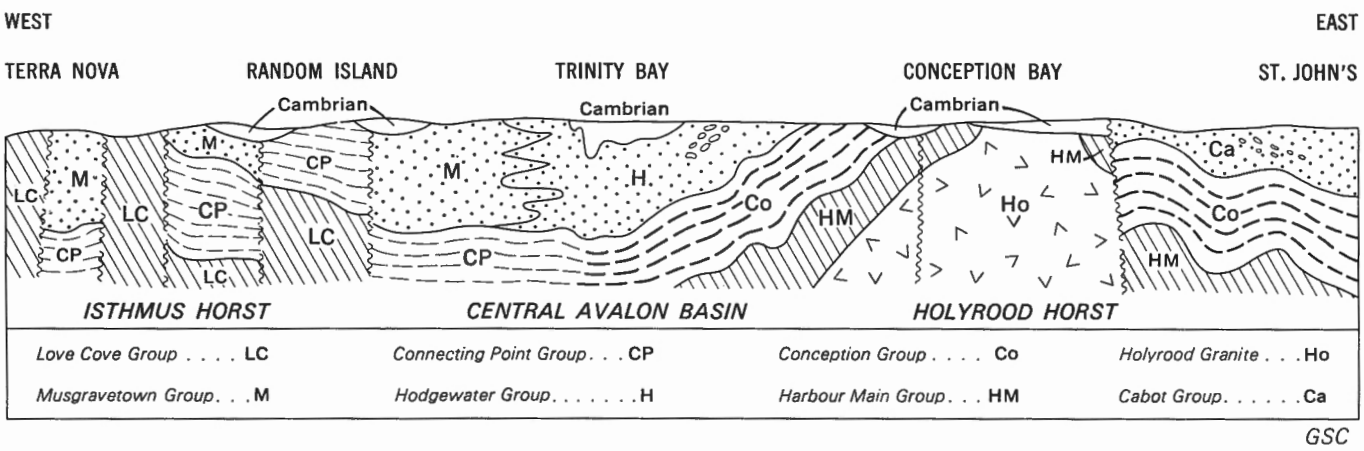


FIGURE VI-3. Diagrammatic cross-section of Precambrian and Cambrian rocks of Avalon Platform, eastern Newfoundland (modified from McCartney, et al., 1966).

that have been slightly cataclastized and altered. The granite yielded a whole-rock Rb–Sr date of 574±11 m.y. (McCartney, *et al.*, 1966), a date close to the Precambrian–Cambrian boundary and believed to be too young.

Evidence for the Avalonian Orogeny in Cape Breton Island and southern New Brunswick is sparse. The George River Group is overlain unconformably by Middle Cambrian rocks that farther east conformably overlie late Hadrynian volcanic rocks of the Fourchu Group. Structural relations of the Green Head Group and late Hadrynian volcanic rocks of the Coldbrook Group are obscure. Granitic clasts occur in both the Coldbrook and Cambrian strata, but the granite source has not been recognized. Metamorphism leading to schist and gneiss in the Green Head and George River Groups is believed to be Paleozoic.

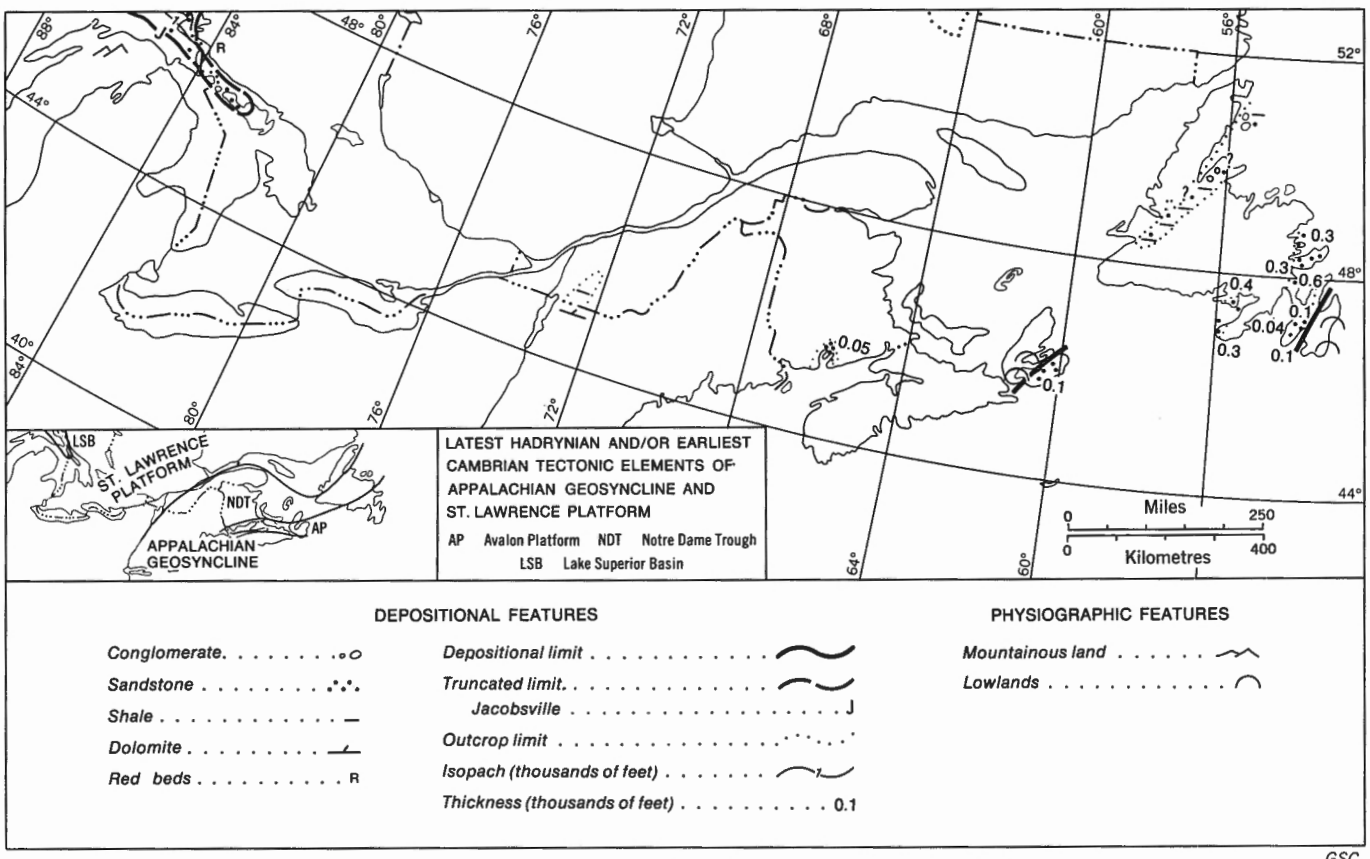


FIGURE VI-4. Latest Hadrynian and/or earliest Cambrian sedimentation and tectonism in Southeastern Canada.

Late Hadrynian (Fig. VI-2)

On Avalon Peninsula, Newfoundland, the Conception and Connecting Point Groups (McCartney, 1967; Jenness, 1963) are stratigraphically equivalent and lie at the base of the upper Hadrynian sequence (Fig. VI-4). The groups consist of slate, siltstone, and greywacke, with minor quartzite, conglomerate, and volcanic rocks. They were deposited in Trinity Trough. The Conception Group is about 7,000 feet thick and the Connecting Point Group between 9,000 and 30,000 feet thick. Locally, the Conception includes 100 feet of conglomerate and red sandstone at the base, and red sedimentary rocks of the Hibbs Hole Formation at the top. These lithologies suggest alternating changes between dominantly deep-water marine deposition and shallow-water continental deposition. Impressions of soft-bodied frond-like Metazoa as long as 30 cm have been recently discovered by S.B. Misra in ripple-marked mudstone of the upper part. The Conception Group conformably overlies the Harbour Main Group except near the Holyrood granite where it is unconformable; there the basal Conception conglomerates contain pebbles of Holyrood granite. The Connecting Point is faulted against older and most younger rocks. Local Hadrynian deformation is suggested by the direction of fold axes within the Connecting Point Group that in places trend a little west of north, at variance with prominent northeasterly trending fold axes in overlying Hadrynian rocks.

The Bull Arm Formation (McCartney, 1967; Jenness, 1963) consists of basalt, andesite, mafic and silicic volcanic breccia and pyroclastic rocks, and silicic lavas, as well as tuffaceous sedimentary rocks, red and green arkose, slate, siltstone, and conglomerate. A whole-rock Rb–Sr date on the volcanic rocks is 494±30 m.y. which is too young (Fairbairn, *et al.*, 1966). Other Rb–Sr dates on equivalent rocks in Cape Breton Island and southern New Brunswick are 509±40 m.y. and 495±30 m.y., respectively, and these also are too young. The Bull Arm Formation has a maximum thickness of 8,000 feet. Slate and siltstone are most common near the base and are similar to underlying beds of the Connecting Point Group. The contact is conformable, and the base of the Bull Arm Formation is drawn arbitrarily at the lowest volcanic member. In some places the Bull Arm is underlain by the Cannings Cove Formation that overlies the Connecting Point Group with angular unconformity. The Cannings Cove consists of more than 2,000 feet of red and green conglomerate and sandstone with clasts of volcanic rocks that resemble the Bull Arm, suggesting that its deposition was contemporaneous with the volcanism in adjoining regions.

Throughout Burin Peninsula some of the volcanic rocks included in the Deer Park Pond and Southern Hills Formations may be equivalent to the Bull Arm Formation.

Late Hadrynian sedimentary rocks in Newfoundland are generally referred to the Musgravetown, Hodgewater, and Cabot Groups (McCartney, 1967; Jenness, 1963; Rose, 1952), except in the vicinity of Fortune Bay where they are referred to the Dantzig Group, or Doten Cove and Chapel Island Formations. All these formations are overlain by white quartzite of the Random Formation or its equivalent, the Blue Pinion Formation; the quartzites are in turn succeeded by fossiliferous Cambrian shales. All upper Hadrynian sedimentary rocks below the Random Formation are considered equivalent, at least in part, but individual formations are not co-extensive except locally where Musgravetown beds have been traced through facies changes into strata of the Hodgewater Group. The Musgravetown Group consists of about 10,000 feet of arkosic sandstone, siltstone, slate, and conglomerate that conformably overlie the Bull Arm Formation. The Hodgewater and Cabot Groups are similar and have approximate maximum thicknesses of 17,000 and 14,000 feet, respectively. The Hodgewater and Cabot Groups lie east of the Musgravetown Group and both rest conformably on the Conception Group on the opposite flanks of the Holyrood horst.

On Cape Breton Island the upper Hadrynian Fourchu Group (Weeks, 1954) consists of silicic to intermediate pyroclastic rocks locally with intercalated flows, minor shale, siltstone, and greywacke. A distinctive pyroclastic member commonly occurs at the top of the succession. Many rocks of the Fourchu Group display a schistosity or parallel elongation of fragments in some places, whereas in others they are difficult to differentiate from nearby unmetamorphosed volcanic rocks of Middle Cambrian age. The Morrison River Formation conformably overlies the Fourchu Group and consists of a basal unit of crossbedded red sandstone with conglomerate interbeds and a middle unit of interbedded sandy slate and quartzite. Pebbles in the conglomerate are mainly vein quartz and sedimentary rocks, with only a few volcanic rocks. The formation has a maximum thickness of about 2,400 feet in the northeast, a minimum of 130 feet in the southwest, and apparently thins to zero in subsurface to the northwest.

In southern New Brunswick the upper Hadrynian Coldbrook Group consists of 18,000 to 33,500 feet of flows and pyroclastic rocks in roughly equal proportions, with silicic to intermediate varieties predominating over mafic types (Alcock, 1938). Minor conglomerate, phyllite, and schistose sandstone are also included. One conglomerate bed contains red granite pebbles, and the sandstones include numerous fragments of red feldspar, both of unknown provenance. The contact with the underlying Green Head Group has been interpreted as an angular unconformity, but this relationship is uncertain. The Coldbrook Group is overlain by the Ratcliffe Brook Formation (Hayes and Howell, 1937; Alcock, 1938), from 20 to 2,000 feet of fine-grained red micaceous sandstone with conglomerate at its base. Clasts are mainly volcanic pebbles and granite in one locality. Rapid variation in thickness suggests filling of an irregular erosion surface

on Coldbrook volcanics, but the nature of the contact is uncertain.

Latest Hadrynian and/or Earliest Cambrian (Fig. VI-4)

On Avalon Platform in eastern Newfoundland, Cape Breton Island, and southern New Brunswick, deposition of upper Hadrynian red beds in Trinity Trough changed gradationally to marine, white, pure quartzose sand without evidence of interruption in deposition. The quartzite is overlain conformably by fossiliferous Lower Cambrian shale. It is probably of the same general age in the three areas, but it may be partly transgressive and diachronous.

In Newfoundland the Random Formation consists of white quartzite pure enough to be a source for industrial silica (McCartney, 1967; Jenness, 1963). It is 200 to 300 feet thick. The lower part contains red shale and siltstone similar to the underlying Musgravetown and Hodgewater Groups; the upper part has beds of grey, red, and green shale like the overlying Cambrian formations.

On Cape Breton Island the upper part of the Morrison River Formation (Weeks, 1954) consists of as much as 350 feet of white quartzite that thins northwesterly to zero in central Cape Breton where Middle Cambrian rocks rest directly on the George River Group.

In southern New Brunswick the Glen Falls Formation (Hayes and Howell, 1937) consists of 20 to 50 feet of quartzose sandstone. The lower part is white sandstone and the upper 3 to 6 feet is black sandstone.

The Jacobsville Formation occurs as subsurface outliers beneath Paleozoic strata and rests with angular unconformity upon Aphebian and older crystalline rocks between western Manitoulin Island and Keweenawan Peninsula of northern Michigan (Hamblin, 1958). It consists of red sandstone, conglomerate, and shale of non-marine origin. The maximum thickness is more than 700 feet at Sault Ste. Marie. The Jacobsville was deposited along the north side of an east-trending land that extended across northern Michigan. The age of the formation is uncertain; it has been variously classified as Proterozoic or Lower, Middle, and Upper Cambrian. Inasmuch as the Jacobsville is disconformably overlain by Upper Cambrian coarse marine clastic strata and is lithologically unlike other Cambrian rocks in the region, it is probably Hadrynian and represents the oldest post-Grenvillian strata on St. Lawrence Platform.

PALEOZOIC

Cambrian

Tectonic Summary

During the Cambrian the limits and tectonic characteristics of the St. Lawrence Platform and Appalachian Geosyncline were clearly established for the first time (Figs. VI-5, 6). Within the geosyncline, thick non-volcanic clastic rocks were deposited along the northwestern and southeastern flanks of Notre Dame Trough that was bordered in southern Quebec by a zone of volcanism probably in late Cambrian time. A thin sequence of fossiliferous Cambrian shale with minor limestone was deposited on the folded and flat-lying rocks and granite within the Hadrynian Trinity Trough. Cambrian seas spread laterally to overlap the Precambrian lands, and by the late Cambrian at least, largely covered the lands and probably were connected with the seas of Notre Dame Trough. The thickness, distribution, sedimentary facies, and overlap relationships show that the Cambrian strata accumulated in northeast-trending depositional basins on Avalon Platform, and the faunal similarities suggest that they had direct marine connections. The platform was a gently subsiding stable element from Cambrian to early Ordovician. It was partly deformed by subsequent Paleozoic orogenies and partly preserved as stable blocks that shielded the Cambrian–Ordovician cover from the deformation in adjacent troughs. The relationships are unknown between the Cambrian strata of Avalon Platform and those of Notre Dame Trough to the west and Meguma Trough to the east.

Southeast of Avalon Platform very thick greywacke quartzite accumulated in Meguma Trough, largely by turbidity current deposition. The detritus was probably derived from a landmass lying to the southeast in an area now covered by the Atlantic Ocean.

Orthoquartzitic sandstone and carbonate rocks were deposited on St. Lawrence Platform. Eastern Anticosti Basin near Notre Dame Trough was inundated by shallow seas in the early Cambrian. The basal arkosic rocks and minor basaltic flows are overlain by a thick succession of Lower to Upper Cambrian orthoquartzitic sandstone and dolomite. During the late Cambrian the sea transgressed into Quebec Basin and Ottawa Embayment and onto Algonquin Arch in southwestern Ontario leaving marine deposits of sandstone and dolomite that rest unconformably on the Precambrian basement. Voluminous quartz detritus was shed into southern Quebec Basin from mountains along the north side of Frontenac Arch. The Cambrian strata deposited in Quebec Basin thicken southeastward and probably grade into geosynclinal rocks now exposed in the northwesterly displaced thrust sheets of Appalachian Orogen.

The fauna in the pelitic sequence covering Avalon Platform are typical of an European or Atlantic faunal realm of the Cambrian and are characterized by the trilobite *Paradoxides*. Those in the carbonates of the St. Law-

rence Platform are typical of the North American or Pacific faunal realm and include such characteristic fossils as archeocyathids and the trilobite *Olenellus*. The presence of the two distinct faunal provinces in relatively close proximity has been attributed to: (1) a linear land barrier—the "New Brunswick Geanticline", or a series of island arcs separating the two areas of deposition; (2) a deep-water trough along the central axis of the geosyncline that prevented mixing of the shallow-water fauna on each side; (3) Cambrian and older rocks on the southeast side of the Appalachian Geosyncline representing part of another continent that was separated from North America during Cambrian and Ordovician time by a proto-Atlantic Ocean. Presumably this early Paleozoic ocean basin closed during middle Paleozoic time and later reopened along a new line to become the present Atlantic Ocean during Mesozoic time. The simple geographical separation of the two realms has been complicated by the recent discovery of Middle Cambrian trilobites typical of both the Atlantic and Pacific realms in the axial part of Notre Dame Trough in northeastern Newfoundland.

Appalachian Geosyncline

Notre Dame Trough. Cambrian rocks are not sufficiently exposed or dated by fossils to reconstruct fully the tectonic conditions of deposition within the trough. At Quebec City, along the northwestern flank of the trough in *southern Quebec,* the oldest rocks are the Charny Formation (Osborne, 1956), which is part of what was formerly called the Sillery Formation. The Charny is perhaps 2,000 feet or more thick and is bounded above and below by thrust faults. It is mainly Lower Cambrian and consists of red, green, and grey marine shale, sandstone, and fine pebble-conglomerate. The Lauzon Formation is mainly Upper Cambrian and comprises about 2,000 feet of red shale; grey, black, and green siltstone and sandstone; and some limestone-conglomerate. Rocks of similar character occur here and there along the margin of the trough in southern Quebec. In the Eastern Townships of southern Quebec the uppermost Hadrynian and Cambrian Oak Hill Group forms the core of the Sutton Mountain anticlinorium (Cady, 1960, 1967; Eakins, 1964). The group consists of several formations totalling about 2,000 feet of rocks whose dominant lithologies in ascending order are intermediate to mafic volcanic rocks, greywacke, dolomite, quartzite, black slate, quartzite, dolomite, siltstone, quartzite, and black slate. These formations reflect deposition alternating between geosynclinal and platformal conditions and grade into metamorphosed rocks of the Sutton schist. Northeastward along trend (Béland, 1962) the core of the anticlinorium consists of the Rosaire Group overlain in the northwest by the Armagh Group and in

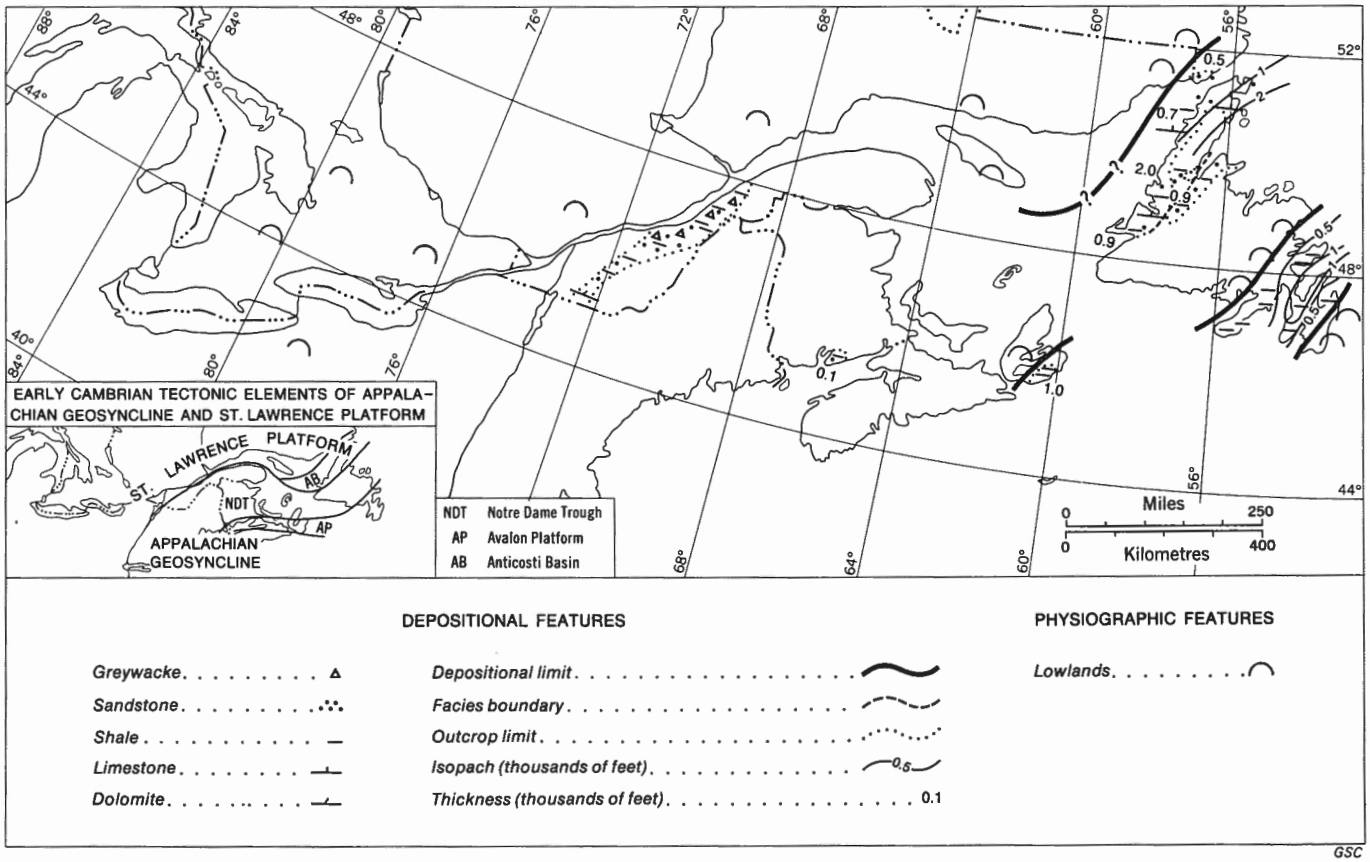


FIGURE VI-5. Lower Cambrian sedimentation and tectonism in Southeastern Canada.

the southeast by the Caldwell Group. The age and structural relationship of these groups is unknown, but perhaps most of these unfossiliferous rocks are of Cambrian age. The Rosaire consists of white and grey quartzite, siltstone, and slate a few thousand feet thick; the rocks are characteristic of marine shallow-water deposition in a tectonically stable environment such as a platform. The probably younger and equivalent Armagh and Caldwell Groups are each about 25,000 feet thick and consist of green, red, and grey impure sandstone; arkosic sandstone and greywacke; siltstone; slate; and quartzite. The Caldwell also contains some mafic volcanic rocks. The Rosaire and Caldwell Groups grade into schists of the Bennett Group. The sedimentary rocks of the Armagh and Caldwell Groups are similar to the Cambrian formations at Quebec City and their equivalents to the northeast and southwest. Parts of the Armagh and Caldwell Groups are probably early Ordovician. The abundant feldspar detritus, mainly plagioclase, suggests derivation from a crystalline terrane like the Canadian Shield to the north. Similar rocks of Cambrian and possibly Cambrian age occur along trend in northern Gaspé Peninsula. Late Middle and early Upper Cambrian trilobites have been collected from dark grey slate within the Quebec Group, as well as a Middle Cambrian trilobite from a limestone clast in a conglomerate (Ollerenshaw, 1967). A small area of late Middle and early Upper Cambrian limestone and shale of the Corner-of-the-Beach and Murphy Creek Formations occurs on eastern Gaspé Peninsula (McGerrigle, 1950). These rocks are somewhat unique because they lie south of the trend of the Quebec and Schickshock Groups. The Cambrian trilobites in Gaspé Peninsula are characteristic of the Pacific faunal realm.

In *western Newfoundland,* Cambrian and possibly Cambrian rocks occur along western Notre Dame Trough and in allochthonous sequences derived from the trough now resting on the St. Lawrence Platform (Poole, 1967). In the trough the Fleur de Lys Group of Burlington Peninsula (Neale and Nash, 1963; Neale and Kennedy, 1967) and its probable equivalents near Deer Lake the Mount Musgrave Formation and Grand Lake Brook Group (Brückner, 1966; Lilly, 1967), are undated, and stratigraphic relationships to adjacent rocks are not clear. The rocks are dominantly psammitic and pelitic schists several thousand feet thick, with minor plutonic boulder meta-conglomerates. All are regionally metamorphosed to the amphibolite or greenschist facies. Locally on Bur-

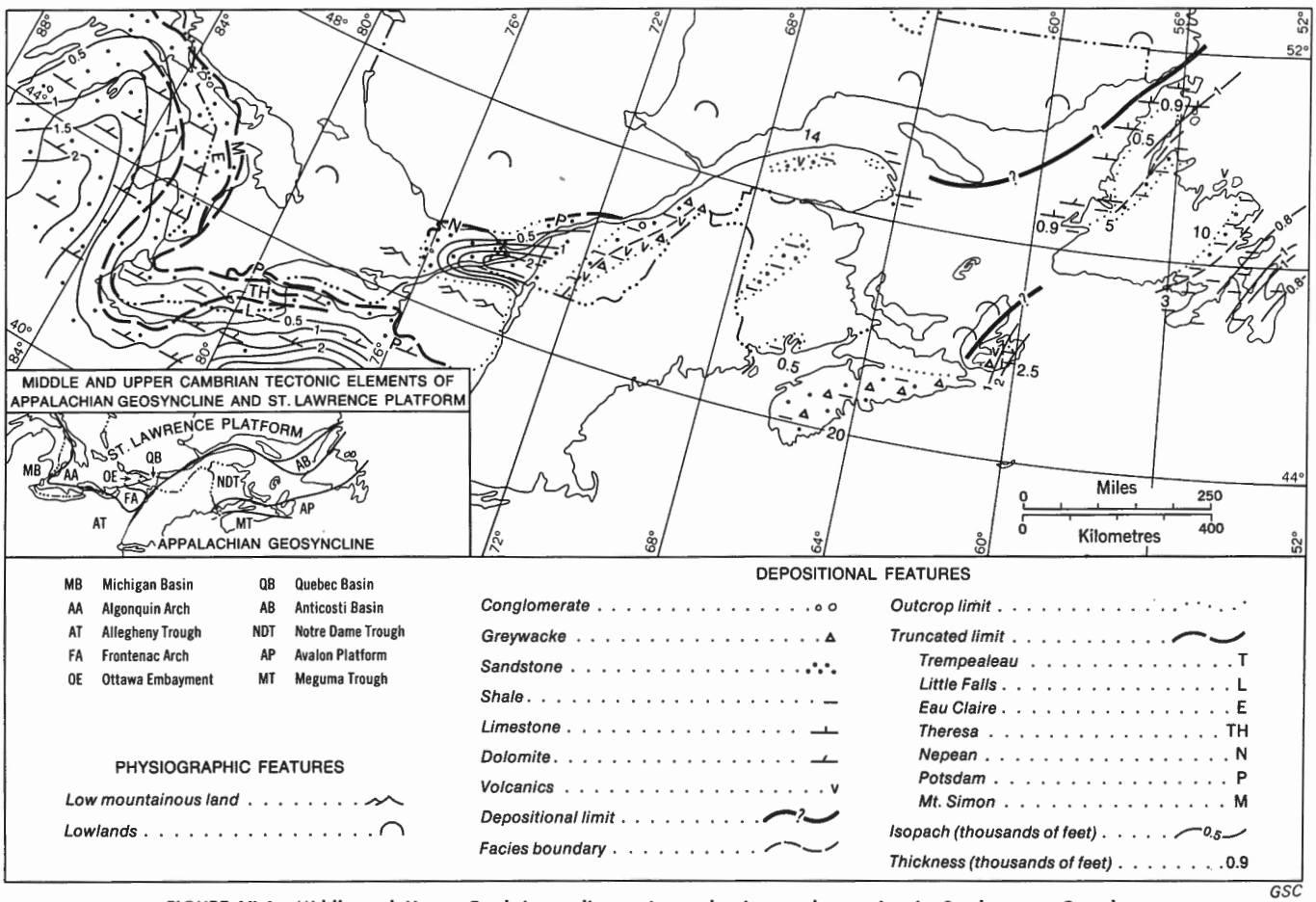


FIGURE VI-6. Middle and Upper Cambrian sedimentation, volcanism, and tectonism in Southeastern Canada.

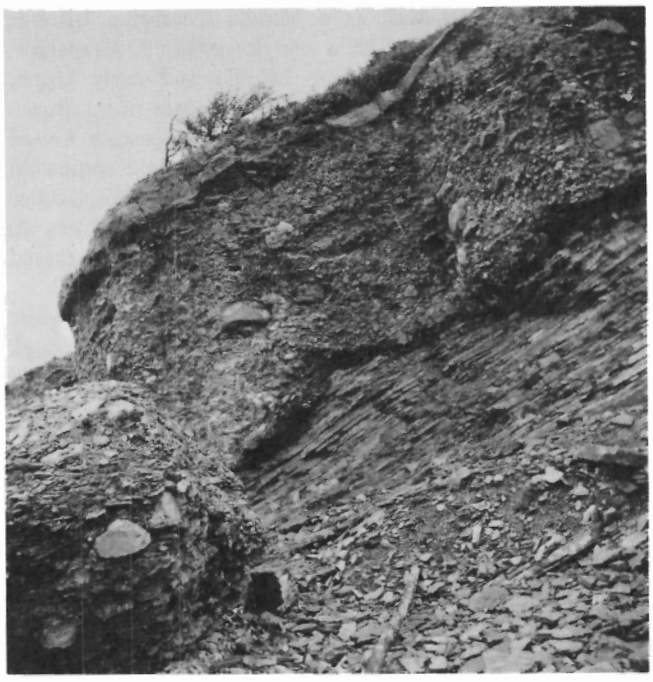


PLATE VI-1. Cambrian thin-bedded limestone and limestone breccia of Cow Head Group at northern end of Humber Arm klippe, Cow Head, Newfoundland.

lington Peninsula thin-bedded marble and limestone breccia beds occur along the western part of the Fleur de Lys outcrop belt, and green amphibolitic and chloritic schists of probable volcanic origin occur in the eastern exposures. The Mount Musgrave Formation and Grand Lake Brook Group appear to underlie or grade westerly into autochthonous Lower Cambrian carbonates of the Labrador Group on St. Lawrence Platform. Their distribution suggests that the contact with the Labrador Group is conformable and that they are partly older. These trough rocks have had a more involved structural and metamorphic history than nearby Paleozoic rocks in the trough to the east, implying that the contact is an unconformity or a fault. From the information at hand, the Fleur de Lys rocks and equivalents probably represent a marginal trough succession equivalent to or lying below the Cambrian strata of the St. Lawrence Platform.

The allochthonous Cambrian sequences of western Newfoundland occur within the Humber Arm klippe and the Hare Bay klippe (Rodgers and Neale, 1963). The Humber Arm klippe contains Cambrian rocks of two facies: a carbonate facies of the Cow Head Group to the north, and a clastic facies to the south and east. The lower part of the Cow Head Group (Kindle and Whittington, 1958) consists of 300 to 400 feet of fossiliferous limestone, shale- and limestone-conglomerate, and Middle and Late Cambrian breccia (Pl. VI-1). Limestone clasts are approximately the same age and lithology as the immediately underlying beds, but are of a different facies and contain a different fauna than the now exposed carbonate rocks of St. Lawrence Platform. Presumably the limestone clasts were derived from nearby contemporaneous carbonate banks not now exposed, the brecciated material being periodically discharged down submarine slopes into the limy muds accumulating at the base. This depositional environment was not local, for the conglomerates at Levis, Quebec, and in the Quebec Group along the south shore of the St. Lawrence River also contain boulders of Cambrian limestones remarkably similar lithologically and faunally to those in the Cow Head Group. Thus the facies from which the boulders were derived extended along strike, although not necessarily continuously, for about 800 miles. Probably this was an easterly facies of the carbonates of the St. Lawrence Platform. The clastic facies in the southern and eastern part of the klippe (Brückner, 1966; R. K. Stevens) consists of 2,500 to 3,000 feet of generally thin bedded shale, sandstone, limestone, and limestone-conglomerate, referred to in ascending order as the Summerside, Irishtown, and Cooks Brook Formations. Both the Irishtown and Cooks Brook Formations contain clasts of Early Cambrian limestone, and in addition the Cooks Brook contains indigenous Middle Cambrian fossils in its lower part and Tremadocian graptolites at the top. These rocks were at one time considered to be Middle Ordovician and to stratigraphically overlie Cambrian to Middle Ordovician autochthonous carbonates of the St. Lawrence Platform. With the discovery of Cambrian and Lower Ordovician fossils, however, it now appears certain that the Cow Head Group and clastic rocks to the south are in large part contemporaneous with the autochthonous carbonate sequence and are indeed allochthonous (Rodgers and Neale, 1963). The clastic facies was probably deposited farther east within the Notre Dame Trough than the platform–trough border zone where the Cow Head Group was deposited.

The Hare Bay klippe at the northern end of Great Northern Peninsula contains thrust slices of Maiden Point and Goose Cove Formations (Cooper, 1937) whose ages are unknown but are probably partly Cambrian. The Maiden Point consists of 6,000 feet or more of coarse greywacke sandstone and minor grey and red slate, and near the base, pillowed flows and pyroclastic rocks of andesitic and basaltic composition. Some of the volcanic rocks may be Lower Ordovician like those near Notre Dame Bay. The Goose Cove Formation consists of schist, gneiss, and amphibolite derived from andesitic and basaltic volcanic rocks that border ultramafic bodies within the klippe. The Goose Cove has apparently been thrust over the Maiden Point; their stratigraphic relationships are unknown.

In the axial part of Notre Dame Trough in northeastern Newfoundland, early Middle Cambrian trilobites of a probable Atlantic affinity have recently been discovered by M. Kay in limestone lenses within volcanic rocks of a formation hitherto believed to be Middle Ordovician. The volcanic rocks outcrop over only a few

hundred feet and may comprise a large exotic block, as suggested by outcrops of chaotic rocks nearby. If transported, however, the block has not moved far and indicates that Cambrian rocks of eugeosynclinal facies were deposited, at least locally, within the central parts of the trough in Newfoundland.

Along the eastern margin of Notre Dame Trough, in *east-central Newfoundland,* clastic rocks occupy a position in the trough comparable to the Fleur de Lys Group along the western margin. The lower unit of the Gander Lake Group (Jenness, 1963) and its probable equivalent rocks to the southwest, the Baie d'Espoir Group, form a monotonous sequence of siltstone, micaceous sandstone, slate, greywacke, and quartzite, some 10,000 feet thick or more. The lower unit is unfossiliferous and is overlain, presumably conformably, by Middle Ordovician slate and volcanic rocks of the middle unit. The base abuts the Hadrynian Love Cove Group along a contact that has been interpreted as a fault rather than an unconformity, chiefly because the Precambrian rocks are less intensely metamorphosed where the two groups are in juxtaposition.

Avalon Platform. Cambrian strata of the Avalon Platform are dominantly shale with minor sandstone and limestone, and locally, on Cape Breton Island, thick volcanic rocks. The total thickness ranges from less than 1,000 feet to a maximum of more than 4,000 feet in eastern Newfoundland. The sedimentary strata are mainly of marine shallow-water deposition and their lithologic sequence remains remarkably uniform throughout most of the region. The strata are abundantly fossiliferous in places, and the sequence of similar faunas suggests that the seas covering the platform in eastern Newfoundland, Cape Breton Island, and southern New Brunswick were continuous and that they extended into the Boston area to the southwest and also probably to the British Isles and Scandinavia. All the fauna are of the Atlantic faunal realm.

The Cambrian strata conformably overlie the basal transgressive quartzites of the Random, Morrison River, and Glen Falls Formations. Where Cambrian strata unconformably overlap Precambrian rocks, their age and thickness distribution indicate that the Lower and Middle Cambrian beds accumulated in northeast-trending troughs with the sea transgressing northwesterly. The deposition of late Middle and Upper Cambrian shales was more widespread, and the seas probably covered the entire Avalon Platform. The presence of thicker and more arenaceous beds near the western extremity of Avalon Platform at Fortune Bay, Newfoundland, suggests that the later Cambrian strata grade into the eastern deposits of Notre Dame Trough.

Cambrian strata of Avalon Platform are preserved in graben or synclinoria within a Precambrian terrane (Fig. VI-3). They are simply folded. Locally, where Cambrian strata overlap the deformed Precambrian Harbour Main Group and Holyrood granite of eastern Newfoundland, they are not folded and are broken only by normal faults.

In *eastern Newfoundland* (McCartney, 1967; Jenness, 1963; Hutchinson, 1962) the axis of maximum Cambrian deposition remained near Trinity Bay where more than 4,000 feet of shale and minor limestone was deposited. Successive Lower Cambrian formations overlapped eastward and westward across lowlands of Precambrian rocks that did not shed coarse clastic detritus. The oldest formation, the Bonavista, consists of red and green shale, minor limestone, and local basal conglomerate, succeeded



PLATE VI-2

Hadrynian Harbour Main Group, folded in the Hadrynian Avalonian Orogeny, overlain with angular unconformity by Lower Cambrian slate and limestone, folded probably in the Devonian Acadian Orogeny, Brigus South Point, Newfoundland.

by the Smith Point Formation of red and green algal limestone, the Brigus Formation of red and green shale with minor limestone interbeds and nodules, and the green shale of the early Middle Cambrian Chamberlains Brook Formation that locally contains manganese-rich beds and red shale at the base. The equivalent Youngs Cove Formation in Fortune Bay contains more siltstone, indicating the presence of a source land to the west. Overlying the Chamberlains Brook Formation is the late Middle Cambrian Manuels River Formation of grey and black shale with limestone interbeds and a few intercalated basalt flows and tuff, and the Upper Cambrian Elliott Cove Formation of grey, green, and black silty shale and siltstone. The same formational sequence in Bonavista Bay, northwest of Avalon Peninsula, is separated into a Lower and early Middle Cambrian red and green shale-bearing Adeyton Group and a late Middle Cambrian to Lower Ordovician grey and black shale-bearing Harcourt Group.

On eastern *Cape Breton Island* (Hutchinson, 1952; Weeks, 1954) Cambrian shale containing brachiopods and trilobites and a local greywacke–volcanic assemblage were deposited in a northeast-trending trough that was bounded on the west by a lowland during the Lower and Middle Cambrian. In the east part of the trough the upper Hadrynian and/or lowest Cambrian Morrison River Formation is overlain by 1,000 feet of grey, green, and red shale and siltstone of the late Lower Cambrian MacCodrum and Canoe Brook Formations. They are succeeded in the northwestern part of the trough by 1,500 feet of late Middle Cambrian dark grey and black shale and siltstone of the Trout Brook and MacLean Brook Formations, and in the southwestern part by more than 500 feet of silicic and mafic lava, pyroclastic rocks, greywacke, and shale of the late Middle Cambrian Bourinot Group. This is succeeded in turn by the Kelvin Glen Group comprising at least 2,300 feet of red and green shale and siltstone, and in the basal part red and green shale and siltstone and red and grey conglomerate with quartz and quartzo-feldspathic pebbles and red arkosic sandstone. In the west part of the trough the Middle Cambrian Bourinot Group rests with angular unconformity on the Hadrynian George River Group. In this region the Bourinot thins from 2,000 feet of entirely volcanic rocks at or near the centre of volcanism, northeastward to 600 feet and southwestward to 1,200 feet to several miles from the centre where the volcanics are interbedded with greywacke, quartzite, and shale. The overlying Middle Cambrian MacMullin Formation of grey siltstone and shale thins from about 1,000 feet to zero at the volcanic centre from both the northeast and southwest. The Upper Cambrian MacNeil Formation, several hundred feet of dark grey and black shale with rare black limestone interbeds, overlies Middle Cambrian strata in both eastern and western parts of the trough, and like the equivalent Elliott Cove Group strata of Newfoundland, shows that the sea had inundated the entire area by this time. In the western part of the trough the MacNeil Formation overlaps the MacMullin Formation to rest directly on the central high of Bourinot volcanic rocks.

Near Saint John, *southern New Brunswick*, within the St. John Group (Hayes and Howell, 1937; Alcock, 1938), about 1,000 feet of Cambrian dark grey and black shale with minor sandstone and limestone lies with structural conformity on uppermost Hadrynian and/or lowermost Cambrian quartzite of the Glen Falls Formation. The seven thin Cambrian formations have been defined largely by their age and faunal content. The oldest fossiliferous rocks, the late Lower Cambrian Hanford Brook Formation, consists of about 75 feet of grey sandstone and shale. It is overlain by about 145 feet of late Middle Cambrian black shale with thin interbeds and nodules of grey limestone, referred to in ascending order as the Fossil Brook, Porter Road, and Hastings Cove Formations. The overlying Upper Cambrian strata consist of more than 500 feet of dark grey and black shale with minor limestone and sandstone interbeds, referred to in ascending order as the Agnostus Cove, Black Shale Brook, and Narrows Formations.

Meguma Trough. In Nova Scotia the Goldenville Formation (Taylor, 1967), the lowest in the Meguma Group, consists of as much as 18,000 feet of lithic greywacke and feldspathic quartzite with minor conglomerate and slate. The strata are partly metamorphosed to biotite, greywacke, schist, and gneiss of the greenschist and almandine amphibolite facies. The Goldenville is unfossiliferous, but may be upper Cambrian as the overlying Halifax Formation contains Tremadocian graptolites. Clasts in the conglomerate are quartz, quartzite, and greywacke. Sole markings on the bases of greywacke beds suggest deposition by turbidity currents flowing northeasterly and easterly, subparallel to the present structural trend of the rocks. The source must have lain to the southeast in an area now covered by the Atlantic Ocean and the currents must have moved parallel with the axis of the trough, because thin black shale was deposited to the northwest on Avalon Platform.

St. Lawrence Platform

Lower Cambrian. Lower Cambrian strata of Anticosti Basin occur in western Newfoundland and nearby Labrador. The deposits record a stable environment of deposition. They thicken gradually southeasterly toward the present exposures of the Long Range and Indian Head Range basement complexes, indicating that these Precambrian inliers were not positive elements during the early Cambrian. Bordering Strait of Belle Isle the basal Bradore Formation of the Labrador Group (Schuchert and Dunbar, 1934) lies with angular unconformity on Precambrian crystalline rocks and consists of red to light grey arkose, orthoquartzitic sandstone, conglomerate, and shale. Columnar basalt flows are interlayered with red arkosic rocks near the base. The rocks are unfossiliferous

except for the occurrence of vertical worm tubes, *Scolithus linearis*. The formation thickens southeastward from Labrador to about 300 feet on each side of the Strait of Belle Isle. The equivalent Cloud Mountains Formation near Canada Bay is 585 feet thick. Crossbedding in sandstone in Labrador indicates southeasterly transport of detritus. The Bradore Formation is overlain conformably, or with slight disconformity, by the Forteau Formation, which consists of thin-bedded red and grey variegated shale and grey limestone containing archeocyathid reefs and sandy and oölitic beds. The Forteau Formation thickens southeasterly from 185 feet in southeastern Labrador to 386 feet in St. John Highlands of western Newfoundland and 700 feet near Canada Bay. Inland from the coast of Labrador the character of the Forteau Formation changes: the archeocyathid reefs become thinner and fewer, and the limestones gradually become more evenly bedded and more argillaceous. The Forteau Formation is overlain by orthoquartzitic sandstone, limestone, and shale of the Hawke Bay Formation, which is about 300 feet thick at the type section on Hawke Bay, western Newfoundland.

In central western Newfoundland Lower Cambrian strata of the Labrador Group locally exceed 2,000 feet in thickness. They are dominantly orthoquartzitic sandstone, limestone, and shale, but are readily divisible into the formations recognized to the north. The rocks unconformably overlie the Precambrian basement and grade southeasterly into metamorphic rocks that may be partly Precambrian. Near Port au Port in southwestern Newfoundland, 900 feet of fossiliferous Lower Cambrian strata of the Kippens Formation overlies plutonic rocks of the Indian Head Range Complex with marked unconformity (Riley, 1962). The Kippens Formation consists of shale, sandstone, and limestone, and has many lithological characteristics similar to the Forteau Formation.

Middle Cambrian. Middle Cambrian strata are known at only two localities in Anticosti Basin, both in western Newfoundland. At Canada Bay about 900 feet of limestone, dolomite, and orthoquartzitic sandstone comprise the Cloud Rapids and Treytown Pond Formations (Betz, 1939). The Cloud Rapids Formation, mainly bluish black limestone and grey shale, overlies the Forteau Formation with apparent disconformity marked by a thin pebble-conglomerate. It is conformably overlain by grey, fine-grained limestone of the Treytown Pond Formation, which in turn is succeeded by sparsely fossiliferous Lower Ordovician strata. In the seacliffs of Port au Port Peninsula about 850 feet of Middle Cambrian dolomite, shale, and

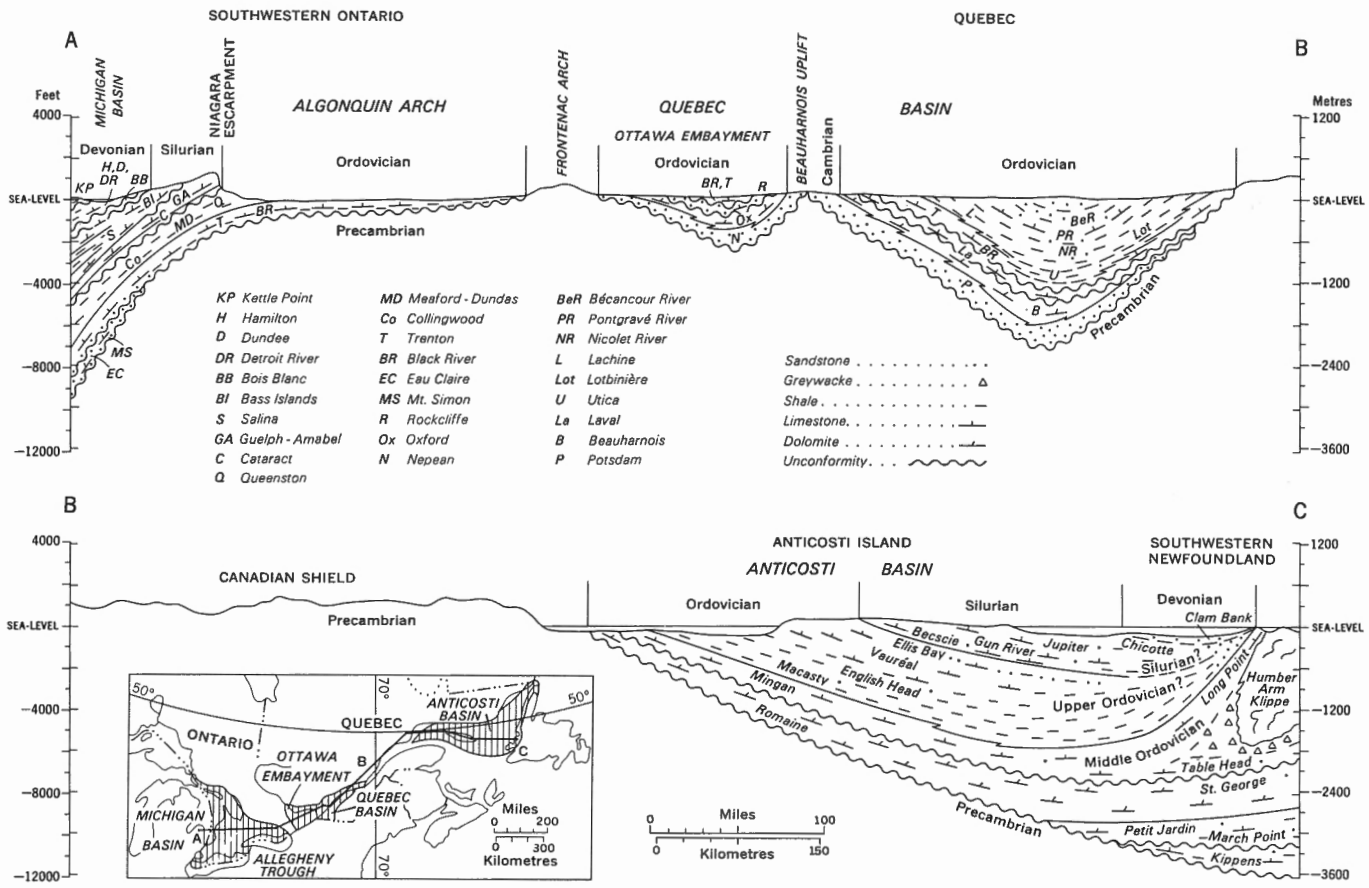


FIGURE VI-7. Cross-section of Paleozoic formations of the St. Lawrence Platform from Michigan Basin to Anticosti Basin (by B. V. Sanford).

a lower orthoquartzitic sandstone is referred to as the March Point Formation (Riley, 1962). The base is not exposed, and the Middle Cambrian beds are conformably overlain by Upper Cambrian and Lower Ordovician strata.

Middle Cambrian limestone clasts of platform facies are common in the conglomerates of the Humber Arm klippe and in those of the Quebec Group.

Upper Cambrian. Upper Cambrian strata were deposited throughout St. Lawrence Platform with the exception of Anticosti and Mingan Islands and possibly local areas in northwest Newfoundland.

In western Newfoundland, deposition in Anticosti Basin of dominantly carbonate rocks, initiated in Middle Cambrian time, probably continued in most places throughout the late Cambrian and into the early Ordovician (Fig. VI-7). Upper Cambrian strata have not been distinguished everywhere, largely because they are lithologically similar to, and grade upwards into, Lower Ordovician strata; they can only be separated faunally. Locally, about 500 feet of limestone, dolomite, and shale of the East Arm Formation near Bonne Bay and a more silty and shaly equivalent, the Petit Jardin Formation on Port au Port Peninsula, have been recognized. Faunal evidence has indicated recently that the basal part of the St. George Group, traditionally assigned to the Lower Ordovician, is Upper Cambrian.

In the Quebec Basin the basal Paleozoic transgressive sandstones are referred to as the Potsdam Formation (Clark, 1965), or locally in the Ottawa Embayment as the Nepean Formation (Wilson, 1946). They are considered correlative in part to rocks of known Upper Cambrian age exposed along the northern margins of the Adirondack Mountains in New York, although undoubtedly there are areas in eastern Ontario and Quebec where the Potsdam spans the Cambrian–Ordovician systematic boundary. The Potsdam Formation is as much as 2,000 feet thick. It is composed of a white, grey-buff, and reddish grey orthoquartzitic sandstone of exceptional maturity with quartz-feldspar pebble-conglomerate at the base or locally intercalated higher. Primary sedimentary structures point to depositional environments varying from terrestrial talus sheetwash fans bordering the areas of high relief such as the Frontenac Arch to those of low-energy lagoonal and tidal flats.

In southwestern Ontario the distribution of Upper Cambrian rocks was initially controlled by the prominent Algonquin and Frontenac Arches (Sanford and Quillian, 1959), but these tectonic elements were presumably covered by late Upper Cambrian to Lower Ordovician strata later eroded prior to the advance of Middle Ordovician seas (Fig. VI-8). During the Upper Cambrian the Algonquin Arch extended southward to connect with the Waverly Arch of west-central Ohio (Burgess, 1962). Basal Upper Cambrian deposits are white, grey, and locally red orthoquartzitic sandstone that thicken from zero at their erosionally truncated edges to 150 feet. These are termed the Potsdam southeast of Algonquin Arch, and the Mount Simon Formation to the west in Michigan Basin. The Potsdam is gradationally succeeded by the Theresa Formation, 350 feet of alternating orthoquartzitic sandstone and dolomite, and the Mount Simon by the lithologically similar Eau Claire Formation, 200 feet thick. Both formations locally overlap the older strata to rest directly on the Precambrian crystalline rocks. In the northern part of Michigan Basin the Eau Claire and Mount Simon Formations merge to form the Munising Formation consisting entirely of orthoquartzitic sandstone (Hamblin, 1958). This formation has a maximum thickness of 30 feet in the western part

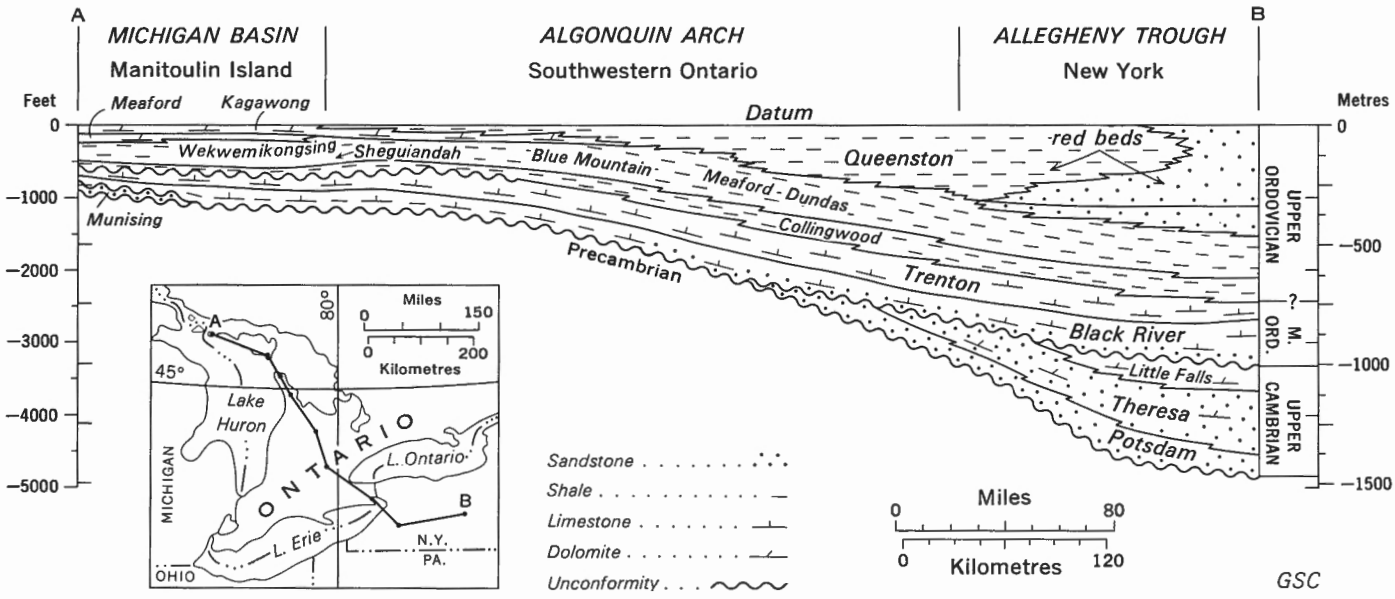


FIGURE VI-8. Cross-section of Cambrian and Ordovician formations in southwestern Ontario and northern New York (by B. V. Sanford).

of Manitoulin Island and increases to more than 200 feet in the northern peninsula of Michigan. Although detritus was presumably derived from the Precambrian crystalline rocks over which the Upper Cambrian sea transgressed, large land areas existed in northern Michigan and Wisconsin throughout Dresbachian and Franconian time, and constituted the main provenance for the considerable volume of detritus deposited within the northern and western parts of Michigan Basin.

The youngest Cambrian rocks in southwestern Ontario, the Little Falls Formation, lie beneath Lake Erie gradationally overlying the Theresa Formation. They consist of tan, finely crystalline dolomite, 100 feet thick at the International Boundary.

Ordovician

Tectonic Summary

During the Ordovician very thick, mainly eugeosynclinal sedimentary and volcanic rocks were deposited conformably on Cambrian rocks in Appalachian Geosyncline. On St. Lawrence Platform relatively thin carbonates and shale were deposited with angular unconformity on Precambrian crystalline rocks of the eastern craton or disconformably on Cambrian sandstones. In the Taconian Orogeny, during the Middle and Late Ordovician, parts of the geosyncline were folded, metamorphosed, intruded by ultramafic, mafic, and granitic bodies, uplifted, and eroded. St. Lawrence Platform was intermittently epeirogenically uplifted and eroded.

During the Early and early Middle Ordovician many thousands of feet of volcanics and greywacke accumulated in Notre Dame Trough, and thick paraplatformal orthoquartzite and shale accumulated in the central part of the trough in New Brunswick (Figs. VI-9, 10). Thin shale was deposited on Avalon Platform of the mainland, and thick, shallow-water siltstone and sandstone with interbeds of oölitic iron-formation accumulated on the platform in southeastern Newfoundland. Rather thin Lower Ordovician dolomite was deposited on St. Lawrence Platform and probably did not cover much of the shield. At the end of the Early Ordovician the platform was uplifted, and the Lower Ordovician and Cambrian rocks were eroded from Algonquin Arch and the margins of Quebec and Anticosti Basins. During the early Middle Ordovician, limestone, sandstone, and shale were deposited in Anticosti and Quebec Basins, whereas the site of Michigan Basin and Algonquin and Frontenac Arches remained land.

The early phase of the Taconian Orogeny occurred

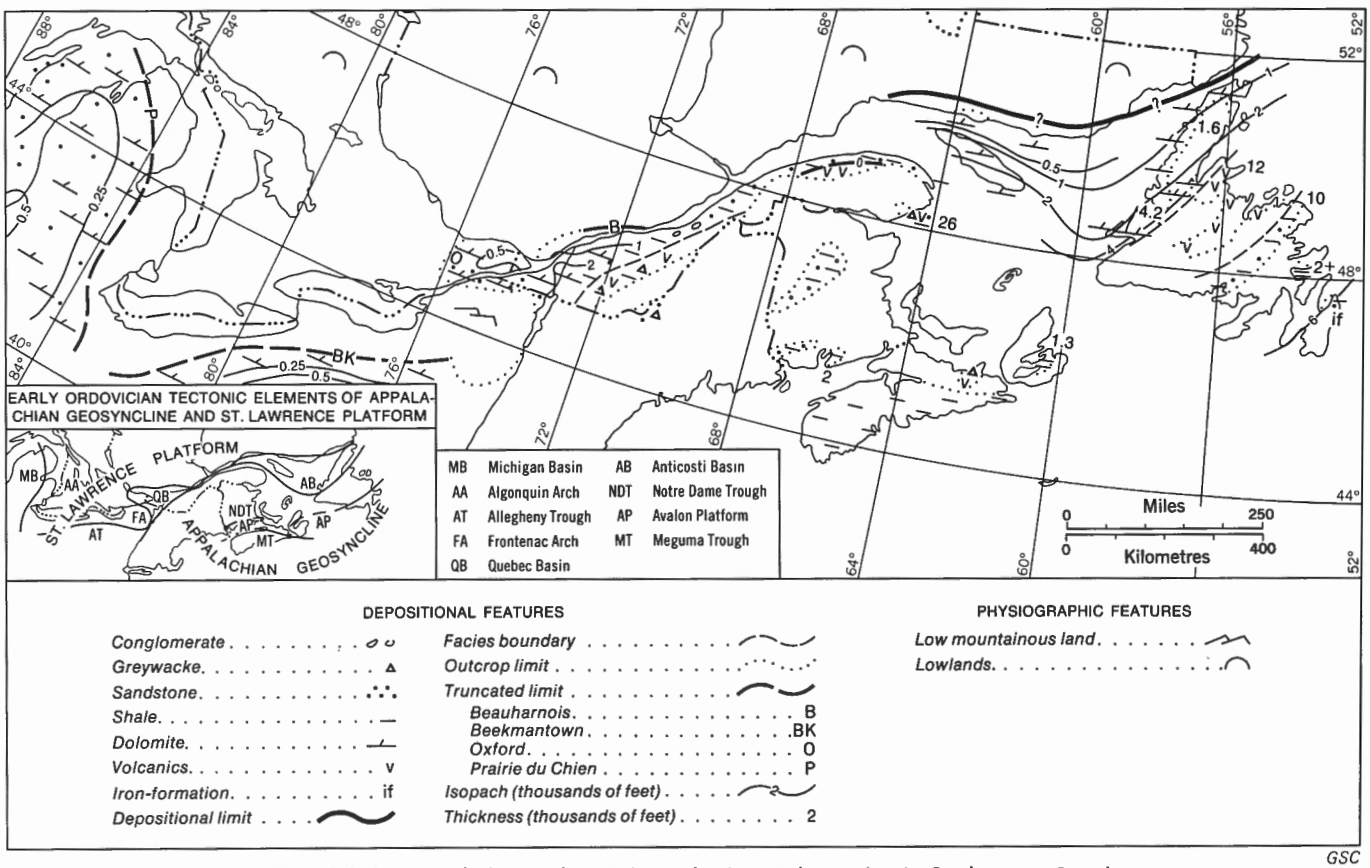


FIGURE VI-9. Lower Ordovician sedimentation, volcanism, and tectonism in Southeastern Canada.

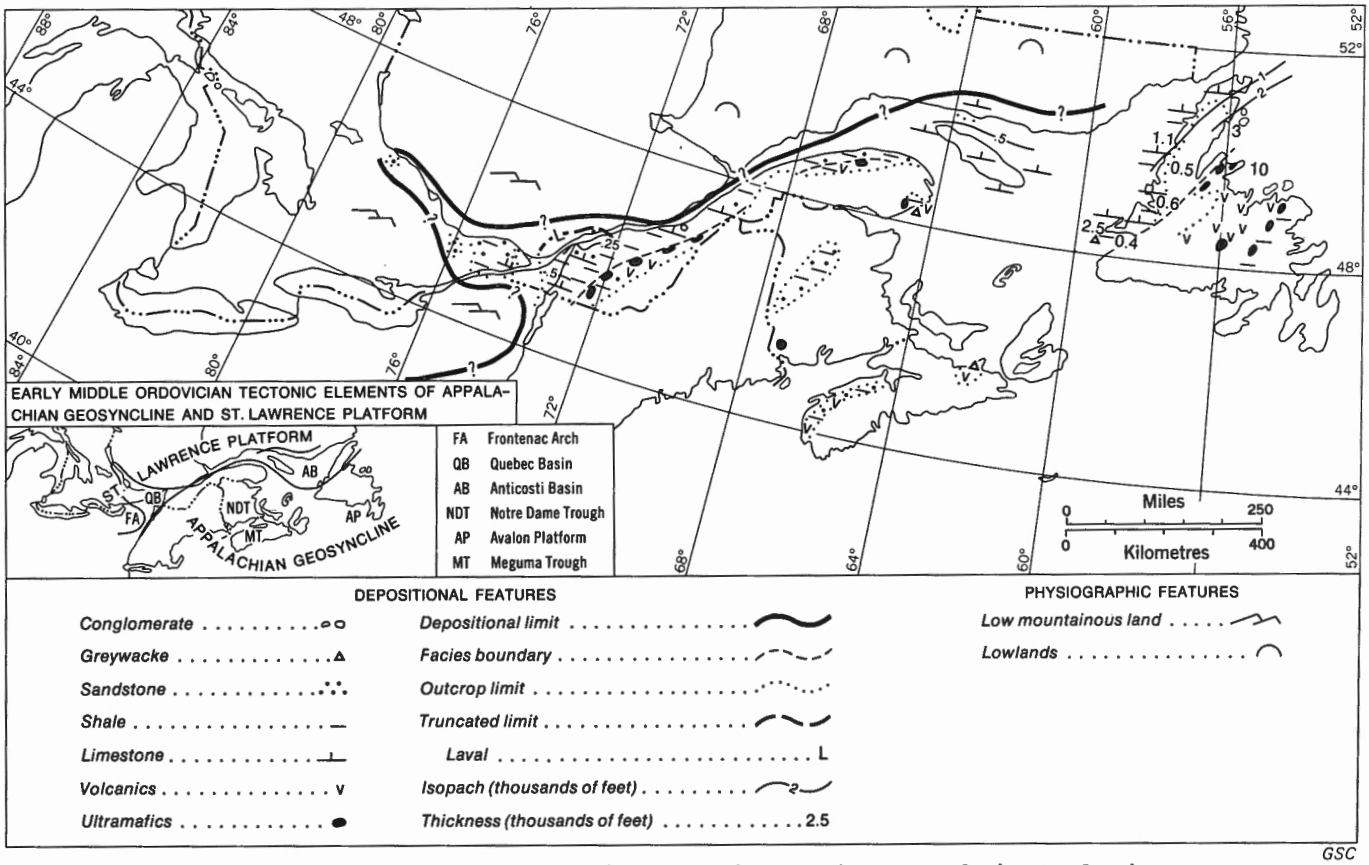


FIGURE VI-10. Early Middle Ordovician sedimentation, volcanism, and tectonism in Southeastern Canada.

during middle Middle Ordovician, about Ashby to early Wilderness (Fig. VI-11). Ultramafic bodies were emplaced in the volcanic belts of Notre Dame Trough. Quebec and Burlington Geanticlines were initiated, and folding, metamorphism, granite intrusion, uplift, and erosion occurred. Northwest of the geanticlines, adjacent to St. Lawrence Platform, a complementary trough developed in which greywacke and shale were deposited. In western Newfoundland Cambrian–Ordovician rocks and ultramafic–mafic bodies were thrust westward onto the platform. Strata of southwestern Avalon Platform were probably folded, metamorphosed, intruded by granite, uplifted, and eroded. No orogenic effects have been recognized on Avalon Platform in Newfoundland or in Meguma Trough. On St. Lawrence Platform, epeirogenic uplift coinciding with the early phase of Taconian Orogeny resulted in retreat of the sea and minor erosion.

During late Middle Ordovician thick black shale and greywacke were deposited along the northwest side and southwestern end of Quebec Geanticline. In axial Notre Dame Trough eugeosynclinal deposition was initiated in central New Brunswick and continued in central Newfoundland. Avalon Platform lacks late Middle and Upper Ordovician strata and during this time may have been a lowland of little deposition or under erosion. Eugeosynclinal deposition and volcanism also continued in Meguma Trough throughout the Middle and Late Ordovician without interruption by Taconian Orogeny. Late Middle Ordovician limestones were deposited over wide areas of the St. Lawrence Platform and these now form the oldest strata preserved in most outliers on the Canadian Shield. In Allegheny Trough and Quebec Basin limestone graded southeastward to thick shale sequences that were probably derived from Quebec Geanticline. In southwestern Newfoundland the Humber Arm klippe was partly covered by sandstone and limestone of late Middle Ordovician age.

The late phase of Taconian Orogeny, during the Late Ordovician, produced folding, metamorphism, granite intrusion, uplift, and erosion (Fig. VI-12). The older geanticlines were enlarged, and the Miramichi Geanticline was initiated. These elements subsequently controlled sedimentation until the Middle Devonian. Quebec Geanticline and the region to the northwest were subjected to folding and northwest-directed thrusting. The volume of late Ordovician detritus suggests that the southwestern part of the geanticline was uplifted more than the northeastern part. The Burlington Geanticline was probably uplifted and possibly folded at this time. Miramichi Geanticline originated within a volcanic belt; the rocks were folded, metamorphosed, intruded by granites, uplifted, and eroded. In the Matapedia facies belt, in axial Gaspé Trough between Quebec and Miramichi Geanticlines, deposition of limestone and shale continued from Middle Ordovician to early Silurian with-

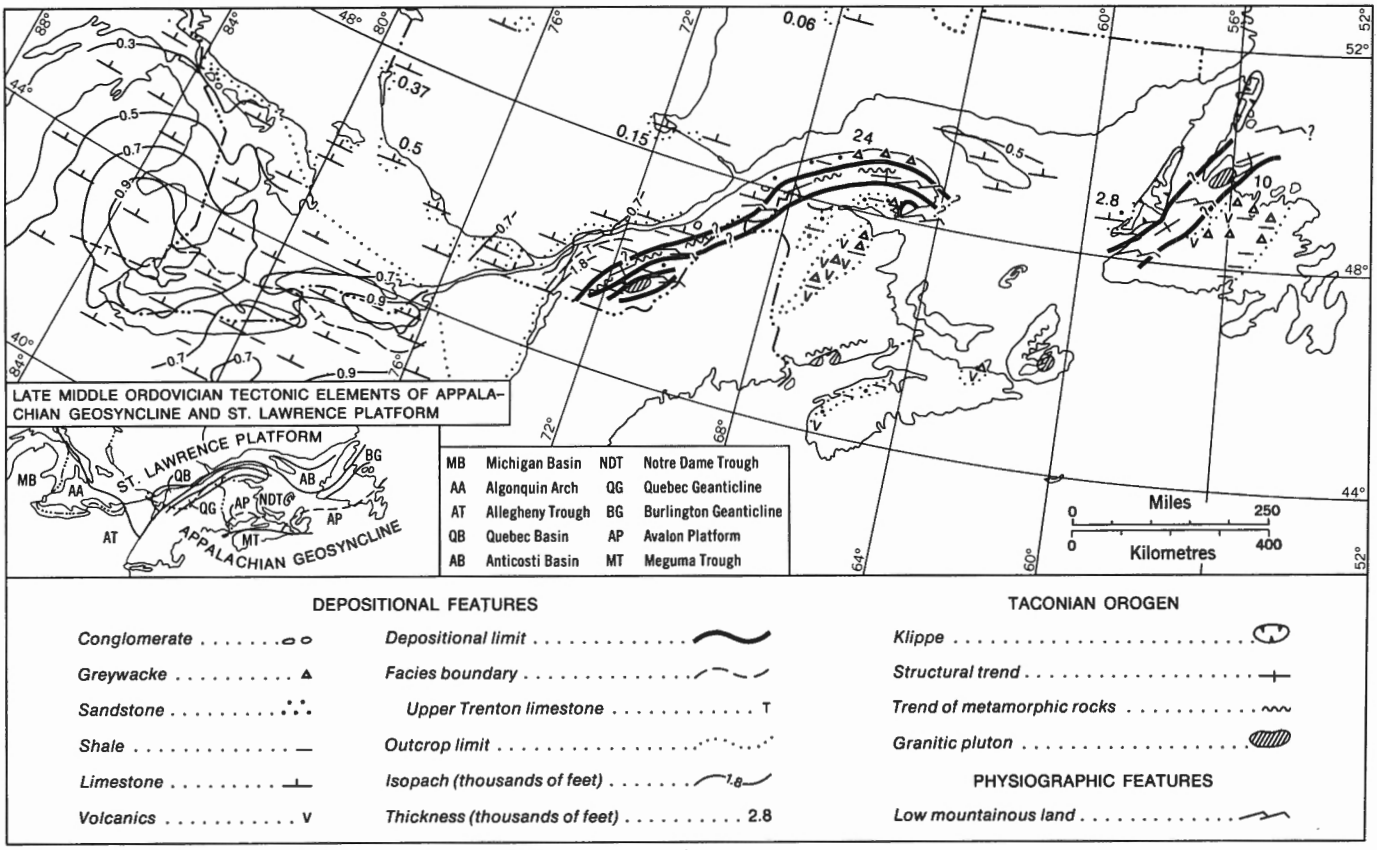


FIGURE VI-11. Late Middle Ordovician sedimentation, volcanism, and tectonism in Southeastern Canada, with some features of Taconian Orogeny produced during middle Middle Ordovician.

out reflecting the orogeny in the nearby geanticlines. Similarly, no orogenic effects of this phase have been recognized on Avalon Platform or in Meguma Trough.

Late Ordovician deposition in Appalachian Geosyncline was restricted by the growth of the geanticlines. Some sandstone and shale were deposited on Quebec Geanticline in the Eastern Townships. Deposition was continuous in the Matapedia facies belt. In Notre Dame Trough in Newfoundland volcanism gave way to deposition of greywacke and conglomerate which continued into the Silurian. On St. Lawrence Platform early Upper Ordovician black and grey marine shale was deposited in Allegheny Trough, Michigan, Quebec, and Anticosti Basins, and on the southern Canadian Shield. In latest Ordovician (Richmondian) fine to coarse clastic deltaic red beds derived from Quebec Geanticline were carried into Allegheny Trough and Quebec Basin and across Algonquin Arch into Michigan Basin. Deposition of normal marine limestones prevailed in western and northern Michigan Basin and western Anticosti Basin.

Lower and Early Middle Ordovician (Figs. VI-9, 10)

Appalachian Geosyncline. In *Notre Dame Trough,* Lower and early Middle Ordovician sedimentary and volcanic rocks have not generally been differentiated from similar, probable Cambrian rocks because fossils are sparse or lacking and deposition was apparently continuous almost everywhere. In southern Quebec the rocks can be separated into two facies belts that parallel the northwestern edge of the trough. Both the non-volcanic belt to the northwest and the volcanic belt to the southeast probably contain several thousand feet of strata. The facies boundary in the Eastern Townships corresponds roughly to the present anticlinorium, exposing older rocks of the Oak Hill, Sutton, Bennett, and Rosaire Formations; in Gaspé it falls along the north side of Shickshock Mountains. In the non-volcanic belt at Quebec City (Osborne, 1956) the Levis Formation is latest Cambrian to early Middle Ordovician and is bounded at the base and top by thrust faults. It consists of an unknown thickness of slate, siltstone, sandstone, and limestone-conglomerate. Slate and siltstone are grey, green, and red. The limestone-conglomerate occurs in lenticular beds of great lateral dimensions. About 95 per cent of the clasts are white to grey, fine-grained, massive to oölitic limestone; the remainder is dolomite and shale. The clasts vary from angular blocks and slabs to well rounded pebbles, cobbles, and boulders; the matrix is mainly carbonate. Clasts have yielded fossils of Lower, Middle, and Upper Cambrian and of Lower Ordovician age. One large Middle Cambrian block is more than 500 feet long and 40 feet thick. The source beds are not exposed; they probably lie to

the southeast beneath the thrust sheets of the folded Quebec Group.

Along the south shore of St. Lawrence River, northeast of Quebec City (Hubert, 1967), the Lower Ordovician Kamouraska Formation of orthoquartzite and limestone-conglomerate is underlain by Upper Cambrian and Lower Ordovician feldspathic sandstone, orthoquartzite, and limestone-conglomerate, and overlain by Lower Ordovician slate and siltstone. Detritus in the feldspathic sandstone was probably derived from the southeast near the anticlinorium bounding the non-volcanic facies belt on the southeast, whereas the detritus in the orthoquartzite and limestone-conglomerate was derived from the northwest. Still farther northeast, in northern Gaspé Peninsula, there is an unknown thickness of mainly Lower and some early Middle Ordovician graptolitic green-grey feldspathic sandstone and dark slate with minor grey, green, and red shale, conglomerate, limestone, and chert. Beds of orthoquartzite of the Kamouraska type and limestone-conglomerate are intercalated with Cambrian–Ordovician strata along almost the entire south shore of St. Lawrence River. Volcanic rocks are rare.

The volcanic facies belt in the Eastern Townships of southern Quebec (St-Julien, 1967) comprises unfossiliferous probable Upper Cambrian and Lower Ordovician rocks and possibly early Middle Ordovician rocks. They are unconformably overlain by late Middle Ordovician rocks, and farther southeast are blanketed by unconformably overlying Silurian and Devonian strata. In the Eastern Townships, the Caldwell Group and Brompton, Ascot, and Weedon Formations occur in separate areas; all are presumed to be about the same age. They consist of several thousand feet of silicic and mafic lava and tuff; grey, green, and red slate; and greywacke. Pillowed mafic lavas are common. Rocks of some formations have been metamorphosed to phyllite and schist. Orebodies of massive pyrite with some chalcopyrite and traces of lead and zinc sulphides are associated with schistose rhyolite and porphyry. In the southeasternmost Eastern Townships the Arnold River Formation is exposed in the core of the Boundary Mountain anticlinorium which is flanked and partly overlapped by Silurian strata. The formation is unfossiliferous but is probably Upper Cambrian and Lower Ordovician. It consists of thick, schistose feldspathic quartzite. The volcanic rocks reappear in the Shickshock Group in northern Gaspé Peninsula (Mattinson, 1964; Ollerenshaw, 1967). The group consists of several thousand feet of mafic lava, tuff, minor feldspathic sandstone, slate metamorphosed to amphibolite, amphibole schist, and mica schist. Pillow structures are common in some lavas. Stratigraphic and structural relations with the non-volcanic strata to the north are uncertain. Facies equivalence, locally complicated by faults, seems to be the most reasonable hypothesis. In south-

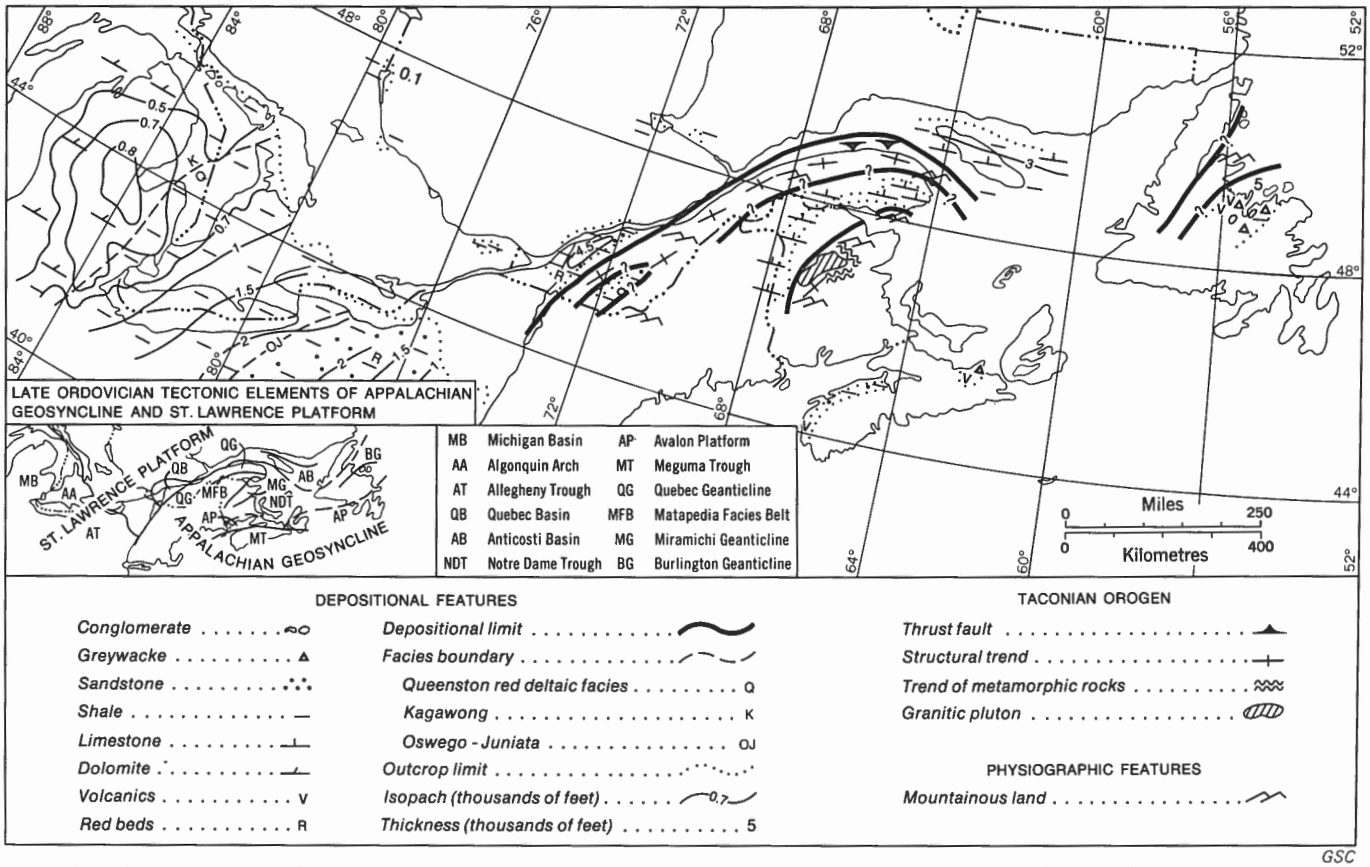


FIGURE VI-12. Upper Ordovician sedimentation, volcanism, and tectonism in Southeastern Canada, with some features of Taconian Orogeny.

eastern Gaspé Peninsula the Maquereau Group (Ayrton, 1967) is faulted against the late Middle Ordovician Mictaw Group and is unconformably overlain by the Silurian Chaleurs Bay Group. It is unfossiliferous and has been considered Lower Ordovician, Cambrian, or Precambrian. It is probably Upper Cambrian to early Middle Ordovician like the rocks south of St. Lawrence River. The Maquereau Group consists of several thousand feet of quartzose greywacke, red and green siltstone and slate, quartzite, and mafic volcanic rocks.

In central New Brunswick (Poole, 1963) several thousand feet of orthoquartzite and grey and green-brown slate and phyllite, in part metamorphosed to schist and gneiss, comprises the oldest rocks of the Miramichi Geanticline. Shelly fauna within the top 50 feet are of Lower or early Middle Ordovician age (R. B. Neuman). Similar quartzose strata in eastern Maine are considered to be Cambrian. The higher maturity of the strata indicates a more stable tectonic and shallow-water depositional environment than strata to the south or the volcanic–greywacke rocks to the north in southern Quebec. In southern New Brunswick black slate and quartzite of the Dark Argillite Division of the Charlotte Group, of unknown thickness, has yielded Arenigian graptolites.

In northern Nova Scotia, in Antigonish Highlands, the oldest rocks constitute the Browns Mountain Group (Williams, 1914). It is unfossiliferous and overlain with a disconformity or low-angle regional unconformity by the Silurian Arisaig Group. The Browns Mountain consists of 15,000 feet of silicic and mafic lava, tuff and breccia, greywacke, slate, and quartzite. Some slate is red and some felsic volcanics are ignimbrites. Hematitic quartzite and oölitic hematite beds a few feet thick occur in two places within the group. The iron-formation and some associated linguloid shells suggest correlation with Lower Ordovician hematite-bearing strata of southeastern Newfoundland.

Lower Ordovician strata in Newfoundland are autochthonous in Notre Dame Trough of central Newfoundland and allochthonous on strata of the St. Lawrence Platform, having been moved westward during the Taconian Orogeny in the middle Middle Ordovician. In Notre Dame Trough rocks of known and probable Lower Ordovician and early Middle Ordovician age cannot be separated. Those of the western trough are dominantly volcanic and have been assigned to the Baie Verte, Snooks Arm, Lushs Bight, and Wild Bight Groups, whereas those of the central trough are dominantly sedimentary and are included in the Exploits and Gander Lake Groups (Williams, 1964, 1969; Jenness, 1963). Graptolites and shelly fauna are scarce; age assignments and equivalence of strata are established generally by lithology and stratigraphic position. The Snooks Arm, Lushs Bight, and Wild Bight Groups consist of about 15,000 feet of mafic volcanic rocks with minor sedimentary rocks. The volcanic rocks are mainly massive and pillowed basalt lava, breccia, and tuff, all altered to greenstone. Less common are andesite, dacite, trachyte lava, and varicoloured pyroclastic rocks. Some basalts are spilitic having 4 to 4.5 per cent Na_2O. The sedimentary rocks are mainly greywacke, sandstone, and black, grey, green, and red slate. Chert forms beds between some basalt flows, a filling between pillows, and is interbedded with other sedimentary rocks. Limestone beds are rare and are commonly fossiliferous. The Baie Verte Group, west of Notre Dame Bay, is unfossiliferous but is lithologically similar to the lower Ordovician rocks described above. It apparently overlies schist and gneiss of Fleur de Lys Group with structural conformity. Massive chalcopyrite–pyrite bodies locally containing galena and sphalerite are associated with schistose mafic volcanic rocks. The lower part of Exploits Group contains some volcanic rocks and much slate, greywacke and minor conglomerate, chert, and limestone. Along the east side of Notre Dame Trough quartzite and grey slate within the lower member of Gander Lake Group is probably early Ordovician and has been described with the Cambrian system.

In western Newfoundland, Lower and early Middle Ordovician sedimentary and volcanic rocks form the upper parts of two klippen that rest on early Middle Ordovician limestone and black shale of the Table Head Formation. Within the Humber Arm klippe of southwestern Newfoundland (Brückner, 1966) black shale near the top of the Cooks Brook Formation contains Tremadocian graptolites. The conformably overlying Middle Arm Point Formation, 400 to 900 feet thick, consists of black and green shale with some interbeds of siltstone, dolomite breccia, quartzite, and sandstone, and has yielded Arenigian graptolites. The two conformably overlying formations are unfossiliferous, but are probably no younger than early Middle Ordovician because they were moved as part of the klippe during middle Middle Ordovician. They consist of the Blow-me-Down Brook Formation, 800 to 1,650 feet thick, of brown thick-bedded impure feldspathic sandstone with some conglomerate and interbedded red and green argillite, that is gradationally overlain by the "Humber Arm Volcanics", 1,200 to 3,000 feet thick, of mafic lava (both massive and pillowed) and pyroclastic rocks. Along the northeastern end of the Humber Arm klippe, Lower Ordovician limestone, shale, and limestone-conglomerate are believed to be part of the klippe, but unlike the klippe rocks to the south, were derived from a more stable depositional environment on the east edge of Anticosti Basin. The Green Point Formation (Kindle and Whittington, 1958) is 400 feet thick with top and bottom unexposed. It consists of grey and green-grey thin-bedded shale and limestone, minor red and green shale, and some interbeds of limestone-conglomerate (Pl. VI-3). Penecontemporaneous slumping or some diagenetic process has led to crumpling of thin limestone interbeds in shale to produce unusual folds with irregular trends that trace out polygonal outlines several inches in diameter. Anticlinal crests are bent, cracked, or broken (Pl. VI-4). Farther north in the klippe,

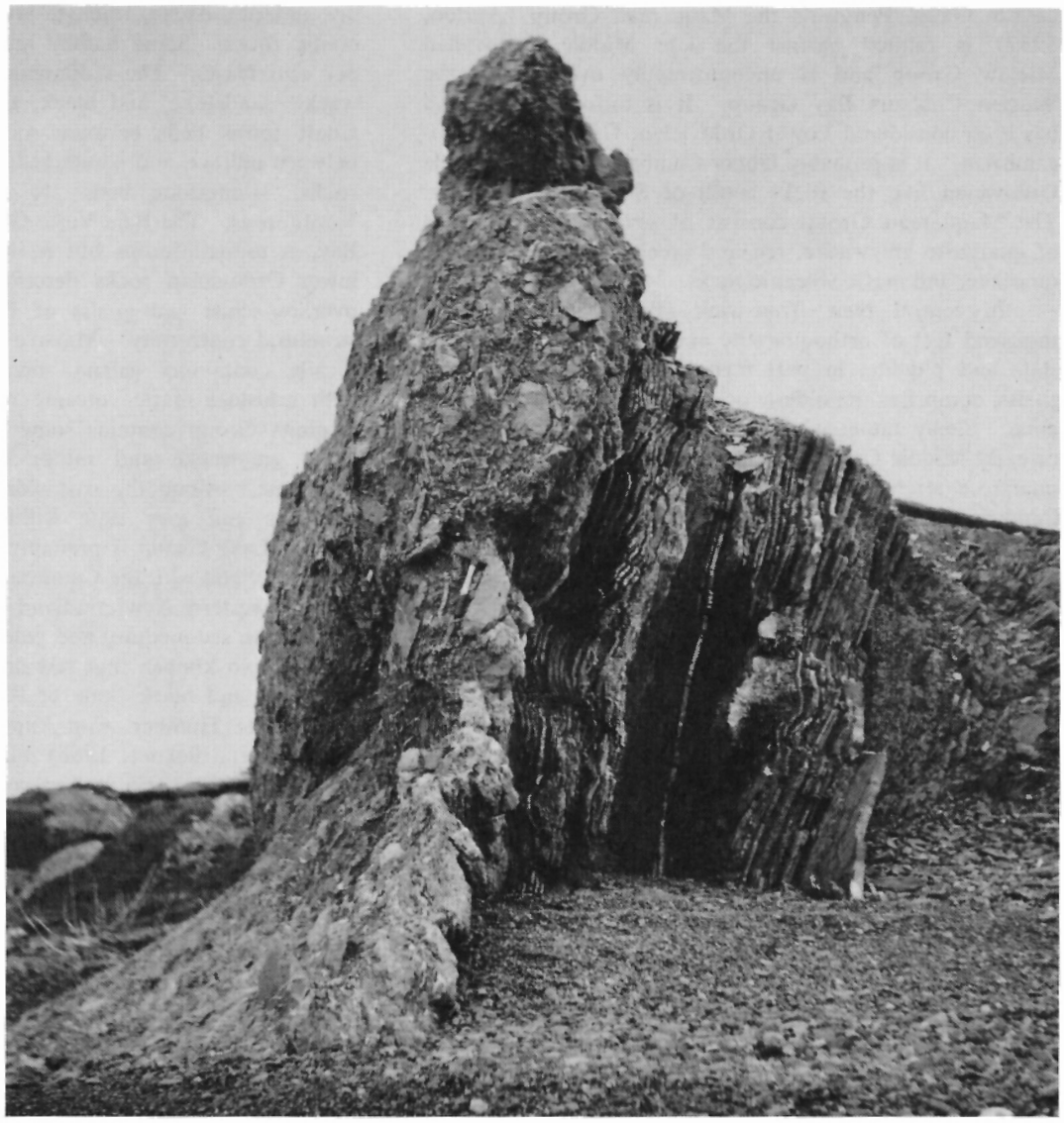


PLATE VI-3

Limestone breccia with interbedded shale and limestone of the Lower Ordovician Green Point Formation, Green Point, Newfoundland.

on Cow Head Peninsula, the upper 400 feet of the Cow Head Group consists of Lower and early Middle Ordovician limestone, limestone-conglomerate, and shale partings that are lithologically similar to the Cambrian lower part. Limestone boulders are slightly older than the strata; the boulders in the Ordovician strata range from Upper Cambrian in the lower beds to early Middle Ordovician (Whiterock) in the upper beds. The conglomerate beds, like those of the Green Point Formation, apparently originated as masses of limestone clasts that slid into areas of deposition of normal limestone. A few miles north and south of Cow Head Peninsula the Cow Head Group is overlain conformably by an unnamed formation—a few hundred feet of green sandstone and grey and red shale with Llanvirnian graptolites.

Within the Hare Bay klippe of northern Newfoundland (M. F. Tuke) the Lower Ordovician Northwest Arm Formation consists of 200 feet or more of black and green slaty argillite with interbeds of sandstone, limestone, greywacke, and conglomerate. The Northwest Arm Formation has been overthrust from the east by the Maiden Point and Goose Cove Formations that contain greywacke and volcanic rocks that are partly Lower Ordovician. The strata resemble the Canada Head Group and overlying volcanics exposed farther south in the Hare Bay klippe.

On *Avalon Platform* Lower Ordovician black shale was deposited in southern New Brunswick and Cape Breton Island; shale, sandstone, and iron-formation were deposited in southeastern Newfoundland. These strata are the youngest recognized Ordovician strata on the platform. At Saint John, New Brunswick (Hayes and Howell, 1937), the Navy Island Formation consists of 100 feet of thin-bedded black shale, with Tremadocian graptolites, that overlies similar black shale of the Upper Cambrian Narrows Formation, probably with slight dis-

conformity. The succeeding, Arenigian Suspension Bridge Formation consists of 50 feet of black graptolitic shale. In central Cape Breton Island (Hutchinson, 1952) the McLeod Brook Formation conformably overlies Upper Cambrian black shale and minor limestone of the MacNeil Formation, and consists of less than 1,300 feet of grey shale with a sparse brachiopod and graptolite fauna of Tremadocian and Arenigian age. In southeastern Newfoundland (Rose, 1952; Jenness, 1963) Tremadocian grey and black shale of the Clarenville Formation, perhaps as much as 2,000 feet thick, lies conformably on similar shales of the Upper Cambrian Elliott Cove Group on the west side of the Avalon Platform west of Trinity Bay. In Conception Bay, on the eastern side of the platform, 2,000 to 3,000 feet of Clarenville shale is believed to underlie the bay. The islands in the bay consist of Tremadocian and Arenigian strata. The older, Bell Island Group comprises an estimated 4,000 feet of grey, brown, and greenish micaceous sandstone, siltstone, and shale with minor hematitic beds. The conformably overlying Wabana Group, at least 1,000 feet thick, consists of grey, brown, and greenish shale, siltstone, micaceous sandstone, and oölitic hematite. Shallow-water deposition is indicated by crossbedding, ripple-marks, and mudcracks. Hematite beds in Wabana Group have been mined and have yielded an ore with relatively high phosphorus.

In *Meguma Trough* the Halifax Formation (Taylor, 1967) contains Tremadocian graptolites and conformably and gradationally overlies the probably Upper Cambrian Goldenville Formation. The Halifax consists of thin-bedded grey and minor greenish and black slate that thins southwesterly from more than 12,000 feet in the northeast near Wolfville, to 1,500 feet in westernmost Nova Scotia near Yarmouth.

St. Lawrence Platform. In the *Quebec Basin* (Clark, 1965; Wilson, 1946) the March Formation consists of interbedded orthoquartzitic sandstone and dolomite that form a transitional facies between the underlying orthoquartzitic sandstone and the overlying Oxford dolomite. The March varies in thickness from about 30 feet near Ottawa, to 250 feet at Montreal, and 500 feet in southeastern Quebec Basin. The Oxford Formation in the Ottawa area and its equivalent, the Beauharnois Formation in western Quebec, consist of bluish grey and brown, fine- to medium-

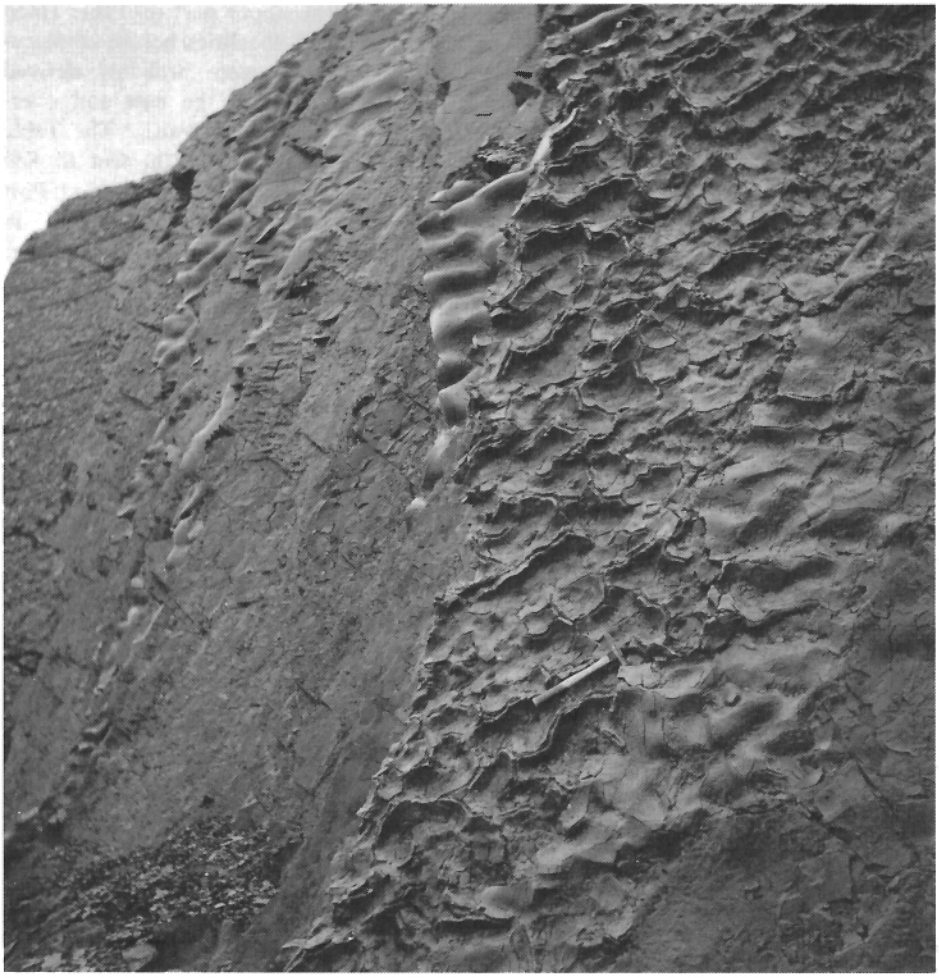


PLATE VI-4

Bedding planes of Lower Ordovician Green Point Formation showing crests of small anticlinal folds cracked and broken possibly by penecontemporaneous slumping, Green Point, Newfoundland.

crystalline dolomite that was deposited in a shallow-water near-shore environment. The upper beds consist of dark greyish brown, microcrystalline dolomite with interbeds of limestone, siltstone, and shale. They occur in the southeastern part of the basin and reflect a somewhat deeper water depositional environment. The Oxford is 410 feet thick; the Beauharnois is 800 feet thick near Montreal and thickens to 1,500 feet in the southeastern part of the basin.

The Rockcliffe and Laval Formations are of early Middle Ordovician (Marmor) age and form part of the Chazy Group (Wilson, 1946). They disconformably overlie Lower Ordovician dolomite and locally overlap its bevelled edges around the border of the basin to rest directly on Precambrian crystalline rocks. In the Ottawa Embayment, the Rockcliffe Formation, 250 feet thick, consists of olive-green and minor red shale and siltstone with lenses of coarse-grained protoquartzitic sandstone. The Laval Formation in eastern and southern Quebec Basin is more than 600 feet thick and consists of a basal unit of protoquartzitic sandstone and greenish grey shale and siltstone that grades vertically and laterally toward the basin to a thick unit of interbedded limestone, shale, and shaly dolomite. Calcarenite beds (St. Martin Member) intertongue with the Rockcliffe Formation in the eastern part of Ottawa Embayment and interdigitate throughout the middle and upper parts of the Laval Formation in Quebec Basin. Small bioherm reefs composed of coral and bryozoa colonies developed locally in the calcarenite facies at various horizons in the upper part of the Laval.

In western *Anticosti Basin,* the Lower Ordovician Romaine Formation is 260 feet thick where it unconformably overlies Precambrian crystalline rocks on the Quebec mainland (Twenhofel, 1938). It thickens southward to 1,200 feet in the subsurface beneath central Anticosti Island. The Romaine consists of greyish brown and bluish grey stromatolitic dolomite with a thin basal veneer of orthoquartzitic sandstone. On Mingan Islands the early Middle Ordovician (Marmor and Ashby) Mingan Formation rests disconformably on Lower Ordovician dolomite and consists, in ascending sequence, of as much as 39 feet of basal conglomerate, sandstone, and shale, 30 feet of bioclastic limestone and coarse calcarenite, and 96 feet of finely crystalline and aphanitic limestone. The formation thickens southward from 165 feet on Mingan Islands to about 500 feet in the subsurface of Anticosti Island.

In eastern Anticosti Basin, the St. George Group (Schuchert and Dunbar, 1934) outcrops in western Newfoundland and passes beneath the two klippen. It is primarily Lower Ordovician but locally includes some Cambrian. The group consists of thick-bedded, massive, grey- and buff-weathering dolomite with some interbedded dolomite and limestone in its upper part. It thickens southeastward from 1,600 feet along the north coast to 4,200 feet near Corner Brook. On the east side of Great Northern Peninsula the Chimney Arm Formation at Canada Bay lies disconformably on Middle Cambrian carbonates and dips beneath the Hare Bay klippe (Betz, 1939). It comprises more than 1,200 feet of black and grey dolomite and dark grey limestone with shaly interbeds. Mudcracked beds indicate a shallow-water depositional environment. West of White Bay the Doucers Formation is similar but unfossiliferous (Neale and Nash, 1963). It consists of 120 to 300 feet of fine-grained, white to blue-grey marble. The early Middle Ordovician (Whiterock to early Marmor) Table Head Formation at its type locality on the west coast of Newfoundland consists of 1,110 feet of limestone and shale that disconformably overlies Lower Ordovician dolomite (Whittington and Kindle, 1963). It comprises a lower unit of 810 feet of dark grey limestone, a middle unit of 270 feet of dark grey limestone and black shale with some contorted rubbly and slumped beds, and an upper unit of 30 feet of black shale. The formation is overlain conformably by at least 700 feet of unnamed and undated, but presumably early Middle Ordovician, grey siltstones and greenish sandstones exhibiting crossbeds and some graded beds and load casts. Similar sandstones overlie the Cow Head Group of the klippe and if they are the same formation, as seems probable, deposition there began earlier (that is during deposition of the upper part of Table Head limestone and shale) and in both localities before emplacement of the klippe. The sands were probably derived from within Notre Dame Trough to the east and overlapped progressively younger strata westward. The Table Head Formation thins southerly from 1,100 feet to 600 feet near Corner Brook and to 400 feet or more at Port au Port. The thinning is apparently southeastward, in the opposite direction to that of the underlying St. George Group. On Port au Port Peninsula the Table Head Formation is overlain by several hundred feet of limestone-conglomerate containing clasts of St. George and Table Head carbonate, which in turn is overlain by about 2,500 feet of greenish grey greywacke and shale of late Llanvirnian to early Caradocian age. The lower part of the greywacke sequence farther east on the peninsula was overridden by the Humber Arm klippe. Blocks from the klippe were incorporated in the greywacke sequence, apparently while the klippe rocks were displaced westward. In Canada Bay on northern Great Northern Peninsula, the Bide Arm and Englee Formations are largely correlative with the Table Head Formation. The Bide Arm rests on Lower Ordovician dolomite and limestone and consists of about 1,500 feet of dark grey limestone, blue-grey dolomite, and marble. The overlying Englee is about 1,500 feet of black slate with sandstone near the top.

Taconian Orogeny, Early Phase

Ultramafic intrusions. Bodies of ultramafic and mafic rocks and associated small granitic stocks and dykes occur in a belt along the western flank of Notre Dame Trough

in southern Quebec and western Newfoundland, along the eastern flank in eastern Newfoundland, as parts of two klippen in western Newfoundland, and as an isolated body in southern New Brunswick (Fig. VI-10). They were emplaced into Cambrian, Lower Ordovician, and probable early Middle Ordovician volcanic rocks during the early or middle Middle Ordovician. The associated small granitic bodies are spatially and apparently genetically related, and hence probably of the same age. The ultramafic rocks of southern Quebec contributed detritus to late Middle Ordovician strata. The bodies in the Humber Arm klippe were carried westward as part of the klippe, which after displacement was overlain by late Middle Ordovician (Wilderness) strata. The ultramafic bodies of the Hare Bay klippe and of western Notre Dame Trough are presumed to be the same age. In western Notre Dame Bay clasts of ultramafic rock and detrital chromite, from the body that intruded the volcanics of the lower Ordovician Snooks Arm Group, occur within the unconformably overlying, probably Silurian, Cape St. John Group. During later deformation in the Acadian Orogeny the ultramafic body moved upward as a solid intrusion to cut the Cape St. John Group. The bodies that intruded the Gander Lake Group of eastern Notre Dame Trough were emplaced into Lower or Middle Ordovician volcanic and sedimentary rocks, but the relationship and ages of nearby strata are not precisely known. Near the southwest end of the Gander Lake serpentinite belt, in central Newfoundland, an ultramafic body appears to have intruded the Silurian Botwood Group. Clasts of ultramafic rock do occur locally in conglomerates of the Lower and Middle Silurian Goldson Formation. The St. Stephen ultramafic body in eastern Notre Dame Trough in southwestern New Brunswick has intruded Lower Ordovician black slate and quartzite of the Charlotte Group. The intrusion probably contributed the diorite clasts found in the nearby mid-Silurian Oak Bay Formation. The small, elongate, serpentinized peridotite dyke-like body along an Acadian fault—the Rocky Brook–Millstream break in northern New Brunswick—is of uncertain age but has intruded Silurian strata with no thermal metamorphic effects. It is not known whether the emplacement of the recognized Ordovician ultramafic bodies was genetically related to the Taconian Orogeny or to volcanic processes. Most lack thermal aureoles.

The ultramafic rocks of the Eastern Townships are the host rocks for the famous chrysotile asbestos mines (Chap. VII). They consist of elongate bodies of peridotite, dunite, and minor pyroxenite and gabbro lying nearly concordant with the enclosing strata. Plugs and dykes of cataclastic granite and syenite cut the ultramafic bodies and are confined to them. Muscovite from two such bodies yielded K–Ar dates of 477 and 481 m.y., about early Middle Ordovician. The ultramafic rocks are intensely sheared and serpentinized and locally steatized and carbonatized. Soapstone and talc are mined in some localities. In contrast, the Mount Albert ultramafic body in northern Gaspé Peninsula has an ovoid surface outline and is believed to be a high-temperature mass. It is composed of dunite, peridotite, and minor pyroxenite. It has been little serpentinized and lacks granitic rocks. Gently dipping banding indicates an internal domal structure. In the thermal aureole, as much as 1,000 feet wide, greenschist of the Shickshock Group grades through amphibolite to garnetiferous hornblende–pyroxene granulite at the contact. Muscovite and biotite from garnetiferous micaceous quartzite within the aureole 700 feet from the contact yielded K–Ar dates of 495 m.y., early Ordovician. The St. Stephen body in southwestern New Brunswick consists of several square miles of gabbro and minor peridotite, and it too lacks internal granitic intrusions. Deposits of nickel and copper sulphides occur in the gabbro. Biotite from gabbro near the peridotite yielded a K–Ar date of 462 ± 20 m.y., Middle Ordovician.

In Newfoundland, in western Notre Dame Trough, peridotite and minor pyroxenite, generally serpentinized and locally steatized and carbonatized, contain chrysotile asbestos that is mined at Baie Verte. Small associated bodies of gabbro and diorite (but no granitic plugs and dykes) are probably genetically related to the ultramafic bodies, whereas other larger bodies cut Silurian rocks and are associated with Acadian granites. In the Gander River belt, in eastern Notre Dame Trough, the ultramafic rocks are generally elongate bodies of serpentinized peridotite and pyroxenite with closely associated stocks of gabbro and diorite and a few smaller bodies and dykes of albite granite. Locally the rocks are intensely sheared and altered to talc–carbonate. One such body contains 10 million tons or more of low-grade magnesite.

The ultramafic bodies in the two klippen of western Newfoundland were probably originally emplaced in the western Notre Dame Trough. In the Humber Arm klippe interbanded peridotite and dunite grade upward into banded gabbro in a body originally 60 miles long and now separated into several masses by west-directed thrust and tear faults. Layering dips gently to moderately west. The lower, ultramafic zone, 2½ to 4 miles thick, is concordant with the thermally metamorphosed underlying strata. Contact metamorphic rocks are 200 to 500 feet of biotite phyllite, amphibolite, and calcic hornfels. Gabbro in the upper part has intruded the volcanic rocks in the roof and developed an amphibolite and pyroxene–hornblende granulite contact facies. Elongate bodies, several miles long, of altered hornblende quartz diorite cut the gabbro and presumably are genetically related to and the same age as the ultramafic rocks. In the Hare Bay klippe banded serpentinized peridotite comprises a large dissected sheet, as much as 2,000 feet thick, capping several hills and underlain by greenschist and hornblende gneiss of the Goose Cove Formation. Contact metamorphic effects cannot be identified with certainty. Layering in the peridotite dips at gentle to moderate angles. Several gabbro stocks cut sedimentary rocks nearby and are probably related to the ultramafic rocks.

Deformation, metamorphism, and granitic intrusions. In southern Quebec the belt of Cambrian to possibly early Middle Ordovician volcanic rocks with its emplaced ultramafic–mafic bodies became uplifted during the first stage of development of Quebec Geanticline. Folding and regional metamorphism attributable to the early phase of Taconian Orogeny cannot be distinguished with certainty, although they are probably present. Early folds in pre-Middle Ordovician rocks in Sutton Mountains of the Eastern Townships are isoclinal and recumbent; they trend northwesterly to westerly in contrast with the later northeasterly trending structures. The rocks are in the greenschist facies of metamorphism. Muscovite yielded K–Ar dates ranging from 440 to 414 m.y., about Middle Ordovician to Middle Silurian. In northern Gaspé Peninsula the Shickshock Group is believed to have been regionally metamorphosed before the intrusion of the Mount Albert ultramafic body with its thermal aureole. The Shickshock grades southerly from phyllite and greenstone into biotite schist, and in one locality to garnetiferous schist and blue-green amphibole schist. The metamorphism and presumed associated recumbent and isoclinal folding may be Taconian effects, although muscovite from phyllite has yielded a K–Ar date of 530 m.y., about Middle Cambrian. In southeastern Gaspé Peninsula the undated phyllite of the Maquereau Group is complexly deformed into recumbent folds with subhorizontal axial plane cleavage that certainly developed before the unconformably overlying Silurian and probably before the late Middle Ordovician.

The Moulton Hill intrusion in the Eastern Townships is a coarse-grained granite stock that was later cataclastically deformed and altered to essentially quartz and albite. Clasts of volcanic rocks of the Ascot Formation, ultramafic rocks, and the Moulton Hill granite are contained in the unconformably overlying late Middle Ordovician Magog Formation. In northern Gaspé Peninsula late Middle Ordovician strata lying north of the uplifted belt contain detrital serpentine, chromite, and volcanic rocks probably derived from the Shickshock Group and ultramafic intrusions like the Mount Albert body. Relations of late Middle Ordovician strata to older rocks are uncertain and are obscured by younger faults.

Along the northwest side of Quebec Geanticline a trough developed in which was deposited a thick sequence of shale and greywacke of middle and late Middle Ordovician age. Such strata are a few thousand feet thick within the Notre Dame Trough of most of southern Quebec and as much as 24,000 feet thick in northern Gaspé Peninsula. Equivalent strata, if deposited on St. Lawrence Platform, have been either removed by erosion or overridden by the upper block of Logan's thrust. The southeast-thickening black shale in southeastern Quebec Basin is probably a relic of this sequence.

In western Newfoundland Burlington Geanticline was probably uplifted at this time, but stratigraphic evidence is lacking. The postulated uplift is a necessary part of the hypothesis in which the two Taconian klippen were thrust westward from this region onto the St. Lawrence Platform. Ages of folding, metamorphism, and granite intrusion are difficult to establish. The Fleur de Lys Group of the northern geanticline has been metamorphosed to greenschist and almandine amphibolite facies and contains a complex of superimposed folds, the oldest of which are recumbent and isoclinal. K–Ar dates on micas in the schists and gneisses record Devonian ages, probably as a result of degassing by nearby Devonian granite intrusions to the southwest. South of Corner Brook biotite from gneiss and associated granite yielded dates of 452 and 420 m.y., about early Middle Ordovician to Middle Silurian. The Burlington granodiorite in northern Burlington Geanticline is unconformably overlain by volcanics of probable Silurian age and may be a Taconian intrusion. It is massive, medium to coarse grained, and partly altered to epidote and chlorite.

West of Burlington Geanticline, during late Llanvirnian to earliest Caradocian time, thick greywacke and shale were deposited on the platformal carbonate of the Table Head Formation as the waters deepened and a trough developed. The klippen overrode part of this sequence; the Humber Arm klippe and probably the unnamed greywacke–shale sequence in front of the klippe are overlain by the middle Caradocian (late Wilderness) Long Point Formation.

The klippen either slid or were thrust westward from western Notre Dame Trough and the zone along the trough–platform boundary into the trough of greywacke–shale deposition, a distance of perhaps 25 to 60 miles. The time of displacement is early Caradocian (Porterfield and early Wilderness) as determined by the age of the overridden strata, age of the greywacke–shale with incorporated blocks, and in the Humber Arm klippe, age of the overlying, post-displacement Long Point Formation. The time of displacement of the Taconic klippe of New York and Vermont is considered to be later, in middle or late Caradocian (Barneveld). Rocks of the klippen are moderately to intensely folded, cleaved, and faulted (Pl. VI-5). Folds and cleavage parallel regional northeast trends; faults trend northeast to northwest. Near Corner Brook the southeast part of the Humber Arm klippe has southeasterly overturned folds associated with a southeast-directed thrust. The intensity of recrystallization within muscovite–chlorite subfacies decreases westward as the axial planes become vertical. In the Hare Bay klippe the Maiden Point Formation has recumbent folds. The Goose Cove basic schist and gneiss are in the greenschist and low almandine amphibolite facies. It has been suggested that these rocks were derived in their metamorphic form from some parts such as the Birchy schist of the Fleur de Lys Group in northern extension of Burlington Geanticline that may underlie the ocean east of the klippe.

On Avalon Platform, in Cape Breton Island and southern New Brunswick, Middle and Upper Ordovician

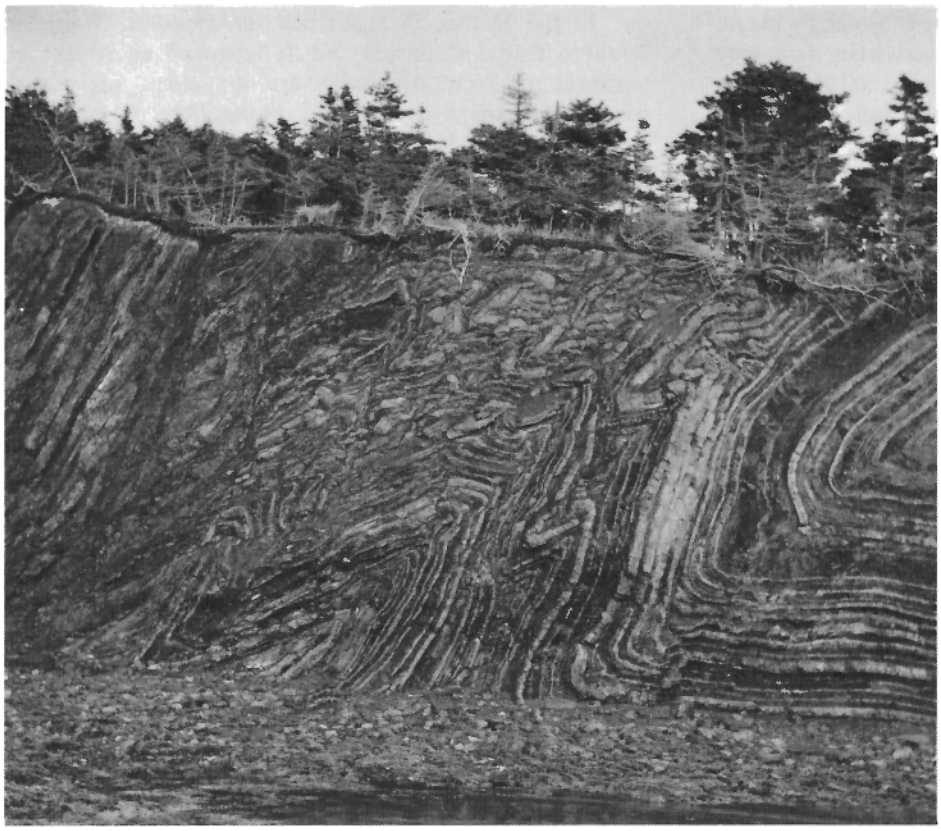


PLATE VI-5
Contorted Ordovician shale near base of Humber Arm klippe, Black Point, Port au Port, Newfoundland. Folds possibly developed during movement of klippe in the Ordovician, Taconian Orogeny.

strata are lacking, and relations of Silurian strata are obscured by Acadian deformation, except in southeastern Cape Breton Island where the undated but probably Silurian or Devonian Middle River Formation rests with angular unconformity on Middle Cambrian rocks. In central Cape Breton Island the Precambrian George River Group and its possible equivalents in northern Cape Breton Island were metamorphosed to marble and schist, possibly during the early phase of the Taconian Orogeny and probably during Acadian Orogeny. Similarly, in southern New Brunswick, parts of the Precambrian Green Head Group were metamorphosed to marble, schist, and biotite gneiss, and parts of the Coldbrook Group to phyllite, sericite schist, and amphibolite. A K–Ar date on biotite from Green Head biotite gneiss yielded 508 m.y., about late Cambrian.

Granites have intruded the metamorphic belt and are characterized by a wide variety of composition in contrast with the relatively uniform composition of the later Acadian intrusions. They range from gabbro through diorite, quartz diorite, granodiorite, syenodiorite, monzonite, and syenite to granite. Some are massive and coarse grained, others are cataclastic and foliated. Most are chloritized. They form the Milkish Head and Golden Grove bodies near Saint John, New Brunswick, and the bodies in the Boisdale and Coxheath Hills of central Cape Breton Island. The last two bodies contain associated copper and molybdenum deposits. The small lenticular anorthosite and diorite plutons of northern Cape Breton Island may belong to this suite. Similarly, some of the mixed granites in Antigonish Highlands of Nova Scotia may be Taconian intrusions.

K–Ar dates from biotite in granite and gabbro in southern New Brunswick are 479 and 500 m.y. Those from biotite and hornblende in two granites in Antigonish Highlands are 432 and 582 m.y., and those from biotite in granite and hornblende in syenodiorite in central Cape Breton Island are 493 and 584 m.y. Most dates suggest intrusion during Cambrian and early Ordovician, but they are interpreted as probably reflecting Taconian events, because in central Cape Breton Island the intrusions cut Middle Cambrian rocks that are succeeded conformably by Lower Ordovician shale. They could, however, reflect an earlier Taconian event in the Lower Ordovician.

Late Middle Ordovician (Fig. VI-11)

Appalachian Geosyncline. Late Middle Ordovician (Wilderness and Barneveld) strata are common in Notre Dame Trough, are probably present in Meguma Trough, but are probably lacking on Avalon Platform. In Notre Dame Trough of southern Quebec they lie in a miogeosynclinal trough northwest of the Quebec Geanticline and on the southeast side of the geanticline. Along the northwest side, near Lake Champlain (Cady, 1960), the Mystic

Formation consists of 75 feet of limestone-conglomerate that rests unconformably on Lower Ordovician dark grey shale and limestone. It contains blocks several feet across of Lower Ordovician limestone derived from as much as 1,350 feet stratigraphically below the unconformity. The overlying late Middle Ordovician Stanbridge Formation consists of about 2,000 feet of dark grey and black slate and minor siltstone. In the same belt, 15 miles south of Quebec City, are black sandstone and chert with pyritic black slate containing upper Porterfield and Wilderness graptolites. At Quebec City the youngest strata immediately southeast of Logan's Line are the Middle and possibly Upper Ordovician Citadel and Quebec City Formations of the Quebec Group (Osborne, 1956). The Citadel consists of an unknown thickness of shale and limestone-conglomerate. Clasts are dark grey limestone bearing Chazy and Black River fauna, and hence the formation is probably of Wilderness and Barneveld ages. The conformably overlying Quebec City Formation consists of more than 2,000 feet of thick-bedded, dark grey argillaceous limestone and shale. The age of the formation is probably Barneveld to Maysville. Near Lake Témiscouata the Trinité Group consists of an unknown thickness (probably more than 1,000 feet) of ribbon limestone, varicoloured slate, quartzite, limestone-conglomerate, and black slate that has yielded Wilderness graptolites (Lajoie, *et al.*, 1968). Along the northern Gaspé Peninsula coast for 100 miles are dark grey argillite and interbedded greywacke and calcisiltite containing Barneveld graptolites; a thickness of about 24,000 feet has been suggested. The rocks are characteristic of a flysch and turbidite facies as grading, scour marks, and flute casts are common. Sole marks at the base of greywacke beds indicate current flow toward the west, probably parallel to the trough axis. Detritus of serpentine, chromite, volcanic rocks, and chert indicate derivation from the Shickshock Group in Quebec Geanticline to the south.

Along the southeast side of Quebec Geanticline are late Middle Ordovician black slate, greywacke, and minor volcanic rocks similar in lithology and general thickness to those on the northwest side. In the Eastern Townships, the Beauceville and Magog Formations consist of dark grey slate, greywacke, sandstone, minor conglomerate, and chert; mafic and silicic volcanic rocks are common in some localities (St-Julien, 1967). Clasts in conglomerate and sandy rocks are feldspar, quartz, shale, volcanic rocks, and albite granite, all similar to nearby older rocks. Graptolites are of Wilderness and Barneveld ages. The basal contact is probably either a disconformity or low-angle unconformity. Thicknesses of the formations are unknown, but probably are several thousand feet.

In Gaspé Peninsula hard black shale lying south of Shickshock Group rocks of biotite metamorphic grade has yielded Normanskill graptolites. The strata probably form a horst within the Shickshock normal fault which has downdropped Silurian strata on the south.

In the Matapedia facies belt the Honorat Group in southern Gaspé Peninsula and its unnamed equivalent in northwestern New Brunswick are the oldest rocks and consist of perhaps as much as 14,000 feet of grey and greenish grey slaty argillite, siltstone, quartzite, quartzose greywacke, and grit (Skidmore, 1965). Some arenaceous beds are feldspathic and some are rich in volcanic detritus. The formations have yielded Middle or Upper Ordovician graptolites and shell fragments. Near the southeast coast of Gaspé Peninsula the Mictaw Formation contains Barneveld graptolites and is lithologically similar to the Honorat Group, but is separated from the Honorat and Matapedia Groups by Silurian strata (Ayrton, 1967). The Mictaw comprises more than 3,500 feet of dark greenish grey volcanic-rich sandstone, greywacke, siltstone, and slate of which the basal 1,000 feet is made up of conglomerate and minor interbeds of red siltstone. Clasts in the conglomerate are angular, generally from 3 to 6 inches in diameter but with boulders up to 1½ feet in diameter, and include greywacke, feldspathic quartzite, greenstone, quartz, and locally pink granite gneiss, all probably derived from Maquereau Group and associated rocks. Detrital muscovite from Mictaw sandstone and from unnamed quartzite equivalent to the Honorat Group in northwestern New Brunswick yielded K–Ar dates of 563 and 505 m.y. respectively, that is, Cambrian. The conglomerate has recently been interpreted as a talus deposit rather than a basal conglomerate (Ayrton, 1967).

The Matapedia Group conformably and gradationally overlies the Honorat Group and is at least 9,500 feet thick (Skidmore, 1965). It consists of dark blue-grey, dense argillaceous limestone in beds as thick as 3 inches interbedded with thinner beds and partings of grey shale and siltstone that produce a distinctive ribboned appearance. In southern Gaspé Peninsula the Matapedia Group comprises a more shaly Pabos Formation overlain by the White Head Formation with relatively more limestone. The Matapedia Group in easternmost Gaspé Peninsula overlies the Upper Cambrian Murphy Creek Formation with apparent angular unconformity. Graptolites date the Matapedia Group and its equivalents as late Middle Ordovician (Barneveld) to earliest Silurian in Maine and late Upper Ordovician to early Silurian in Gaspé Peninsula. The lower and upper contacts may be diachronous.

A eugeosynclinal assemblage of late Middle Ordovician and possibly younger rocks occurs in a belt extending southwesterly through central New Brunswick. In northeastern New Brunswick the Tetagouche Group consists of many thousands of feet of silicic and mafic flows, pyroclastic rocks and sills, greywacke, partly graphitic slate, and minor cherty iron-formation (Davies, 1966). A variety of quartz- and feldspar-bearing rocks of uncertain origin are variously referred to as "porphyry", quartz–feldspar– and quartz–augen schist and gneiss, and quartz–sericite–chlorite schist and gneiss. These rocks form tabular bodies, several miles long and a few thousand feet

thick, lying structurally concordant with bounding strata. More than a dozen bodies of massive base metal sulphides totalling over 125 million tons have been found in the Bathurst mining camp. The bodies are generally concordant with Tetagouche strata and are commonly located at or near augen schist. In central New Brunswick, north of Fredericton, similar strata, but with only minor silicic volcanics and without porphyry, rest conformably on unnamed Cambrian (?) and Lower Ordovician quartzite and slate (Poole, 1963). Andesite flows, tuff, and sills are associated with graphitic ribbon chert containing Wilderness graptolites and manganiferous iron-formation in the lower part of the assemblage. The upper part is mainly unfossiliferous greywacke and slate.

Late Middle Ordovician strata have not been recognized in southern New Brunswick. Part of the Browns Mountain Group of northern mainland Nova Scotia is probably of this age.

In Newfoundland, late Middle Ordovician rocks occur within the axial Notre Dame Trough in two belts extending southwesterly from Notre Dame Bay and separated by Silurian strata. The rocks apparently overlie older rocks conformably and pass upward in some places into Upper Ordovician rocks. In southwestern Notre Dame Bay about 8,500 feet of sedimentary rocks and minor mafic volcanic rocks are probably mainly late Middle Ordovician, although the upper 4,000 feet of strata may be faulted against the lower 4,500 feet (Espenshade, 1937). Both sequences are of the same general age. The lower sequence comprises three conformable formations. The Beaver Bight Formation rests conformably on the Wild Bight Formation of mainly mafic volcanic rocks, and consists of about 1,000 feet of thin-bedded tuffaceous greywacke, chert, and sandstone. Crossbedding, grading, and current scour structures are common. The overlying Shoal Arm Formation, about 1,500 feet thick, consists of interbedded grey sandstone and black shale and slate, and in the lowest 300 feet there are red, green, and black cherty shale and slate, some of which weathers to a black manganese oxide. Gull Island Formation consists of about 2,000 feet of massive brown-grey coarse-grained greywacke and minor conglomerate containing subangular detrital quartz, feldspar, hornblende, and epidote. Graded bedding is common. In the upper sequence the Julies Harbour Group comprises roughly 2,000 feet of a heterogeneous assemblage of: grey and black argillaceous sandstone and quartzite; greywacke; coarse conglomerate containing pebbles of granite, andesite, limestone and chert; chert; andesite; and a red and green cherty shale, which weathers to black manganese oxide, identical with beds of the Shoal Arm Formation. The overlying Burtons Head Group is also a heterogeneous assemblage about 2,000 feet thick of rocks much like those of Julies Harbour but containing more andesite.

On New World Island, in Bay of Exploits, the middle part of the Exploits Group contains late Middle Ordovician rocks whose lithology and stratigraphy vary from place to place (Kay, 1966, 1967). Argillite and conglomerate containing pebbles and boulders of greywacke, volcanics, and plutonic rocks are overlain by black argillite and chert with late Middle Ordovician graptolites. In another part of the island tuff is overlain by 100 feet of limestone that is overlain by 100 feet of black cherty argillite with Barneveld graptolites. In still another part, interbedded mafic lava, tuff, and cherty argillite have yielded late Middle Ordovician graptolites.

Southwest of eastern Notre Dame Bay the middle and upper units of the Gander Lake Group conformably overlie the lower unit that is probably early Middle Ordovician (Jenness, 1963). The middle unit consists of many thousands of feet of mafic pyroclastic rocks and flows; minor silicic volcanics; grey, green, black, and red slate; greywacke; and conglomerate. Ultramafic bodies intrude the volcanic rocks. The conglomerate contains pebbles and cobbles of chert, slate, volcanic rocks, quartz, granite, and limestone. The uncertain relationships with known Ordovician rocks and the lithology of the pebbles suggest that the conglomerate may be infolded Silurian. The upper unit comprises many thousands of feet of grey, green, and minor red and black slate with some greywacke, conglomerate, and mafic pyroclastic rocks. Conglomerate contains pebbles of quartz, chert, slate, and volcanic rocks. The middle unit has yielded Wilderness graptolites, and the upper unit Wilderness to early Upper Ordovician graptolites. Thus the Gander Lake Group may represent deposition throughout the entire Ordovician. The upper part of the Baie d'Espoir Group on the south coast of Newfoundland consists of about 20,000 feet of grey slate, phyllite, siltstone, and minor quartzite and greywacke, not unlike the middle and upper units of the Gander Lake Group.

Late Middle Ordovician strata have not been recognized on the Avalon Platform despite their occurrence in Notre Dame Trough and probably in Meguma Trough. The platform may have been a lowland separating the troughs, the site of little deposition or erosion. In Meguma Trough, late Middle Ordovician strata are probably represented within the thick White Rock Formation of sedimentary and volcanic rocks that are post-Lower Ordovician to pre-Upper Silurian.

St. Lawrence Platform. In southwestern Ontario (Sanford, 1961; Beards, 1967), late Middle Ordovician strata lie on Precambrian crystalline rocks on Algonquin Arch, and on the Upper Cambrian and Lower Ordovician in Michigan Basin and Allegheny Trough (Fig. VI-8). The Black River and Trenton Groups thicken from 700 feet on Algonquin Arch, to 900 feet in Michigan Basin, and to more than 900 feet in Allegheny Trough. The Black River Group of Wilderness age comprises three formations. (1) A lower transgressive unit, the Shadow Lake Formation, rests unconformably on Precambrian or Upper Cambrian rocks and consists of 20 to 30 feet of red and green shale, siltstone, minor limestone, and dolomite. In

the region of Manitoulin Island, similar strata are 2 feet to more than 40 feet thick. (2) The conformably overlying Gull River Formation consists of grey and cream-coloured lithographic limestone with interbeds of pelletoidal limestone and dolomite containing thin beds of bentonite in its upper part. From 75 feet over the Algonquin Arch the Gull River thickens to 130 feet on Manitoulin Island and to more than 400 feet beneath Lake Erie and southern Lake Huron. (3) The youngest formation of the Black River Group is the Coboconk Formation, consisting of interbedded lithographic limestone, calcisiltite, and calcarenite that thickens westward from 20 feet on Algonquin Arch to 100 feet near southern Lake Huron. Equivalent and lithologically similar beds on Manitoulin Island have a maximum thickness of about 25 feet.

The coarse clastic limestones of the Trenton Group are of late Wilderness and Barneveld age, and conformably overlie the Black River Group. A lower unit, the Kirkfield Formation, consists of interbedded finely crystalline shaly limestone, bioclastic limestone, and calcarenite. The Kirkfield thickens southeastward from 60 feet on Manitoulin Island to a maximum of 240 feet along the north shore of Lake Ontario. The succeeding Verulum Formation, comprising coarse bioclastic limestone and calcarenite with shale interbeds, increases in thickness from 80 feet on Manitoulin Island to a maximum of 140 feet beneath Lake Erie. The coarse limestones of the Kirkfield and Verulum Formations of shallow marine origin comprise the dominant facies of the Trenton Group throughout most of southwestern Ontario, but become fine textured and more shaly bordering Allegheny Trough. The Cobourg Formation is the highest of the group; it is made up of dark brown argillaceous and aphanitic limestones. From a maximum thickness of 200 feet in a narrow belt extending from the southern tip of Lake Huron to western Lake Ontario, the Cobourg thins northwest to 95 feet on Manitoulin Island and southeast to 40 feet beneath Lake Erie where it grades to black euxinic shale.

In Quebec Basin (Belyea, 1952; Clark, 1965; Wilson, 1946) the Black River and Trenton Groups vary in thickness from 730 feet in the Ottawa Embayment, to 700 feet north of St. Lawrence River east of Montreal, and to as much as 2,000 feet in southeastern Quebec Basin. The upper Trenton limestones in the western and northeastern parts of the basin grade southeastward into interbedded shale and limestone, a reflection of rapidly increasing subsidence of the Notre Dame Trough along the stable margin of the St. Lawrence Platform. The Black River Group comprises three conformable formations. (1) In western Quebec Basin a lower unit, the Pamelia Formation, overlies the Rockcliffe Formation. It consists of 70 feet of grey to black sandy shale grading upward to interbedded brown dolomite, lithographic limestone, and minor shale. The Pamelia thins eastward and may be absent in eastern and southern Quebec Basin. (2) The overlying Lowville Formation comprises dark greyish brown lithographic limestone that thins eastward from 155 feet in Ottawa Embayment to 40 feet in southern part of the Quebec Basin where it disconformably overlies the early Middle Ordovician Laval Formation. In the northeastern part of the basin it locally overlaps the Laval Formation and pre-Middle Ordovician strata and rests on Precambrian crystalline rocks. (3) The upper unit of the Black River Group is the Leray Formation, consisting of interbedded sublithographic limestone, calcisiltite, and calcarenite. It thins eastward from 65 feet in Ottawa Embayment to 21 feet near Montreal.

Formations of the Trenton Group in western and northwestern Quebec Basin are lithologically similar to those of southwestern Ontario, but somewhat unlike those of eastern and southern Quebec Basin where deposition occurred in a more rapidly subsiding environment. The Rockland Formation comprises the oldest strata of the Trenton Group in the Ottawa Embayment. It is 55 feet thick and consists of grey, argillaceous, finely crystalline limestone. The formation thins eastward to 19 feet near Montreal where it consists of dark grey, microcrystalline limestone termed the Ouareau Formation. Equivalent strata in the northeastern part of the Quebec Basin make up the Pont Rouge Formation, which is 32 feet thick. The overlying Hull Formation in Ottawa Embayment consists of 180 feet of dark blue-grey, finely crystalline limestone with shaly partings overlain by clean, coarse-grained calcarenite. Equivalent strata near Montreal are bioclastic limestone and coarse calcarenite, 300 feet thick, of the Deschambault Formation. Toward the northeast the upper 180 to 190 feet of the Deschambault gives place to a variety of limestone facies referred to as the St. Casimir Formation. In the Ottawa Embayment the succeeding Sherman Fall Formation consists of 25 feet of interbedded calcarenitic limestone and shale that presumably is equivalent to the Montreal Formation, 375 feet thick, composed of fine to coarse crystalline calcarenite and shale. Limestones of the Montreal Formation are fine textured toward the northeastern margin of the basin and merge with aphanitic limestones of the Neuville Formation, 470 feet thick. The youngest Trenton unit in Ottawa Embayment, the Cobourg Formation, consists of 180 feet of dark brown, aphanitic limestone that thickens to 500 feet near Montreal where it grades to interbedded dense limestone and shale of the Tetreauville Formation. The Tetreauville thins northeastward and ultimately merges with the Neuville limestone. In the subsurface the Tetreauville limestones pass abruptly into dark grey and black shales in the southern part of the Quebec Basin.

With the exception of the Gilmour Lake outlier, late Middle Ordovician (Wilderness–Barneveld) limestones form most of the strata in the Paleozoic outliers on the southern Canadian Shield where they rest directly on Precambrian crystalline rocks. Thickest and most complete sequences occur at Gilmour Lake (504 feet), Lake Timiskaming (374 feet), and Lake St. John (153 feet).

The limestones are for the most part lithologically similar to and the equivalent of the Black River and lower Trenton Groups of Wilderness age in southwestern Ontario, and in Quebec and Anticosti Basins. Rocks of Wilderness age commonly contain a clastic veneer at their base. At Lake Timiskaming (Sinclair, 1965), the Guigues Formation consists of 97 feet of coarse orthoquartzitic sandstone succeeded by the Bucke Formation—79 feet of grey, green, and red sandy shale. Strata of Wilderness age at Lake St. John are grey-brown limestones of the Tremblay Formation, 15 feet thick, overlain by limestone of Simard Formation, 63 feet thick. To the north in the Lake Manicouagan outlier 35 feet of thin-bedded black limestone is overlain by 25 feet of thin-bedded yellowish limestone, the youngest strata exposed. Grey-brown and tan late Barneveld limestones of the Farr Formation, 98 feet thick, and of the Shipshaw and Galets Formations, totalling 75 feet, were deposited disconformably on Wilderness limestone at Lake Timiskaming and Lake St. John, respectively.

In western Anticosti Basin (Roliff, 1968), in subsurface, about 1,000 feet of unnamed limestone equivalent to Black River and Trenton Groups rest disconformably on the Mingan Formation. The strata consisting of brown cherty limestones are not exposed, but presumably rise to form the bedrock surface between Anticosti and Mingan Islands.

In western Newfoundland late Middle Ordovician strata occur only in western Port au Port Peninsula. There, the Long Point Formation, of late Wilderness age, consists of about 2,800 feet of strata, a relatively thick sequence that indicates strong subsidence of this part of the platform. The lower 250 feet comprises nodular and argillaceous limestone that rests with angular unconformity on the western, leading edge of the Humber Arm klippe. The upper part of the formation consists of dark greenish shale, bluish calcareous sandstone, and minor thin limestone interbeds.

Taconian Orogeny, Late Phase

During the late phase of the Taconian Orogeny in the Upper Ordovician, late Middle Ordovician and older strata northwest of Quebec Geanticline in southern Quebec were folded and faulted, and the entire geanticlinal belt was uplifted and eroded (Fig. VI-12). Folds are open to tight with axial planes vertical to southeast dipping. Slaty cleavage is developed in pelitic beds. Many folds are overturned northwesterly, and thrust faults and associated transverse faults indicate northwestward transport. Late Middle Ordovician strata in the Eastern Townships are cleaved, and the open to tight folds are upright or slightly overturned to the northwest. Regional metamorphism is no higher than slate and phyllite grades. Strata of probable Upper Ordovician age lie conformably or with low-angle unconformity on late Middle Ordovician strata, and Silurian strata lie with angular unconformity on the Ordovician and older rocks. Uplift of southwestern Quebec Geanticline was apparently greater than the eastern part as a large volume of detritus derived from the geanticline was deposited in Quebec Basin and Allegheny Trough, whereas in Gaspé Peninsula little or no detritus was deposited in the Matapedia belt to the south nor in Anticosti Basin to the north.

In western Notre Dame Trough of Newfoundland the Burlington Geanticline was probably uplifted and eroded. It was probably the source of the conglomerates in the Notre Dame Trough of upper Late Ordovician and Lower Silurian age.

The Miramichi Geanticline was probably initiated in the late Ordovician and underwent regional metamorphism, folding, granite intrusion, uplift, and erosion. Much of the pre-late Middle Ordovician quartzite and shale formation of the core of the geanticline was metamorphosed to schist and gneiss, locally with sillimanite of the almandine amphibolite facies. Bedding and foliation in these rocks and in the overlying Middle and Upper (?) Ordovician Tetagouche Group trend partly northeast parallel with later Acadian structures and partly north to northwest quite unlike the Acadian structures in the Silurian and Devonian rocks flanking the geanticline. Strata in the core commonly have gentle dips, an indication of either recumbent folds or little deformation in contrast with the steep and generally open structures in the Silurian and Devonian rocks. Biotite quartz monzonite and granodiorite intruded the quartzite–slate geanticlinal core rocks and were penetratively deformed and altered to a cataclastic granite with weak foliation, and locally to zones of mylonite in and near the Catamaran fault that was reactivated during Acadian Orogeny. Muscovite and biotite from several samples of granite and gneiss give K–Ar dates ranging from 497 to 385 m.y. The younger dates reflect Acadian deformation and granite intrusion. Those in the 430 to 460 m.y. range suggest Upper Ordovician orogeny.

Upper Ordovician (Fig. VI-12)

Upper Ordovician strata are of restricted distribution in Appalachian Geosyncline, largely as a result of emergent land brought about by late Taconian uplifts, but are widespread in St. Lawrence Platform with the exception of western Newfoundland.

Appalachian Geosyncline. In the Eastern Townships the East Branch Pond and Sherbrooke Formations contain an Upper Ordovician shelly fauna (St-Julien, 1967). They are lithologically similar, a few thousand feet thick, lie with angular unconformity on late Middle Ordovician strata of the Magog Formation, and are unconformably overlain by Silurian strata. Farther northeast the Sherbrooke Formation lies conformably on Magog strata. Each formation consists of a basal conglomerate, sandstone, and some slate, overlain by sandstone and slate that pass upward into grey to dark grey slate. Clasts in the conglomerate are pebbles to boulders of various sedimen-

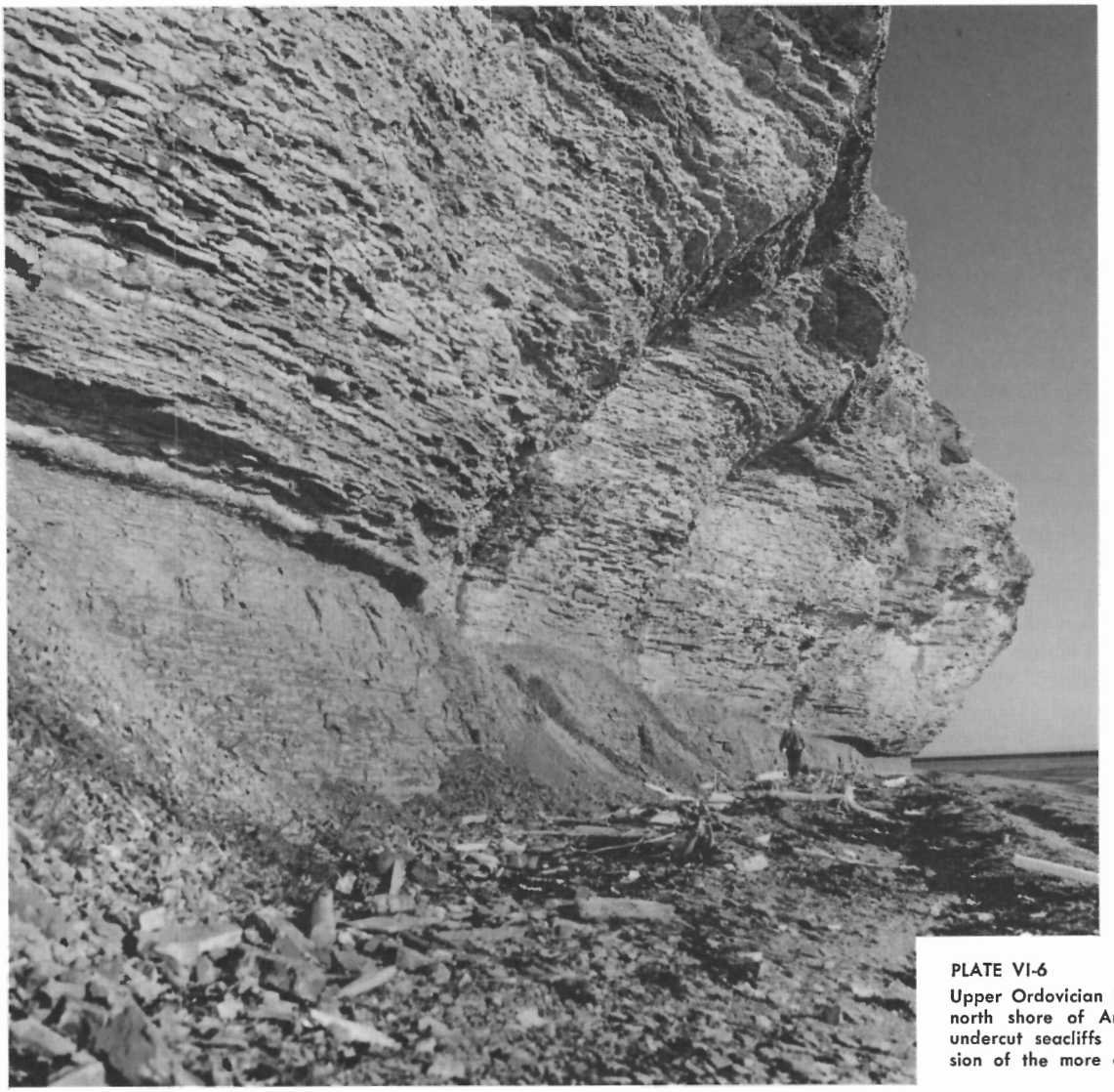


PLATE VI-6
Upper Ordovician English Head Formation, north shore of Anticosti Island, Quebec; undercut seacliffs produced by wave erosion of the more argillaceous limestones.

tary rocks, mafic and rhyolitic volcanics, albite granite, and rare serpentinized peridotite and pyroxenite.

In northern Nova Scotia quartzite containing a Middle or Upper Ordovician shelly fauna occurs as blocks within a late Paleozoic fault breccia and as boulders in nearby Mississippian conglomerate on Cape George. The quartzite is similar to some rocks associated with the iron-formation of the Browns Mountain Group, but otherwise its source is unknown. The nearby Malignant Cove Formation lacks fossils and consists of 20 feet of red, nonmarine, crossbedded conglomerate and grit containing a variety of fragments including chert, granite, rhyolite, and greywacke. It rests with angular unconformity on slate of Browns Mountain Group and is isolated by drift from younger rocks. Its age is unknown; it could be Carboniferous but its high degree of induration suggests pre-Carboniferous and probably pre-Silurian ages.

In axial Notre Dame Trough of northern Newfoundland (Kay, 1967), the upper part of Exploits Group is dominantly greywacke and fine pebble-conglomerate, several thousand feet thick, conformably overlying the Middle Ordovician slate and argillite and locally including lower Silurian. Brachiopods in part of the sequence are Upper Ordovician. In southwestern Notre Dame Bay (Espenshade, 1937) the Crescent Lake and Roberts Arm Formations are unfossiliferous and have been referred to the Upper Ordovician, although lithologically they are unlike presently known Upper Ordovician strata. The Crescent Lake Formation consists of at least 500 feet of red and green tuffaceous shale, chert and minor conglomerate, and sandstone. The conformably overlying Roberts Arm Group is 1,500 feet of pillowed basalt flows and minor rhyolite. The upper unit of Gander Lake Group, already described, has yielded graptolites of Wilderness and of Eden age, and like the Exploits Group it may be conformably overlain by a continuous sequence of uppermost Ordovician and Lower Silurian greywacke and conglomerate.

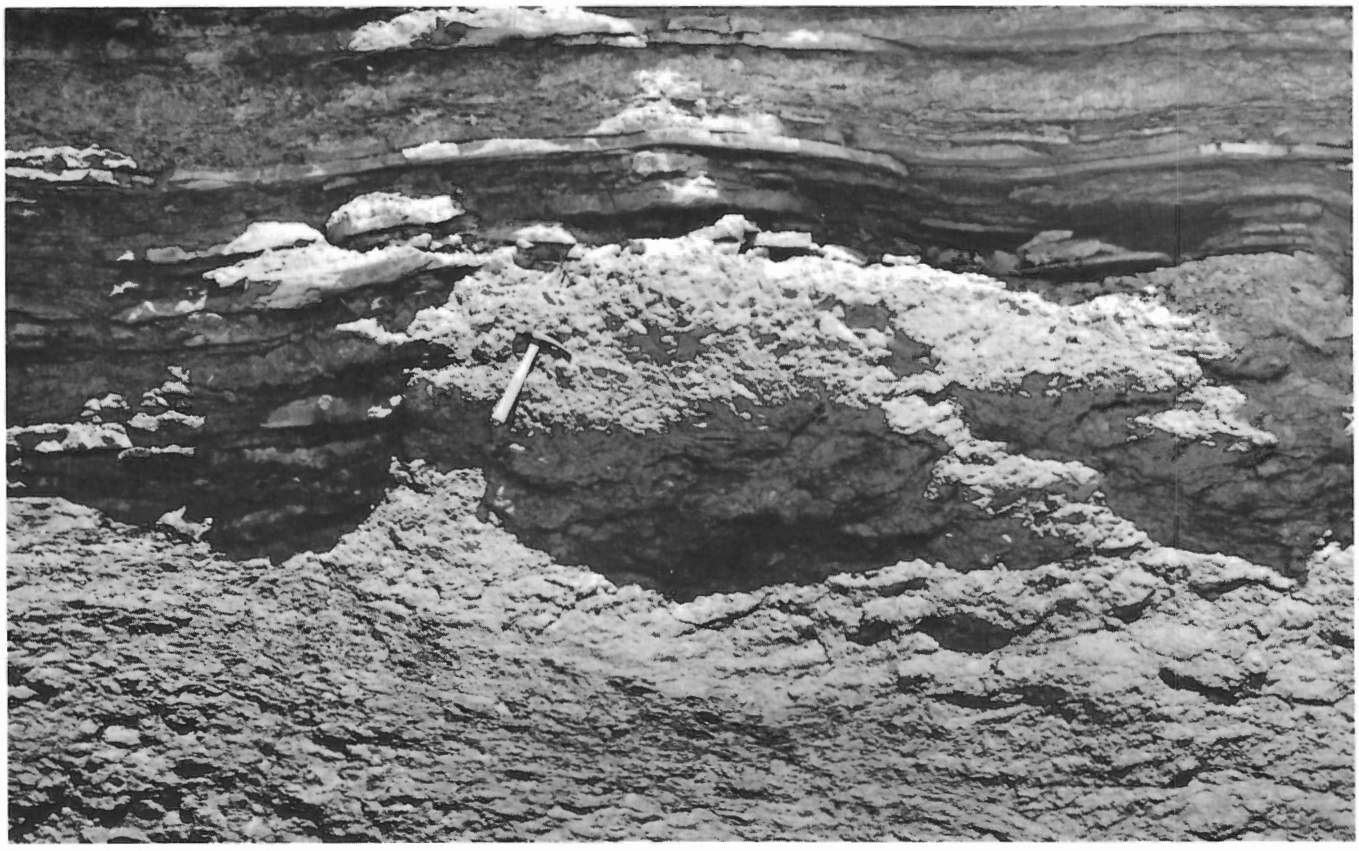


PLATE VI-7. Bioherm in Upper Ordovician, Ellis Bay limestone, Anticosti Island, Quebec.

St. Lawrence Platform. Upper Ordovician fine and coarse clastics of marine and non-marine origin occur in southeastern Allegheny Trough and southeastern Quebec Basin where they are 2,000 and 4,000 feet thick, respectively. The strata thin northwestward and ultimately grade to marine carbonate and shale, as much as 800 feet thick, in Michigan Basin. In western Anticosti Basin as much as 3,100 feet of limestone and shale was deposited.

In southwestern Ontario the Collingwood Formation (Eden) consists of black euxinic shale resting conformably on limestone of the Trenton Group (Sanford, 1961; Beards, 1967). Beneath eastern Lake Erie it may be partly Barneveld and a facies equivalent of the upper Trenton limestones to the northwest. From a thickness of 200 feet beneath eastern Lake Erie the Collingwood thins northwestward to 20 feet near southern Georgian Bay and thickens to 60 feet in the Manitoulin district where it disconformably overlies limestone of the Cobourg Formation. The succeeding Blue Mountain Formation of dark grey and brown shale is 250 feet thick near western Lake Ontario and thins northward to 120 feet near southern Georgian Bay. The equivalent Sheguiandah Formation on Manitoulin Island consists of 100 feet of soft brown, maroon, and grey shale. The Meaford–Dundas Formation (Maysvillian and early Richmondian) is made up of grey shale, siltstone, and minor limestone. These strata have a maximum thickness of 900 feet near western Lake Ontario and thin southward into Allegheny Trough where the upper beds interdigitate with red shale and siltstone of the overlying Queenston Formation. The unit also thins northwestward, and in the Manitoulin region the lower part is termed the Wekwemikongsing Formation —150 feet of grey shale with minor limestone interbeds. The upper part, the Meaford Formation, comprises 60 feet of grey and brown argillaceous limestone. The overlying Richmondian Queenston Formation is 1,100 feet thick beneath eastern Lake Erie and consists of semicontinental red shale and siltstone derived from the Taconian mountains to the east. The Queenston also thins northwestward giving place to 160 feet of bluish grey, argillaceous limestone and dolomite of the Kagawong Formation on Manitoulin Island.

In Quebec Basin (Fig. VI-7) Upper Ordovician strata are similar to those of southwestern Ontario, and except for the Lower Devonian limestone fragments in a Cretaceous diatreme breccia, are the youngest strata exposed (Clark, 1956; Wilson, 1946). They are about 4,500 feet thick in the southern part of the basin and thin to 879 feet in Ottawa Embayment. In the embayment the Edenian Eastview Formation, 20 feet thick, consists of interbedded limestone and black shale that gradationally overlie Cobourg limestones and are apparently absent in the northern part of the Quebec Basin. The gradationally overlying Billings Formation is represented by 170 feet

of black bituminous shale that thickens eastward to 300 feet and forms the Lachine and Lotbinière Formations in northern Quebec Basin disconformably overlying the Trenton Group. In the subsurface south of St. Lawrence River as much as 1,000 feet of similar strata usually referred to the Utica Group lie with gradational contact on the Trenton Group. The Carlsbad Formation (Maysville) in the Ottawa Embayment consists of shale and minor limestone 607 feet thick, increasing to more than 2,000 feet in southern Quebec Basin where the equivalent Nicolet River Formation consists of grey shale with siltstone and sandstone interbeds. These shales are overlain by blue-grey argillaceous limestone of the Richmondian Russell Formation, 40 feet thick, and Pontgravé River Formation near Montreal, 173 feet thick. The Queenston Formation represents the youngest Ordovician strata; it comprises 42 feet of red shale with minor limestone interbeds in the Ottawa Embayment that thicken eastward in southern Quebec Basin to 2,000 feet of nonmarine deltaic red shale, with minor limestone interbeds, and locally abundant red siltstone and sandstone.

On the Canadian Shield Upper Ordovician strata occur in outliers at Lake Timiskaming, Clear Lake, and Lake St. John (Sinclair, 1965). They are the youngest strata exposed at the last two localities. About 20 feet of early Edenian bituminous shale and minor limestone, equivalent to the Eastview and Collingwood Formations, rests conformably above Trenton limestone in the Clear Lake outlier. At Lake Timiskaming, the late Edenian Dawson Point Formation disconformably overlies the Middle Ordovician Farr Formation and consists of dark grey shale 98 feet thick. Equivalent beds at Lake St. John comprise 25 feet of black bituminous shales commonly referred to the Gloucester Formation that disconformably overlies the Galets Formation. Richmond strata occur only at Lake St. John and consist of limestone that is mostly covered by the lake waters.

In Anticosti Basin (Fig. VI-7) Upper Ordovician strata occur on Anticosti Island and beneath the Gulf of St. Lawrence. They are not present in western Newfoundland. On Anticosti Island (Bolton, 1961; Roliff, 1968) Upper Ordovician limestone and shale about 3,100 feet thick overlie Barneveld–Wilderness limestone and are succeeded by Lower Silurian (Alexandrian) rocks. They are lithologically similar to Upper Ordovician strata of western Michigan Basin, but are much thicker and lack the clastic Richmond red bed assemblage of Allegheny Trough and Quebec Basin. The Edenian Macasty Formation forms the bedrock surface beneath part of the Gulf of St. Lawrence north of Anticosti Island and in the subsurface consists of black bituminous shale, 200 feet thick, presumably equivalent to the Utica, Lachine, and Billings Formations of the Quebec Basin. The conformably succeeding 750 feet of alternating grey shale

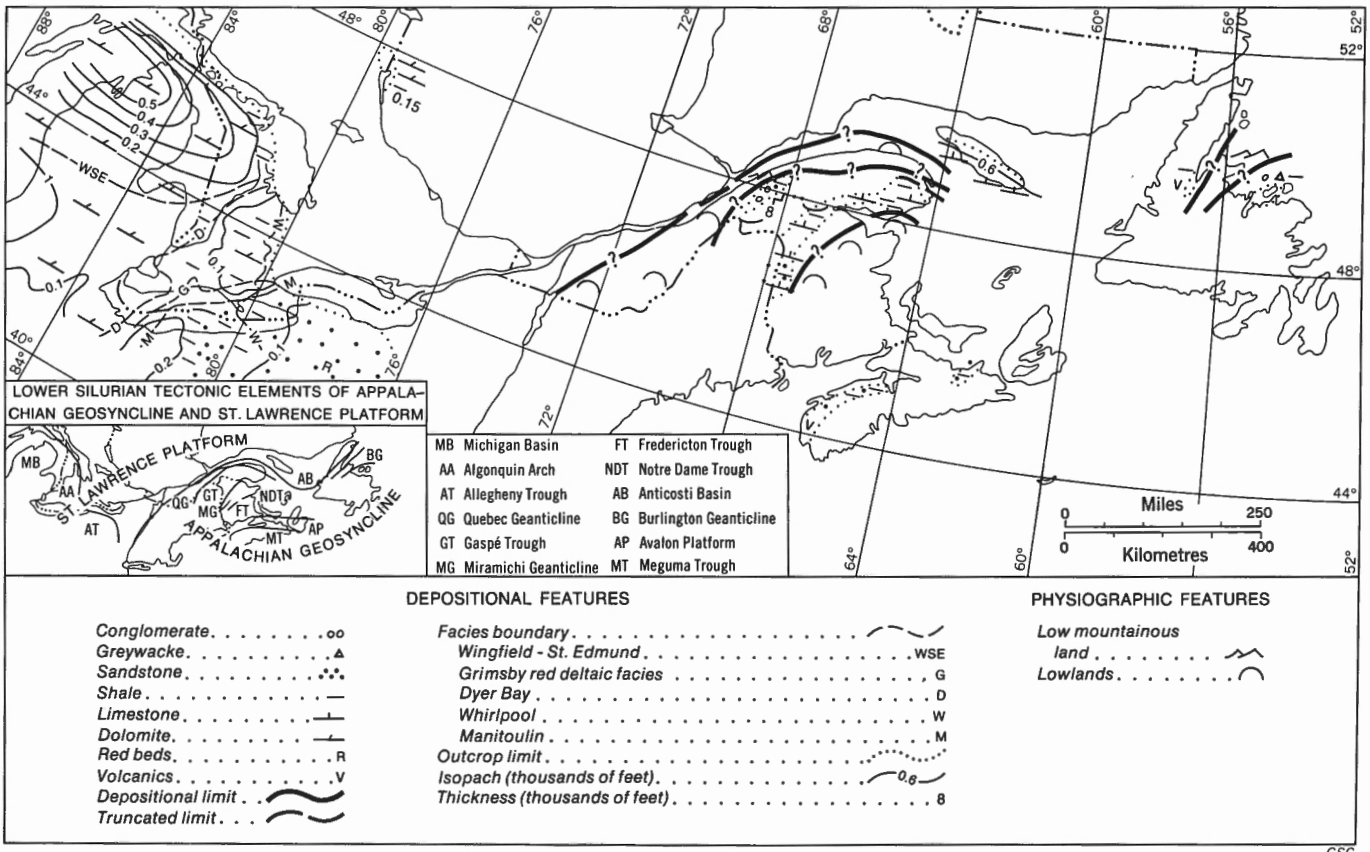


FIGURE VI-13. Lower Silurian sedimentation, volcanism, and tectonism in Southeastern Canada.

and limestone of the lower English Head Formation is known only in subsurface and probably forms the bedrock surface beneath the channel north of Anticosti Island. The succeeding early Richmondian upper English Head (Pl. VI-6) and Vauréal Formations, totalling more than 1,850 feet, consist of bluish grey, finely crystalline to semilithographic limestone and interbedded greenish grey shale. Although predominantly limestone, the shale and silt content of the English Head Formation increases toward the northeast end of Anticosti Island. The late Richmondian Ellis Bay Formation in western Anticosti Island comprises at least 200 feet of argillaceous limestone, shale, and limestone and contains bioherm reefs (Pl. VI-7). It thickens to 300 feet on northeastern Anticosti Island, grading to a dominantly sandstone facies.

Silurian

Tectonic Summary

During the Silurian, sedimentary and volcanic rocks thousands of feet thick were deposited within Appalachian Geosyncline, and carbonates and evaporites were deposited on St. Lawrence Platform. No granitic intrusions or episodes of folding are known to have occurred. Tectonic elements evident in the Ordovician persisted throughout the Silurian (Figs. VI-13, 14, 15).

In northern Gaspé Trough several thousand feet of siltstone, limestone, and quartzose sandstone, derived mainly from the northwest, were deposited unconformably on the folded Ordovician rocks of Quebec Geanticline. The age of the base of the sequence varies from early Llandovery to latest Ludlow. Along the axis of Gaspé Trough in the Matapedia facies belt, deposition of thin-bedded, possibly deep-water, limestone and shale continued from the late Ordovician into earliest Silurian, grading to shale and sandstone with local manganiferous deposits in the southwest, and to shale, siltstone, limestone, and volcanic rocks in the northeast. In southern Gaspé Trough, siltstone, sandstone, volcanics, and locally conglomerate were deposited adjacent to Miramichi Geanticline; these rocks probably lie unconformably on the Ordovician. In Fredericton Trough turbidites were deposited in the north; in the south they grade into mainly silicic volcanics and shallow-water green, red, and brown shale, sandstone, and minor limestone bordering Avalon Platform.

In Notre Dame Trough of Newfoundland lands along Burlington Geanticline and possibly also Avalon Platform bordered the Silurian depositional elements. No geanti-

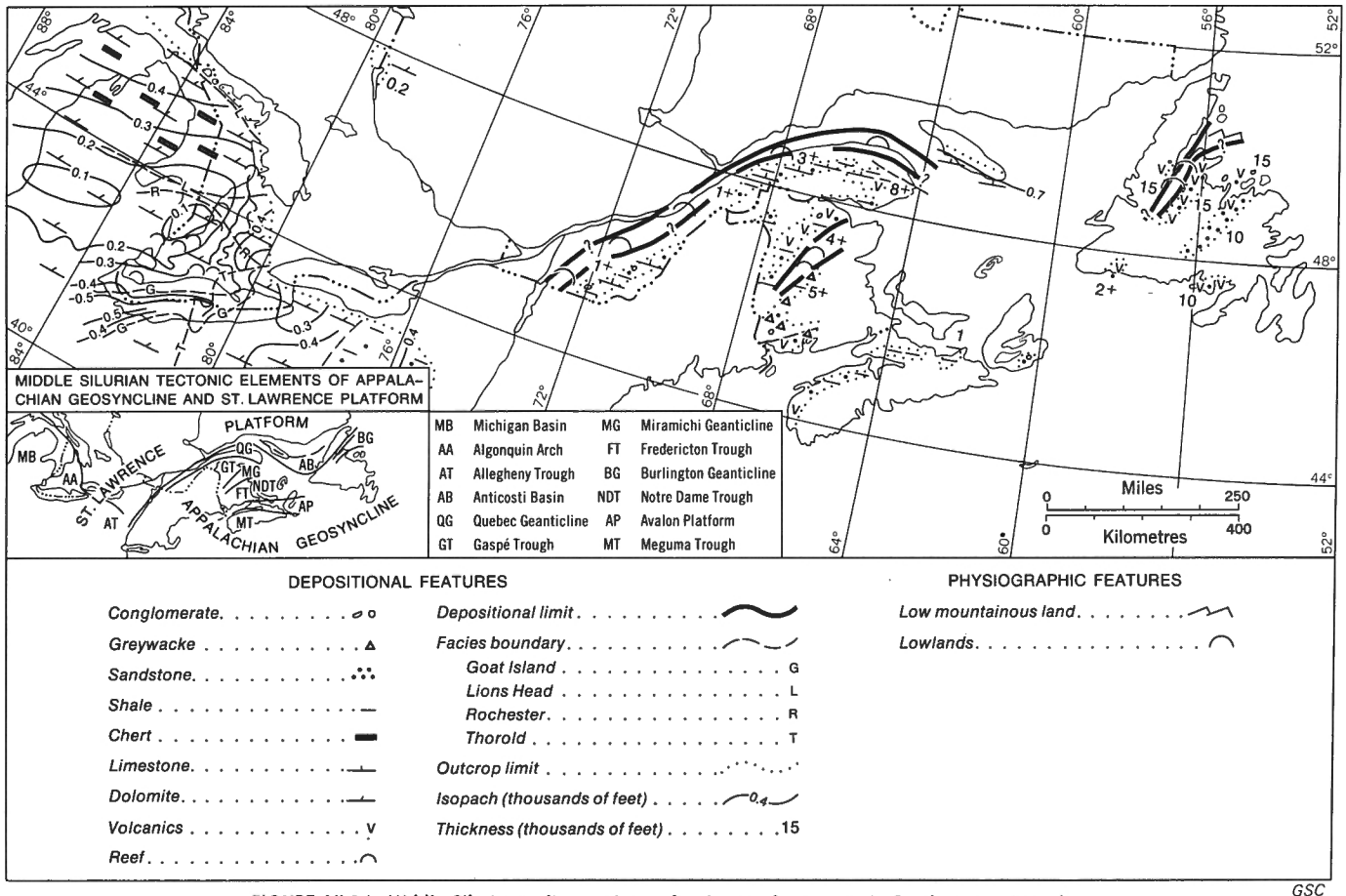


FIGURE VI-14. Middle Silurian sedimentation, volcanism, and tectonism in Southeastern Canada.

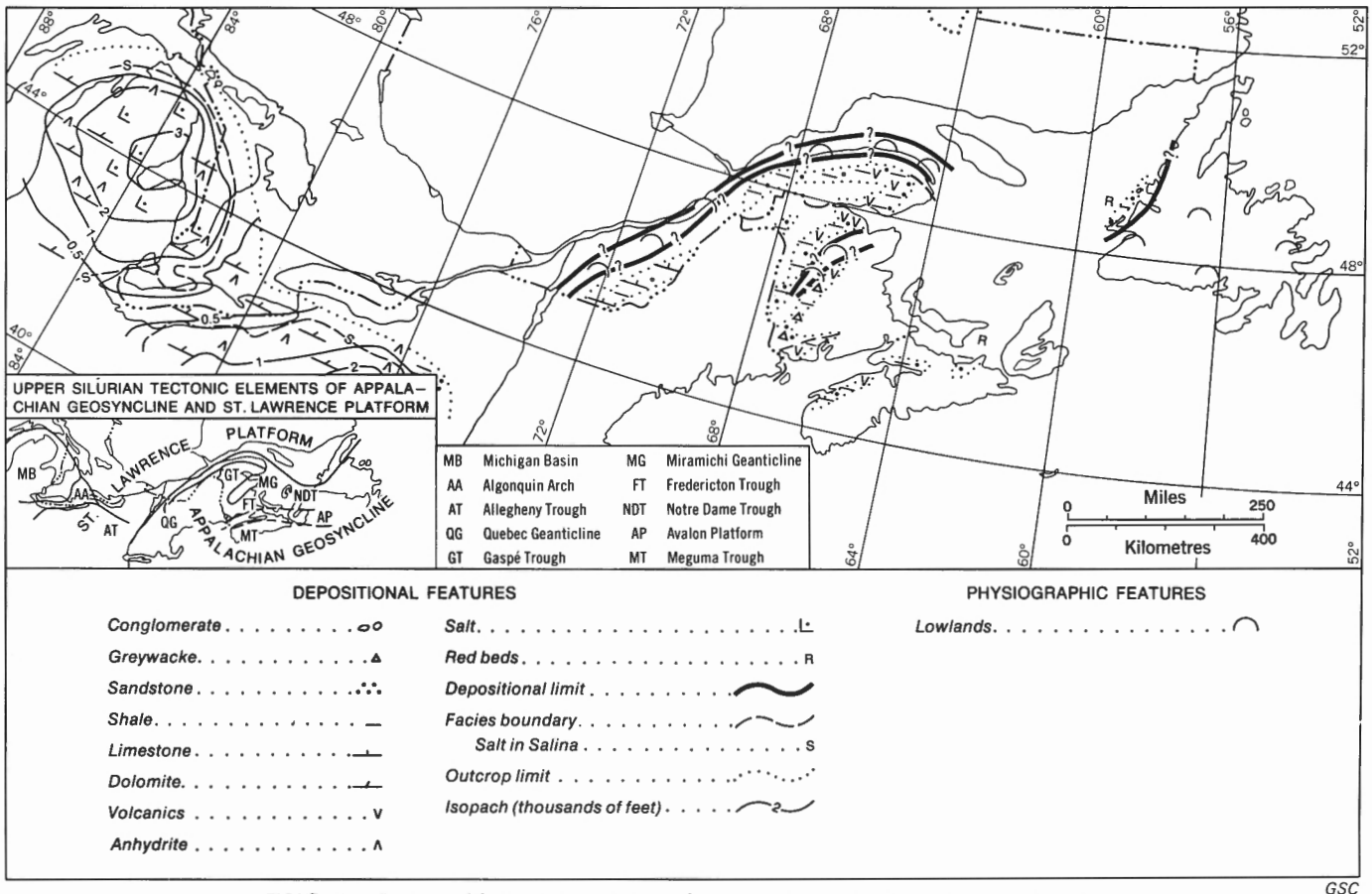


FIGURE VI-15. Upper Silurian sedimentation, volcanism, and tectonism in Southeastern Canada.

cline in the central part of the trough, analogous to Miramichi Geanticline, has been identified, although some facies suggest that such lands existed during much of the Silurian. During Llandoverian and Wenlockian times the depositional environment changed from marine to terrestrial. Thick, very coarse conglomerates indicate either strong uplift to the north beneath Notre Dame Bay, or possibly an extension of Burlington Geanticline.

There is no evidence of Silurian folding and plutonism, but the scarcity of Upper Silurian and Lower Devonian strata makes it difficult to separate older orogenic effects from those of the Devonian Acadian Orogeny. Greywacke was deposited conformably on Ordovician rocks in the northeastern axial part of Notre Dame Trough and is conformably overlain by conglomerate. The Wenlockian volcanics and overlying red and grey, mainly terrestrial sandstones with abundant fluviatile depositional features unconformably or disconformably succeed Ordovician and older rocks. The Silurian strata of Newfoundland generally reflect a change from marine to terrestrial conditions of deposition.

On Avalon Platform, Nova Scotia, a thick, complete Silurian sequence, mainly of shale, was deposited probably unconformably on Ordovician rocks, whereas in eastern Cape Breton Island, undated presumably Silurian sandstone and conglomerate were deposited unconformably on Cambrian rocks.

In Meguma Trough the very thick, undated quartzose sandstone, shale, greywacke, and silicic and mafic volcanics that were deposited conformably on Lower Ordovician slate were succeeded by Upper Silurian shale, siltstone, quartz sandstone, and local volcanic conglomerate.

On St. Lawrence Platform, Silurian strata are now confined to southwestern Ontario, Anticosti Island, and Lake Timiskaming outlier. Most of the southern Canadian Shield was probably covered by marine sedimentary rocks that were formerly connected, at various times, with those of Williston Basin and Hudson Platform. During the Early Silurian coarse red deltaic detritus from the Taconian mountains in eastern New York was spread westward across Allegheny Trough. On the margin of Michigan Basin it grades into marine shales and carbonate rocks. These facies also extended across the southern Canadian Shield to the Lake Timiskaming outlier. During the Middle Silurian large crinoid banks and complex biostromal or barrier reefs formed around Michigan Basin. Numerous patch and pinnacle reefs developed along the flanks of the rapidly subsiding Michigan Basin. Carbonates and fine clastics were being deposited in Allegheny

Trough to the southeast. Evaporite deposits dominated Upper Silurian sedimentation in Michigan Basin and Allegheny Trough, giving place to carbonates and fine clastics over Algonquin Arch in Ontario. In western Anticosti Basin, Alexandrian and Niagaran limestones were deposited conformably on Upper Ordovician limestone and are the youngest strata exposed. In eastern Anticosti Basin the Silurian may be represented in a coarse clastic red bed sequence, the upper part of which contains latest Silurian or earliest Devonian (Pridoli) fossils.

Appalachian Geosyncline

Gaspé Trough. In southern Quebec, Silurian strata of Gaspé Trough lie with angular unconformity on Cambrian to Middle Ordovician rocks of Quebec Geanticline, and in one place with low angular unconformity or disconformity on the Upper Ordovician. The Silurian seas did not cover the geanticline until the late Wenlockian and Ludlovian. Near Lake Memphremagog, Silurian strata form two synclinal remnants (Cooke, 1950). The basal, Peasley Pond Formation consists of unfossiliferous, well-stratified conglomerate and quartzite varying from a few feet to more than 250 feet in thickness. To the southeast the Peasley Pond is missing, and the overlying formations rest directly upon Ordovician rocks. Clasts in the conglomerate are mainly quartz pebbles with minor chert and fine-grained quartzite. The overlying Glenbrooke Formation is grey calcareous slate and slaty argillite of probable early Ludlovian age. In the west it is 900 feet thick but may be much thicker elsewhere. The gradationally overlying Sargent Bay Formation is the youngest Silurian exposed and consists of grey limestone and slaty argillaceous limestone of Ludlovian age. In the west it is 290 feet thick; farther east it is perhaps several thousand feet thick. To the east and northeast Silurian rocks similar to those described comprise the St. Francis, Cranbourne, and Lake Aylmer Groups. All are mainly Ludlovian limestone, in part sandy and argillaceous. Late Wenlockian to early Devonian conglomerates rest unconformably on Cambrian and Ordovician rocks. They may be as thick as 2,000 feet and contain clasts of quartz, various sedimentary rocks, silicic and mafic volcanics, and albite granite.

In the Gaspé Peninsula and Témiscouata and Matapedia regions the Silurian generally consists of four megafacies: Llandoverian shale and coarse clastics, Llandoverian and Wenlockian orthoquartzite, Wenlockian and Ludlovian limestone, and Ludlovian siltstone with minor limestone and coarse clastics. Volcanic rocks are intercalated mainly in the Wenlockian and Ludlovian strata in central and southern Gaspé Peninsula.

In the Témiscouata region, along northern Gaspé Trough, Silurian sedimentary strata thicken from 4,000 to 18,000 feet to the southeast and to 16,000 feet in a local basin in the northeast (Lajoie, *et al.*, 1968). The early and middle Llandoverian Cabano Formation is the oldest formation unconformably overlying the Cambrian to Middle Ordovician Quebec and Trinité Groups. It consists of grey lithic sandstone, conglomerate with clasts of the Quebec Group, and shale. It thickens southeasterly from zero to 8,000 feet where it is covered by younger strata, and thins northeasterly to zero. Source of the detritus is Quebec Geanticline to the northwest. The conformably overlying late Llandoverian Pointe Aux Trembles Formation and its northeastern, finer grained equivalent, the Lac Raymond Formation, thicken southeasterly from 1,000 to 6,000 feet. The Pointe Aux Trembles consists of volcanic sandstone and conglomerate with rare latite flows and tuff. Well-rounded clasts from 1 inch to 8 inches in diameter are volcanic rock. The Lac Raymond Formation is mainly greenish grey shale with interbeds of tuffaceous and pebbly sandstones containing quartz and volcanic rock. The source of the volcanic detritus was probably to the southeast in northwesternmost New Brunswick, now covered by Devonian strata. The overlying Robitaille Formation and laterally equivalent Asselin Formation are believed to be of early Ludlovian age and to overlie disconformably the Pointe Aux Trembles and Lac Raymond Formations. The Témiscouata region is inferred to have been a land area during the Wenlockian. The Robitaille consists of red to white siltstone and lithic and quartz sandstones with minor green shale, limestone, dolomite, and pebble-conglomerates. Along the northwest the Robitaille unconformably overlies the Quebec Group and reaches its maximum thickness of 1,900 feet. It intertongues northeasterly with the Val Brillant and Sayabec Formations that extend southwest from the Matapedia region. Along the southeast the Robitaille in one place is 700 feet thick and grades northeasterly into the Asselin Formation, which is 1,300 feet of granule conglomerate and sandstone containing detritus of quartz and silicic lava. The overlying Mont Wissick Group consists of three conformable formations. The oldest, the Sayabec Formation, is early Ludlovian. It conformably overlies the Robitaille Formation along the northwest and is missing in the southeast. It consists of up to 400 feet of argillaceous and silty limestone. The overlying Ludlovian St. Léon Formation is 2,300 to 3,500 feet thick and consists of grey and green siltstone and sandstone. To the southeast it overlies the Robitaille and Asselin Formations, and to the northwest it overlies the Sayabec Formation and thickens to 10,600 feet in a local basin. The middle 5,800 feet of the St. Léon consists of sandstone, conglomerate, and greenish grey siltstone of the Lac Des Baies Member.

In the Matapedia region (Béland, 1960) along northern Gaspé Trough about 10,000 feet of Silurian strata comprise four conformable formations. The Awantjish Formation, the oldest, consists of 150 to 1,000 feet of late Llandoverian green, grey, and rare red calcareous shale and siltstone that rest unconformably on folded Ordovician rocks in a small inlier. The conformably overlying late Llandoverian to early Wenlockian Val Brillant

Formation is 500 feet thick and thins northward to 90 feet as it overlaps the Awantjish to rest unconformably upon folded Ordovician rocks. It consists of white and pink quartzite with minor argillaceous quartzite, probably derived from the north. It grades upward through calcareous sandstone into limestone of the Wenlockian and lower Ludlovian Sayabec Formation, 1,200 to 1,500 feet thick, which consists of pure, grey limestone with mud-cracked shaly and silty interbeds. Locally, the limestone is a breccia of angular fragments of algae and corals in a limestone matrix. The Sayabec grades upward into the Ludlovian St. Léon Formation, 7,300 feet thick, consisting of grey to green and rarely red calcareous siltstone and fine-grained sandstone with minor red shale, limestone, sandstone, and conglomerate with quartz and quartzite pebbles. Some conglomerate consists of rounded fragments of limestone, silty limestone, and fossils in a limestone matrix; at one locality the fragments are as much as 15 feet across.

In Gaspé Peninsula and western New Brunswick, in the Matapedia facies belt along the axial region of Gaspé Trough, deposition of thin-bedded limestone and shale continued from late Ordovician to about middle Llandoverian without interruption and without recording the effects of late Ordovician uplifts of the Quebec and Miramichi Geanticlines. During this interval the belt may have behaved as an intra-trough platform covered by relatively deep water but shallower than the flanking parts of the Gaspé Trough. In western New Brunswick lower Llandoverian limestone and slate grade upward through slate to upper Llandoverian, Wenlockian, and Ludlovian thin-bedded, grey to green, and locally red micaceous siltstone, quartzite, and slate with local developments of greywacke, conglomerate, limestone, red slate, and manganiferous iron-formation. In central and eastern Gaspé Peninsula (Burk, 1964) local volcanic centres produced flows, pyroclastic rocks, and volcanic conglomerates. At the western end of St. John anticline (Skidmore, 1965) the upper Llandoverian to Ludlovian Sirois Formation is 4,000 feet or more thick and comprises two members. The older, Laforce Member, 1,800 feet thick at its type locality, overlies the White Head Formation with uncertain but probably conformable or disconformable relations. The Laforce consists of sandy limestone, limestone grit and conglomerate, calcareous sandstone, and within the lower 500 feet of the member, some volcanic conglomerate. The upper, unnamed member is 2,200 feet thick at its type locality and consists of green to grey calcareous siltstone, shale, and silty limestone. Farther east along the St. John anticline (Cumming, 1959) the Burnt Jam Brook Formation of late Llandoverian age overlies the White Head Formation and is partly equivalent to and partly older than the Laforce Member. The formation consists of 400 to 500 feet of green to grey shale overlain conformably by the Owl Capes Formation, 1,000 feet of limestone-conglomerate, sandy and silty limestone, and volcanic conglomerate. About 2 miles west is the Cedar Barn volcanic member, 1,800 feet of interbedded basic lava flows and volcanic conglomerate that appears to be equivalent to Owl Capes Formation and contributed volcanic detritus to it. These formations are probably Middle Silurian. The Cedar Barn and Owl Capes are apparently overlain by Sirois beds. About 10 miles to the south is the Mount Alexander Group (Skidmore, 1965) of 8,000 to 16,000 feet of volcanic and sedimentary Silurian rocks possibly ranging from Llandoverian to early Ludlovian. A basal unit of 2,400 feet of silty limestone rests with uncertain relations on the White Head Formation. The limestone is overlain by 2,000 feet of interbedded shaly to silty limestone and amygdaloidal andesite, succeeded by as much as 10,000 feet of dark green and dark red andesite, basalt, and keratophyre lava, agglomerate, and tuff. The uppermost unit consists of 2,000 feet of limestone. Only 3 to 4 miles to the north, where only 8,000 feet of Mount Alexander Group is exposed, there is less limestone and more siltstone and shale, a change apparently related to the distance from the volcanic centre. About 3,000 feet of andesite and basalt lava in the core of an anticline are overlain by 2,000 feet of calcareous siltstone and minor sandstone and limestone. Locally, conglomeratic limestone and conglomerate contain boulders as much as 2 feet across of mainly volcanic rock and limestone. These strata are succeeded by 3,000 feet of non-calcareous shale and siltstone.

In southern Gaspé Peninsula (Burk, 1964; Badgley, 1956) Silurian sedimentary and local volcanic rocks of the Chaleurs Bay Group lie between the Matapedia facies belt on the north and the folded Mictaw and Maquereau rocks on the south coast. The latter is probably a northeast extension of the Miramichi Geanticline. The basal, Weir Formation is early to late Llandoverian (Ayrton, 1967) and consists of 635 to 2,180 feet of greyish green siltstone with interbeds of quartz- and jasper-pebble conglomerate, arkosic sandstone, and silty limestone. The basal contact with Middle Ordovician Mictaw Group is not exposed but is probably an unconformity. The Clemville Formation conformably overlies the Weir in the north, but near the coast it lies unconformably on the Mictaw and Maquereau Groups, and farther west, unconformably on the Honorat Group. It consists of 90 to 2,000 feet of orthoquartzite, calcareous sandstone, siltstone, and quartz-pebble conglomerate. The overlying La Vieille Formation consists of 800 to 1,800 feet of grey to dark grey, partly nodular and reefal limestone, argillaceous limestone, and calcareous shales of late Llandoverian and Wenlockian age. The lower Ludlovian Gascons Formation is from 1,450 to 3,835 feet thick and consists of green to grey calcareous and non-calcareous siltstone, and minor limestone and red siltstone. The Bouleaux Formation is 800 to 2,600 feet of calcareous siltstone and silty limestone. It is transitional between the Gascons Formation and the overlying West Point Formation of crinoidal, coralline, and partly reefal limestones with red and grey siltstone, varying from 400 to 1,700

feet in thickness. The youngest Silurian, the Indian Point Formation, consists of 456 feet of siltstone, sandstone, and minor crinoidal limestone. The top is covered by Chaleurs Bay. At Black Cape, along the southern coast of Gaspé, the lenticular Black Cape Formation comprises 2,100 feet of andesite and basalt lava, breccia, volcanic conglomerate, and minor red siltstone and grey limestone lithologically and faunally similar to the West Point Formation. Farther west along the south coast Lower Devonian rocks overlap the Silurian strata to rest unconformably on the Honorat Group. A belt of Silurian sedimentary and volcanic rocks occurs southwestward along the northwest flank of Miramichi Geanticline. Rhyolite and andesite are interbedded with sandstone, siltstone, slate, and locally volcanic conglomerate, a late Llandoverian to Ludlovian assemblage. Adjacent to Miramichi Geanticline the Silurian andesite rests, possibly unconformably, on Ordovician rocks and is overlain by Ludlovian green and red slate, sandstone, conglomerate, and minor limestone. The volcanic and sedimentary formations appear to interfinger and pinch and swell in a complex fashion (Greiner and Potter, 1966).

Fredericton Trough. Silurian strata of Fredericton Trough are almost continuous with those of southern Gaspé Trough, extending around the southwestern end of Miramichi Geanticline. In the northeast they are covered by Carboniferous sediments and probably extend beneath the central and southern parts of the Gulf of St. Lawrence. North of Fredericton over 10,000 feet of turbidite, consisting mainly of graded grey greywacke and slate containing Wenlockian graptolites, lies in fault contact with Ordovician rocks (Poole, 1963). The basal part contains graded pebble-conglomerate and black slate. The pebbles are chert, quartzite, pelite, and metamorphic rocks possibly derived from Miramichi Geanticline. The middle part consists of green and red slate and greywacke that apparently thin toward the north. The upper part comprises grey slate and a more quartzose greywacke with Silurian graptolites; it becomes pebbly near the top. Amygdaloidal andesite, quartzose greywacke, and conglomerate with early Ludlovian graptolites lie adjacent to Miramichi Geanticline, but their relationships with older and other Silurian strata are unknown. Similar grey greywacke and slate near Fredericton have yielded early Ludlovian graptolites. Near the southwest end of Miramichi Geanticline an assemblage of greywacke, limestone, calcareous slate, andesite, and basalt (in one place with early Ludlovian shelly fauna) is probably mainly Silurian. North of St. Stephen, greywacke and slate, and thin-bedded micaceous siltstone, quartzite, and green slate like that of the Gaspé Trough in western New Brunswick, are probably Silurian and have been referred to as the Pale Argillite Division of the Charlotte Group.

In southeastern Fredericton Trough (Smith, 1966) Silurian volcanic and sedimentary rocks are complexly interfingered, varying markedly in proportions and lithology from place to place, but are not well dated. Northeast of St. Stephen the Oak Bay Formation consists of 300 feet of polymictic conglomerate that rests probably disconformably on Lower Ordovician rocks. Clasts are composed of well-rounded pebbles and some boulders as much as a foot in diameter. They consist mainly of rhyolite and porphyry, quartzite and diorite; one of the limestone pebbles contained a Silurian pentamerid brachiopod. The Oak Bay thins to zero northeasterly and grades upward into several thousand feet of graded greywacke and slate that contain a Ludlovian shelly fauna. Northeast of Passamaquoddy Bay is a belt of mainly Silurian rocks, the Mascarene Group, that was deposited along the border between Fredericton Trough and Avalon Platform. The group is a variable assemblage, probably many thousands of feet thick, of andesite, basalt, and rhyolite flows; breccias and tuffs interbedded with grey, green, black, and red shale; sandstone; and minor conglomerate and limestone—all metamorphosed generally to slate and phyllite grade. In Maine the oldest strata in the assemblage are late Llandoverian, but most strata in New Brunswick are probably Wenlockian and Ludlovian. Lower Devonian rocks have recently been recognized in the Mascarene assemblage. The welded rhyolitic rocks, red shale, and sandstone accumulated under shallow-water and subaerial conditions of deposition.

Notre Dame Trough in Newfoundland. Silurian and probable Silurian rocks, as thick as 15,000 to 20,000 feet, can be grouped into four, generally conformable lithologic units (Williams, 1967). In ascending order these are: (1) marine greywacke with interbeds of conglomerate, siltstone, and slate; (2) marine conglomerate with interbeds of greywacke and argillite; (3) volcanic rocks; and (4) fluviatile red and grey micaceous sandstone. Within the several outcrop belts the assemblages differ in which units are present, the relative proportions of rock types within the units, and their thicknesses. Various local formational and group names are used. The sequence is complete only in the northeastern axial part of the trough adjoining Notre Dame Bay. Elsewhere, only younger units are present on the flanks of the trough.

On New World Island in eastern Notre Dame Bay (Kay, 1967) several thousand feet of unnamed greywacke, siltstone, and pebble-conglomerate, containing Silurian fossils in upper parts of the section, lies conformably on Middle Ordovician graptolitic slate. Sole marks suggest current flow toward the south and southwest. The greywackes are conformably overlain by the Goldson Formation, more than 2,000 feet thick, consisting of grey to buff, brown, and red pebble- and boulder-conglomerate, with interbeds of sandstone, limy argillite, and coralline shale. Clasts are subangular to well-rounded volcanic rocks, chert, sandstone, argillite, diorite, granite, and metamorphic rocks, their proportions varying from place to place. The conglomerates show graded bedding, scouring, channelling, and locally, crossbedding. Sole

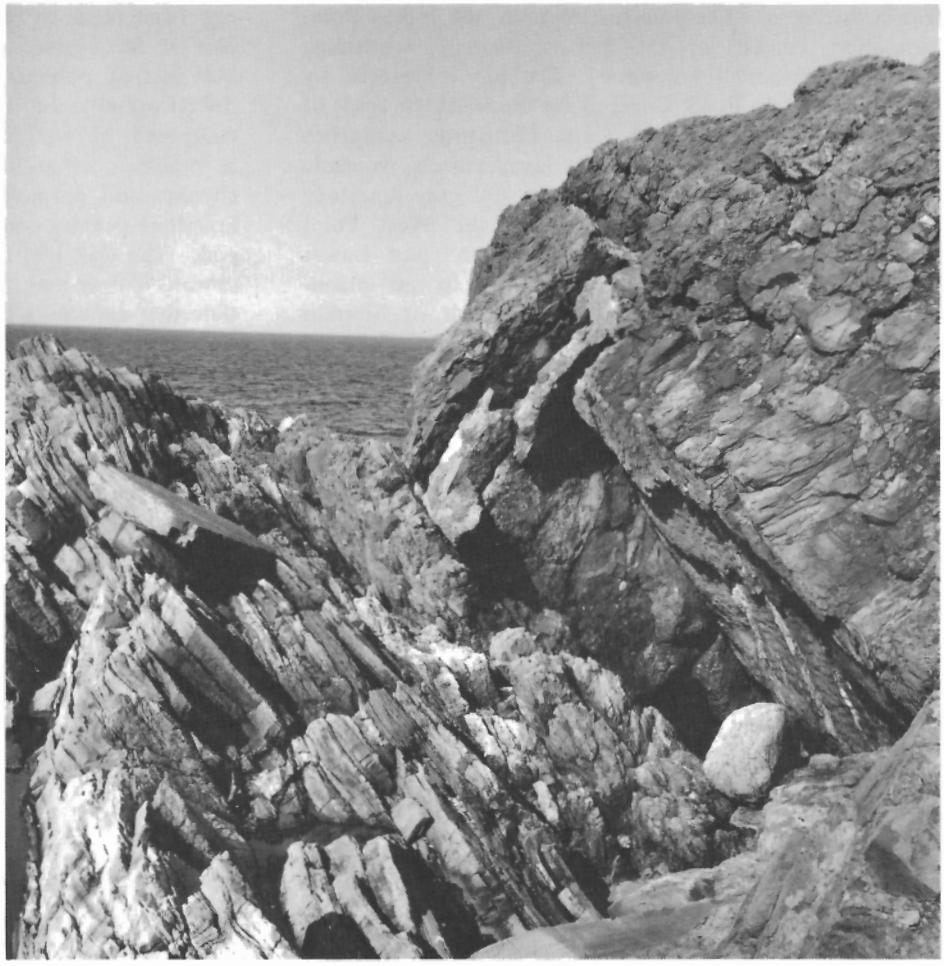


PLATE VI-8

Silurian red sandstone overlain by agglomerate, Botwood Group, Little Fogo Islands, Newfoundland.

marks on greywacke beds suggest a southerly flow, and the conglomerates are coarser and thicker in the northern exposures. A source in northern Notre Dame Bay region is suggested, probably a northeastern extension of Burlington Geanticline. In a few localities conglomerate directly overlies Middle Ordovician slate and volcanic rocks. Shells from clasts and from calcareous interbeds are Llandoverian, but in one locality shells found recently in the matrix of conglomerate are Upper Ordovician, suggesting that Goldson deposition occurred locally during Ordovician as well as Llandoverian time.

In southern Notre Dame Bay (Williams, 1967) near New Bay, about 1,000 feet of Goldson-type grey conglomerate lies conformably but abruptly on greywacke, siltstone, and slate locally containing Middle Ordovician graptolites. Elsewhere thicker undated greywacke may be in part late Ordovician and early Silurian. Clasts consist of pebbles and boulders of fossiliferous limestone and granite and of angular siltstone–slate blocks. One large lensoid limestone mass is 10 by 20 feet and composed almost entirely of Silurian corals. Much of the conglomerate appears to have slid into place.

Along eastern Notre Dame Bay and for 100 miles to the southwest the Silurian rocks have been assigned to a variety of formations by many geologists, but recently have been referred to the Botwood Group and to the Indian Islands Group near Horwood Bay. The Botwood Group consists of three units, each about 5,000 feet thick, in ascending order: (1) conglomerate and greywacke, (2) volcanic rocks, and (3) micaceous sandstone. To the northeast on Change Islands and on the islands north of Fogo Island, the volcanic rocks and micaceous sandstones are interbedded (Pl. VI-8).

South of Notre Dame Bay along the east side of the belt the micaceous sandstone unit overlies the conglomerate without the intervening volcanic formation, or lies with apparent structural conformity on Middle Ordovician slate. The conglomerate commonly contains pebbles of chert, volcanic and granitic rocks, large shale fragments, and fossiliferous limestone boulders. The interbedded greywacke is graded, and sole markings and convolute bedding are common. The volcanic unit includes purplish green, purple, red and green amygdaloidal and porphyritic andesitic lava and pyroclastic rocks. The micaceous sandstones are red, grey, and brown and are muscovitic and quartzose. They generally exhibit crossbedding, ripple-drift laminations, ripple-marks, and less commonly mud-cracks, rain prints, and erosion channels. These features

suggest continental, fluviatile, or shallow-water deposition with periodic subaerial exposure. Interbedded with the sandstone are pebble-conglomerate, micaceous siltstone and shale, and some calcareous shale and limestone. Fossils from the sandstone units, either indigenous to the sandstone or in clasts of a conglomerate, are late Llandoverian to Wenlockian, and where the sandstone lies directly on the Gander Lake Group a Ludlovian graptolite has been found. There is a strong possibility that the various units in the Botwood Group are partly facies equivalents. Toward the southwest the group is regionally metamorphosed to schist and gneiss. The Indian Islands Group consists of conglomerate with an underlying unit of argillite, siltstone, and coralline shale that is not present in the Botwood Group.

Southwest of western Notre Dame Bay is an extensive belt of Silurian volcanic rocks that locally are underlain by conglomerate and overlain by red sandstone. Rocks of the northern end of the belt are referred to the Springdale Group. Most of the rocks are similar to the Botwood Group in the east but the proportions are vastly different. The basal volcanic unit, which rests with uncertain relations on the Ordovician, consists of 10,000 feet of andesitic flows and pyroclastic rocks, and pale red to grey silicic tuff, agglomerate, and flows. Some silicic flows are flow layered and spherulitic. Other silicic rocks are quartz–feldspar porphyries representing both intrusions and crystal tuffs. In one locality the volcanic unit is underlain by a thin granitic pebble-conglomerate. Overlying the volcanic rocks near Notre Dame Bay is 5,600 feet of red sandstone and conglomerate with interlayered silicic and mafic flows. The sandstones contain features indicative of fluviatile or shallow-water deposition. To the southwest, near Red Indian Lake, red sandstone is intercalated in the volcanic rocks. Host rocks of the rich base-metal sulphide orebodies of the Buchans mine are volcanic rocks of probable Silurian age.

On Burlington Peninsula west of Notre Dame Bay the unfossiliferous Cape St. John Group consists of several thousand feet of silicic volcanic rocks, minor mafic volcanic rocks, and less than 5 per cent sedimentary rocks (Neale and Nash, 1963). Large areas of intrusive porphyry are intimately mixed with silicic volcanic rocks. Brown, red, and grey conglomerate, sandstone, and siltstone are intercalated in the volcanic assemblage and lie at its base. Clasts are generally volcanic rocks. Near Notre Dame Bay the Cape St. John Group lies unconformably on Lower Ordovician Snooks Arm Group and contains pebbles of Snooks Arm chert, argillite, volcanic rocks, and ultramafic rocks. To the southwest near Flatwater Pond, silicic volcanic rocks lie nonconformably on the Burlington granodiorite.

At La Poile Bay in southwestern Notre Dame Trough the undated (but probably Silurian) La Poile Group (Cooper, 1954) consists of 900 feet of polymictic pebble- and boulder-conglomerate, succeeded by 1,000 to 2,400 feet of arkosic grit, slate, and minor volcanic rocks, and by more than 10,000 feet of mainly silicic flows and pyroclastic rocks. Relations to older and younger formations are unknown. Along strike the La Poile Group passes into schist and gneiss.

West of White Bay (Neale and Nash, 1963) the Silurian is probably represented by three unfossiliferous formations as well as by the overlying Natlins Cove Formation that contains a late Llandoverian to early Wenlockian shelly fauna, like that of the Goldson conglomerate, and a recently discovered Ludlovian fauna. The Giles Cove Formation overlies Ordovician (?) Doucers Formation with uncertain relations. It consists of 1,000 feet of grey shale, and locally in the upper part includes 700 feet of interbedded andesite flows, tuff, silicic tuff, and agglomerate called the Deadmans Cove Member. The overlying polymictic conglomerate of the Jacksons Arm Formation, 300 feet thick, contains pebbles and boulders of silicic and mafic volcanic rocks, sedimentary rocks, and a few granitic rocks. The succeeding Simms Ridge Formation consists of 2,340 feet of grey to green arkosic sandstone and shale. The Natlins Cove Formation comprises some 10,700 feet of grey to greyish brown, feldspathic and argillaceous sandstone with interbeds of limestone, limy siltstone, and rare conglomerate. Clasts in the conglomerate consist of silicic volcanic rocks and quartz–feldspar porphyries, granite, and sedimentary rocks. Rhyolite and andesite of the Sops Island Member, 1,675 feet thick, occur in the middle of the Natlins Cove Formation and appear to be closely associated with intrusive quartz–feldspar porphyries like those on Burlington Peninsula.

Near Fortune Bay in southern Newfoundland (Williams, 1967) nearly 20,000 feet of unfossiliferous rocks, lithologically similar to the Silurian elsewhere in Newfoundland, overlies the Cambrian strata of Avalon Platform. The Grand Le Pierre Formation, 1,000 feet thick and the overlying Belle Bay Formation, 6,000 feet thick, consist of varicoloured silicic flows and tuff, minor basalt, and lenticular interbedded red, purple, and green siltstone, conglomerate, slate, greywacke, and minor chert. The Belle Bay overlies the Grand Le Pierre with local angular relations, and both rest with angular unconformity on the Cambrian Youngs Cove and Nine Mile Hill Formations. The Belle Bay Formation is locally overlain by 1,000 feet of tuffaceous slate, conglomerate, and greywacke of the Anderson Cove Formation. In most other places the Belle Bay Formation is overlain by the Rencontre Formation, 4,000 feet of red arkosic sandstone, conglomerate, siltstone, and shale with some mafic flows in the lower part. Clasts in the conglomerate are mainly silicic volcanic rocks. The sandstone has crossbeds, ripple-marks, and mudcracks. Crossbeds and cut-and-fill structures indicate a westward flow and suggest that there was land to the southeast.

Avalon Platform. In southern New Brunswick the Silurian rocks that lie on the northwest part of the

Cambrian and Lower Ordovician of Avalon Platform are mainly volcanic and clastic sedimentary rocks, and have been described under Fredericton Trough. In northern Nova Scotia, the Silurian and Lower Devonian Arisaig Group of Antigonish Highlands (Williams, 1914) consists of about 4,000 feet of mainly grey shale and fine-grained sandstone with abundant near-shore type shelly fossils. The lower 2,500 feet represents the entire Silurian system. The lowest formation, the Beechhill Cove, is early Llandoverian and rests disconformably or unconformably on the Ordovician Browns Mountain Group. It consists of 225 to 300 feet of green-grey sandstone and siltstone, locally with less than 2 feet of basal conglomerate containing quartz and rhyolite pebbles. The overlying Ross Brook Formation rests sharply on the Beechhill Cove and comprises 1,050 to 1,300 feet of middle and upper Llandoverian black shale, and towards the top, siltstone and fine sandstone. At least a dozen intercalated thin ash and tuff beds are present in the lowest 60 feet at Beechhill Cove. The conformably overlying French River Formation is early Wenlockian and consists of 280 feet of grey siltstone and shale. It is succeeded conformably by the McAdam Formation, 555 to 625 feet of upper Wenlockian and lower Ludlovian grey and black shale, siltstone, and some limestone concretions, phosphatic nodules, and a thin fossiliferous oölitic hematite bed. The upper Ludlovian Moydart Formation is 390 feet thick and comprises green shale, siltstone, and rarely limestone, overlain by nearly 30 feet of "Red Stratum" consisting of red shale and nodular limestone. The currents that transported detritus comprising the Beechhill Cove and McAdam Formations flowed south and southwest. Near Sunnybrae, strata equivalent to the Arisaig Group consist of two formations (Maehl, 1961). The older Glencoe Brook Formation is 910 feet thick and is similar in age and lithology to the Beechhill Cove. The conformably overlying Kerrowgare Formation is probably Llandoverian to Ludlovian and consists of several hundred feet of shale, slate, fine-grained sandstone, and an intercalated 2½-foot fossiliferous oölitic hematite bed like that of the McAdam Formation. In Cobequid Mountains farther west, several thousand feet of Silurian strata similar to the upper part of the Arisaig Group is underlain by silicic and mafic volcanic rocks.

On southeastern Cape Breton Island the unfossiliferous Middle River Group (Weeks, 1954) is probably Silurian and early Devonian and lies with angular unconformity on the Cambrian Kelvin Glen and Bourinot Groups. It consists of an unknown thickness of cross-bedded feldspathic sandstone, quartzite, and conglomerate. The conglomerate contains quartz and quartzite pebbles and volcanic cobbles derived from the Bourinot Group probably to the north and west.

In southeastern Newfoundland, as in southern New Brunswick, thick Silurian volcanic and sedimentary rocks of Notre Dame Trough rest on the northwestern part of Avalon Platform. Inasmuch as no other Silurian rocks of either geosynclinal or platformal aspect occur elsewhere on the platform, their existence and extent during the Silurian cannot be appraised.

Meguma Trough. In western Nova Scotia (Taylor, 1965; Crosby, 1962) Silurian rocks comprise three conformable formations. The upper two, the Kentville and New Canaan Formations, are largely if not entirely Ludlovian, and the lower one, the White Rock Formation, is undated. The entire assemblage consists of quartzite, slate, and volcanic rocks that thicken from about 2,000 feet in the northeast to nearly 20,000 feet in the southwest. Apart from the volcanic rocks and conglomerates, the strata represent the upper part of a clastic sequence that is probably a nearly continuous deposition from the Cambrian into the Devonian. The White Rock Formation increases in thickness from 100 to 500 feet near Wolfville to over 15,000 feet near Yarmouth. It lies conformably on the Halifax Formation except at Cape St. Mary where the basal White Rock rhyolite lies with an angular unconformity of 10 to 20 degrees on Halifax slate. This relationship is believed to be of local significance only. Near Wolfville the White Rock Formation consists of two massive orthoquartzite beds separated by grey slate and siltstone lithologically indistinguishable from the underlying Halifax Formation. It grades southwestward into an assemblage of andesite lava and tuff, slate, quartzite, conglomerate, rhyolite, and greywacke. Clasts in the conglomerates are pebbles or boulders as much as 2 feet across mainly of quartzite with minor quartz and rhyolite. The Ludlovian Kentville Formation conformably overlies the White Rock Formation, but in westernmost Nova Scotia it has not been recognized and equivalent strata may be included in the White Rock Formation. The Kentville varies in thickness from 1,600 feet near Wolfville to 3,500 feet in the southwest in the Nictaux–Torbrook area; it comprises grey siltstone and slate with minor argillaceous quartzite. The conformably overlying Ludlovian New Canaan Formation occurs only near Wolfville and consists of more than 1,000 feet of marine sedimentary andesite-breccia and minor siltstone, slate, and andesite flows. The top is not exposed.

St. Lawrence Platform

Southwestern Ontario. Alexandrian and early Niagaran rocks that form the Cataract Group of southwestern Ontario reflect a transgressive–regressive cycle of sedimentation (Sanford, 1964). They are 500 feet thick in Michigan Basin, thinning to 80 feet over Algonquin Arch (Fig. VI-16), and thickening to 175 feet in northwestern Allegheny Trough beneath eastern Lake Erie. The transgressive phase consists of tan and blue-grey, fine- to medium-crystalline dolomite of the Manitoulin Formation which is 80 feet thick bordering Michigan Basin. It thins southeast toward Niagara Peninsula and grades into the Whirlpool Formation—as much as 25 feet of white and grey, fine-grained, orthoquartzitic sandstone. As the

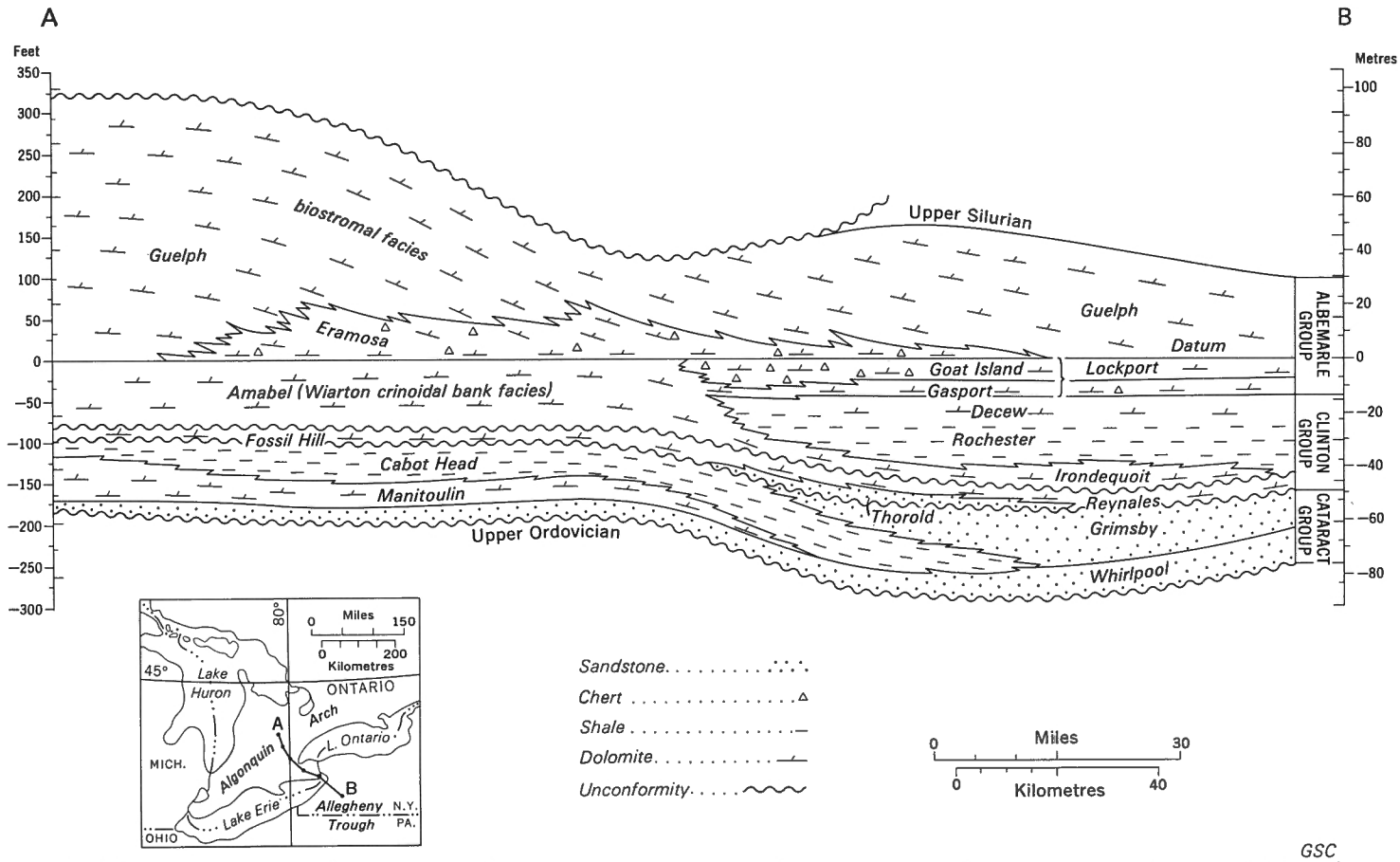


FIGURE VI-16. Cross-section of Lower and Middle Silurian strata, Algonquin Arch to northwestern Allegheny Trough, southwestern Ontario and New York (by B. V. Sanford).

Cataract sea regressed to the northwest, a time–transgressive deltaic red bed sequence was deposited (represented by the Grimsby Formation) that consists of protoquartzitic sandstone and siltstone, and the intertonguing more basinal grey-green and locally red marine shale of the Cabot Head Formation. The Grimsby thins northwestward from a maximum thickness of 80 feet beneath eastern Lake Erie, whereas the partly equivalent and underlying Cabot Head shale thickens northwestward to a maximum of 120 feet near Lake Huron. Around the margin of Michigan Basin (Fig. VI-17) the Cabot Head grades into the Dyer Bay Formation, 20 feet of fine- to medium-crystalline blue-grey mottled dolomite. Conformably succeeding the Dyer Bay is the Wingfield Formation composed of greyish brown and locally red and greyish green shaly aphanitic dolomite, 36 feet thick. It is overlain by the St. Edmund Formation consisting of greyish brown aphanitic dolomite varying in thickness from 12 feet in Bruce Peninsula to 90 feet in Manitoulin Island and the northern peninsula of Michigan.

Detritus in the Whirlpool and Grimsby Formations was derived from the southeast. The partly equivalent Medina red sandstones in western New York and Pennsylvania lie on Upper Ordovician coarse red beds, but east of the Lower Silurian depositional boundary much of the Upper Ordovician is missing and presumably eroded—the detritus constituting the Lower Silurian clastic beds of Allegheny Trough, Algonquin Arch, and southeastern Michigan Basin.

Middle and late Niagaran deposits comprising crinoid bank and complex biostromal barrier reefs constitute the Clinton and Albemarle Groups (Sanford, 1964). They formed on the shallow margins of the rapidly subsiding Michigan Basin. In Ontario these deposits form Manitoulin Island, Bruce Peninsula, and extend southeastward onto Algonquin Arch and into the subsurface beneath central and western Lake Erie (Fig. VI-16). Along the barrier complex, clean carbonate rocks of shallow marine origin, 450 to 550 feet thick, grade westward into basinal carbonates that thin to less than 100 feet in central Michigan Basin, and eastward into basinal carbonates and shale, 200 feet thick in northwestern Allegheny Trough beneath eastern Lake Erie. The Fossil Hill Formation of the Clinton Group consists of brown, fine- to medium-crystalline dolomite locally characterized by coral biostromes and massive lenses containing abundant *Pentamerus oblongus*. From a maximum thickness in Ontario of 80 feet in southern Bruce Peninsula, it thins

northwestward to 40 feet on Manitoulin Island and southeastward to 8 feet over Algonquin Arch. Southeast of the arch, the Fossil Hill grades into as much as 22 feet of fine-grained, orthoquartzitic sandstone of the Thorold Formation, 7 feet of grey fissile shale of the Neahga Formation, and 8 to 22 feet of brown, fine- to medium-grained dolomite of the Reynales Formation.

After deposition of the Fossil Hill and equivalent strata, the sea regressed from Michigan Basin and Algonquin Arch. Major erosion occurred on the arch during the hiatus. As the Niagaran sea advanced again into Ontario and Michigan, a variety of facies were produced, the most persistent of which is the white to bluish grey and beige, coarsely crystalline crinoidal dolomite of the Colpoy Bay and Wiarton Members of the Amabel Formation bordering Michigan Basin. The members are 150 feet thick on the crest of Algonquin Arch and intertongue with grey, cherty, fine-grained dolomite of the Lions Head Member in central Michigan Basin. Equivalent rocks in Niagara Peninsula and beneath eastern Lake Erie are the Irondequoit white to grey, crinoidal dolomite, 10 to 20 feet thick; the Rochester grey silty and dolomitic shales, 80 feet thick; the Decew grey argillaceous dolomite, 30 feet thick; and the Gasport and Goat Island Members of the Lockport Formation, white to grey crinoidal dolomite succeeded by grey, cherty dolomite, 60 feet thick (Pl. VI-9). Lateral migration of the Amabel crinoid bank facies occurred from time to time, and accounts for the presence of local wedges of coarse crinoidal deposits constituting the Irondequoit and Gasport Members within an otherwise basinal sequence. The Irondequoit is confined to Niagara Peninsula and eastern Lake Erie, but the Gasport is widely distributed throughout southwestern Ontario and reaches its maximum thickness of 150 feet near Lake St. Clair. During the final phase of Amabel–Lockport sedimentation the structural centre of Michigan Basin shifted southward. In Ontario this resulted in a southwestward migration of the Wiarton crinoid bank from the Algonquin Arch along a broad arc to central Lake Erie, and thence westward into northwestern Ohio to the newly established southern rim of Michigan Basin. The grey, fine-grained dolomite facies of the Lions Head also migrated southward, transgressing in time, to form the basinal facies of the carbonate bank throughout southern Michigan Basin. There, and in Niagara Peninsula and beneath eastern Lake Erie, these strata are known as the Goat Island Member of the Lockport Formation.

The youngest and most distinctive Niagaran rocks are the carbonates of the Guelph Formation, for they contain the major bioherm reef development of southwestern Ontario and Michigan. Consisting partly of a biostromal complex superimposed on the subjacent carbonate bank deposits of the Amabel Formation, the Guelph formed a fringing barrier reef that completely enveloped the rapidly subsiding Michigan Basin (Fig. VII-26). Cream to white, coarsely crystalline dolomite, 327 feet thick, forms the barrier reef that grades basinwards into inter-reef facies composed of brown and black, bituminous and cherty dolomite. Adjacent to the barrier reef in Michigan Basin are large bioherm patch reefs with a vertical relief of as much as 160 feet and a base several thousand acres in extent. Toward the centre of the basin, where subsidence was more rapid, small pinnacle reefs developed with a relief of 300 to 550 feet and a base of 300 to 500 acres. Although a few scattered bioherms occur within Guelph strata on the barrier and along its southeast margin in

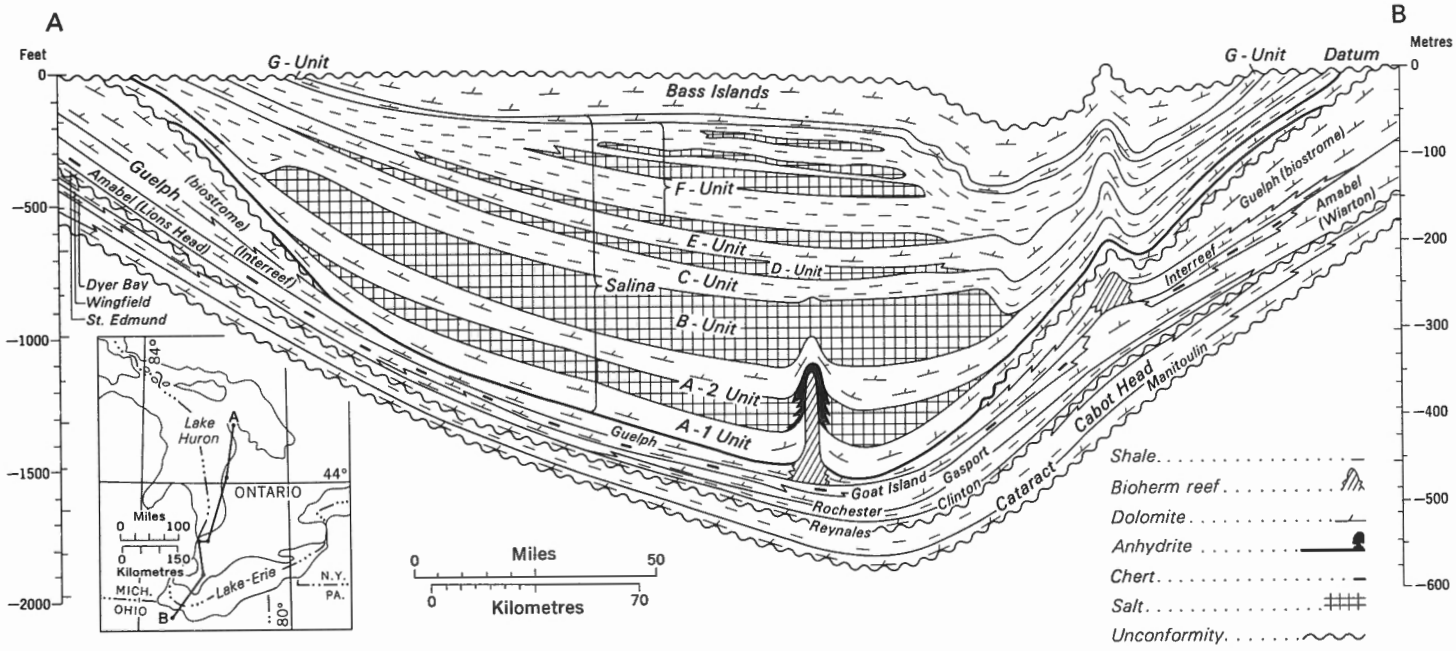


FIGURE VI-17. Cross-section of Silurian strata, eastern flank of Michigan Basin, southwestern Ontario (by B. V. Sanford).

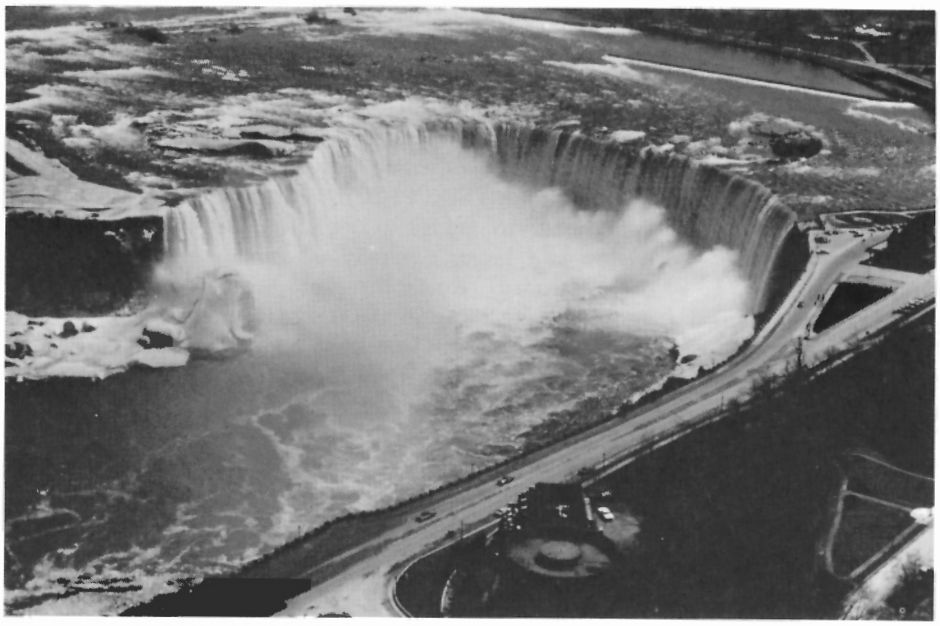


PLATE VI-9
Niagara Falls, Ontario. Falls are formed by resistant dolomites of the Silurian Lockport Formation overlying the softer Clinton and Cataract shales and sandstones and the Upper Ordovician Queenston red shales.

northwestern Allegheny Trough, nowhere in these regions are they known to have more than a few feet of vertical closure, and in general they reflect lack of continuous subsidence.

The Upper Silurian (Cayugan) Series of southwestern Ontario contains a wide variety of evaporite, carbonate, and clastic facies that comprise the Salina and Bass Islands (Bertie) Formations (Sanford, 1964). They have a combined maximum thickness of 1,700 feet in Lambton County near southern Lake Huron (Fig. VI-17) and thin eastward to about 345 feet on Algonquin Arch. The Salina Formation, with seven units designated from A to G in ascending sequence, consists of dolomite, limestone, shale, salt, and anhydrite. Where it has a maximum thickness of 1,340 feet near southernmost Lake Huron, an aggregate of 733 feet consists of salt. The salt beds occur within units designated A, B, D, and F; these gradually wedge out one by one toward the edge of Michigan Basin and grade into anhydrite, dolomite, and shale. The reconstructed depositional edge of Salina salts bordering the Michigan Basin roughly coincides with the margin of the subjacent Guelph barrier reef (Koepke and Sanford, 1965). The gradual migration of the salt edge to its present position much lower in the basin is the result of leaching during post-Silurian time. Completing the Cayugan sequence in Michigan Basin is the Bass Islands Formation consisting of cream to tan, finely sucrosic dolomite as thick as 300 feet, and the equivalent Bertie Formation, 30 to 100 feet thick, in Niagara Peninsula and eastern Lake Erie, composed of dark brown, aphanitic argillaceous dolomite.

Lake Timiskaming outlier. The Alexandrian and early Niagaran Wabi Group (Ollerenshaw and MacQueen, 1960) rests disconformably on early Upper Ordovician (Edenian) Dawson Point Formation and consists of 152 feet of impure carbonate rocks and shale that are lithologically similar to the formations that comprise the Cataract Group of Bruce Peninsula and Manitoulin Island and those names are applied accordingly. The lowest unit, the Manitoulin Formation, is 54 feet of grey limestone and dolomite with a shaly and sandy zone at the base. The overlying Cabot Head Formation comprises 33 feet of red and green shale containing anhydrite and gypsum. The shale grades upward into 25 feet of grey and blue-grey, medium-crystalline limestone and dolomite of the Dyer Bay Formation. The Dyer Bay Formation is succeeded by the St. Edmund Formation, 40 feet of light greenish grey aphanitic dolomite with interbeds of green and red nodular shale. The Thornloe Formation, the youngest Paleozoic strata at Lake Timiskaming, is middle Niagaran. It rests disconformably on the Wabi Group and consists of 196 feet of grey-buff to light brown, fine- to medium-crystalline limestone and dolomite with a foot of limestone-conglomerate at the base. The Thornloe is lithologically similar to and coeval with the Fossil Hill Formation of southern Ontario.

Anticosti Basin (Fig. VI-7). On Anticosti Island (Bolton, 1961; Roliff, 1968) rocks equivalent to the Alexandrian and early Niagaran Cataract and Wabi Groups of Ontario are represented by the Becscie and Gun River Formations which reach a combined thickness of about 600 feet. The Becscie, 265 feet thick, overlies the Upper Ordovician Ellis Bay conformably and consists of a lower whitish grey to blue-grey, finely crystalline limestone grading upward to green shale with nodular limestone interbeds. The succeeding Gun River, 308–343 feet thick, consists of ash-grey to yellowish white limestone alternating with shale in the upper part. Ripple-marks and intraforma-

tional conglomerate in both formations indicate shallow-water depositional environments.

The Niagaran Jupiter and Chicotte Formations, possibly equivalent to Clinton and Amabel strata of southwestern Ontario, are the youngest Paleozoic strata exposed and have a combined thickness of about 725 feet. The Jupiter Formation, 650 feet thick, consists of ash-grey finely crystalline limestone with thin, grey shale interbeds grading upward to light green and grey calcareous shale and argillaceous limestone. The Chicotte Formation, 73 feet thick, rests with abrupt contact on the Jupiter. It is composed of bioclastic limestone with skeletal fragments of crinoid columns and corals, and resembles the Wiarton Member of the Amabel Formation of Ontario.

On Port au Port Peninsula, Newfoundland, the basal beds of the Clam Bank Formation may be Silurian (Rodgers, 1965). They lie with structural conformity on the Middle Ordovician Long Point Formation. The Clam Bank consists of 1,500 feet of red sandstone, siltstone, shale, and some calcareous beds with Pridoli fossils in the upper part.

Devonian

Tectonic Summary

Lower Devonian sedimentation and volcanism in the Appalachian Geosyncline and St. Lawrence Platform reflect a continuation of Silurian depositional patterns, facies, and tectonic elements. In the geosyncline they were profoundly changed by the Acadian Orogeny during the Middle and Late Devonian. During the Late Devonian new tectonic elements were initiated, and terrestrial sedimentation began in the geosyncline that continued into the Carboniferous. Fine clastic sediments derived from the Acadian mountains spread across the western St. Lawrence Platform.

During the Early and early Middle Devonian (Figs. VI-18, 19) thick marine sediments and volcanics, passing upward into non-marine beds, were deposited conformably on the Silurian in Gaspé and Fredericton Troughs. The troughs were bounded by Quebec and Miramichi Geanticlines and Avalon Platform. Devonian volcanism occurred in the same regions as Silurian volcanism. Limestone was deposited in northwestern Gaspé Trough grading southeastward into shale, sandstone, and volcanic rocks of central and southeastern Gaspé Trough. Non-marine sandstone and conglomerate followed in the late Early and early Middle Devonian, perhaps reflecting the early Acadian orogenic phases in Miramichi Geanticline. On Avalon Platform in Nova Scotia deposition of marine siltstone was followed by non-marine red beds, and in the Middle Devonian by volcanic rocks, red sandstone, and conglomerate. In Meguma Trough marine shale was deposited with some limestone and oölitic iron-formation. In Notre Dame Trough in Newfoundland, shale and siltstone, largely non-marine and several thousand feet thick, were deposited in the southwestern part in a tectonic position comparable to the thick, non-marine sandstone sequence in northwestern Gaspé Trough. Rhyolite was apparently extruded in southeastern Notre Dame Trough.

During the Acadian Orogeny almost the entire Appalachian Geosyncline was deformed into northeasterly trending folds (Fig. VI-20). Parts were regionally metamorphosed to grades as high as almandine amphibolite facies, and large, generally discordant granite batholiths were intruded. Rocks that were not deformed during the orogeny were those of the Taconian-deformed northwestern part of Quebec Geanticline and a small part of Avalon Platform in eastern Newfoundland. The rocks were thrust northwestward and northward along Logan's Line onto St. Lawrence Platform. The entire belt, probably including the site of the present continental shelf, was uplifted, and the mountains underwent erosion.

Upper Devonian synorogenic and post-orogenic, non-marine red sandstone, conglomerate, and shale were deposited in local basins on the Acadian Orogen as the basal deposits of Fundy Epieugeosyncline. Deposition within the epieugeosyncline, which was bordered by narrow platforms, continued into the Carboniferous and was controlled largely by fault movements attributed to the Maritime Disturbance, the final phase of the Appalachian orogenies.

On western St. Lawrence Platform the Upper Silurian evaporite basins were emergent in the early Early Devonian. Subsequently, a thin, marine, Siegenian orthoquartzite was deposited in southwestern Ontario and Michigan but was largely removed by erosion as a result of epeirogenic uplift. Lower Devonian limestone occurs as blocks in a Cretaceous diatreme breccia near Montreal, but probably once extended throughout Quebec Basin. Late Lower and early Middle Devonian limestone and dolomite were deposited on the western part of the platform; from Algonquin Arch, the strata thicken northwestward into Michigan Basin and southeastward into Allegheny Trough. During the late Middle and the Late Devonian, coarse red beds of the Catskill Delta, derived from the Acadian mountains, spread northwestward into Allegheny Trough, grading into marine shale and siltstone on Algonquin Arch and to shale and limestone in western Michigan Basin. Devonian strata of southwestern Ontario are lithologically and faunally similar to those of Hudson Platform, suggesting that the seas extended through the intervening region. Devonian strata are not present in central and eastern St. Lawrence Platform except locally in southwestern Newfoundland. A coarse red bed sequence is partly Lower Devonian and was derived from mountains to the east.

Lower and Early Middle Devonian

Appalachian Geosyncline (Figs. VI-18, 19). In northwestern Gaspé Trough in southern Quebec, Lower and early Middle Devonian sedimentary and volcanic rocks lie

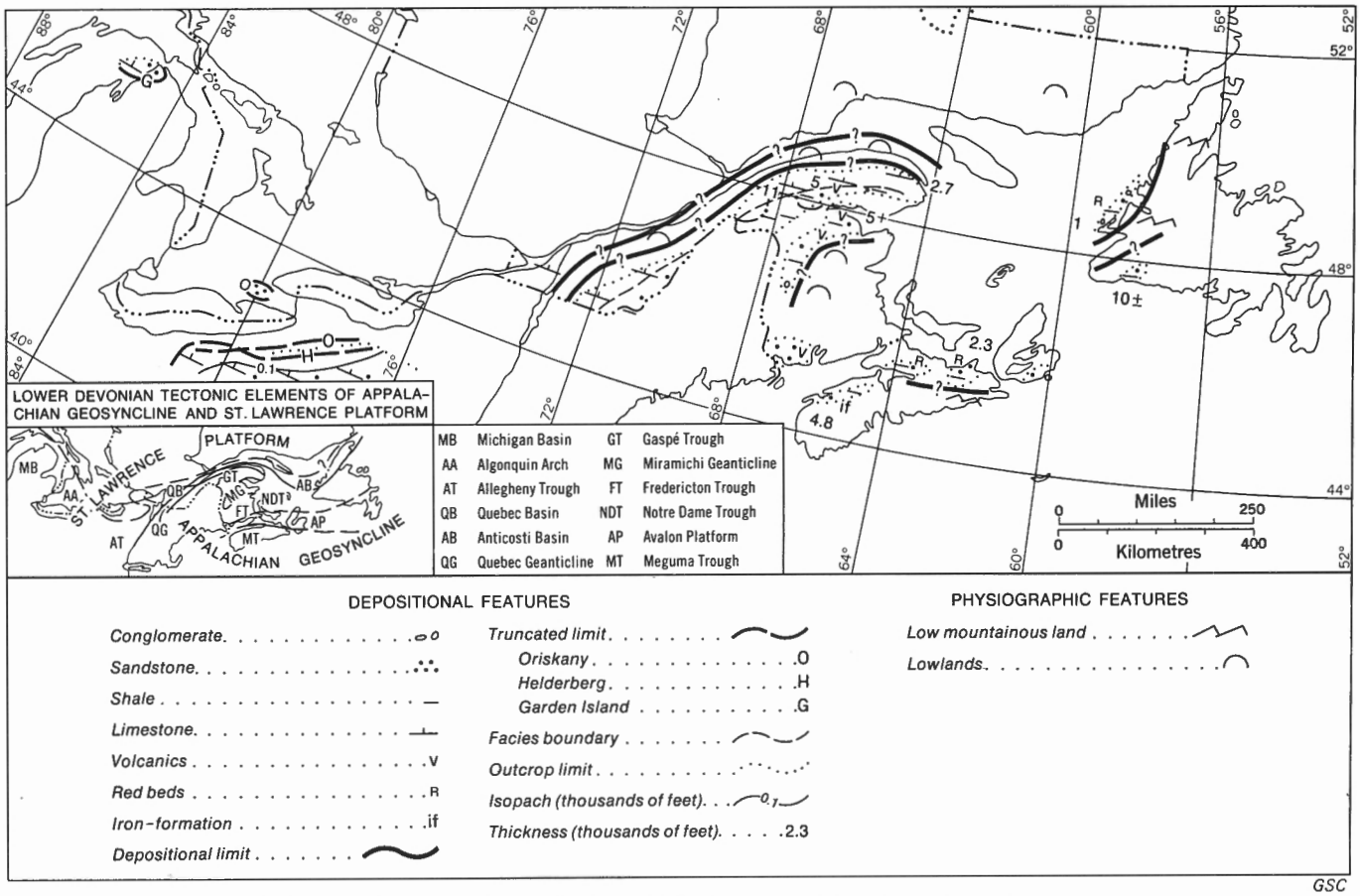


FIGURE VI-18. Early Lower Devonian sedimentation, volcanism, and tectonism in Southeastern Canada.

conformably on Upper Silurian strata, or locally overlap northwestward to rest directly with angular unconformity on the Cambrian and Ordovician of Quebec Geanticline. In the Eastern Townships the similar and laterally equivalent St. Francis and St. Juste Groups consist of an unknown thickness of thin-bedded limestone and calcareous slate with sandstone interbeds, apparently conformably overlain by thin-bedded dark grey slate and sandstone. The lower, calcareous strata are partly Upper Silurian. The base of the sequence is commonly a polymictic conglomerate resting with angular unconformity on Ordovician and older rocks. In one locality at least, a 50-foot conglomerate is apparently Lower Devonian. Thus the base of the sequence varies from perhaps Middle Silurian to Lower Devonian. To the southeast, along the International Boundary, the calcareous strata are missing, and Lower Devonian Seboomook and Compton Formations of thin-bedded dark grey pelites and sandstone have apparently overlapped the older strata in the subsurface to rest on probable Ordovician rocks of a small geanticlinal area. The possibility exists that some of the pelitic and sandy strata may be in part a southeastern facies equivalent to the upper part of the calcareous rocks. The Frontenac Formation overlies the Seboomook–Compton strata and is the youngest formation in Eastern Townships. It consists of dark grey slate and sandstone, and altered basic volcanic rocks many of which are pillowed flows.

In Gaspé Peninsula, northwesternmost New Brunswick, and Témiscouata, (McGerrigle, 1950; Béland, 1960; Lajoie, *et al.*, 1968) Lower and early Middle Devonian rocks consist of the Gaspé Limestone, the conformably overlying Gaspé Sandstone, and the pelitic and sandy Fortin and Témiscouata Formations that lie to the southeast and into which the limestone and lowermost part of the sandstone grade laterally. In the northern half of the belt Ludlovian St. Léon calcareous siltstones grade upward into Lower Devonian limestone of the Cape Bon Ami Formation. The contact may be time–transgressive, but it lies at the base of either the Gedinnian or the Pridoli. The Cape Bon Ami Formation consists mainly of dark grey argillaceous and silty limestone interbedded with dark grey limestone and calcareous shale. On Forillon Peninsula it is 1,100 feet thick and thickens to the west and south; it is from 2,000 to 3,000 feet thick along the northern edge as far west as Témiscouata. Southward into the Gaspé Trough it thickens to 3,500 feet and possibly to a maximum of 6,000 feet in parts of east-central Gaspé. On Forillon Peninsula (Cumming, 1959) the Cape Bon Ami is underlain by the Gedinnian St. Alban Formation and the Pridoli Griffon Cove Formation,

both of which apparently are correlatives of lower parts of Cape Bon Ami elsewhere. The gently dipping Griffon Cove lies unconformably on isoclinally folded Lower Ordovician rocks. It is 400 feet thick and eastward towards the coast wedges out to zero. It consists of quartz-pebble conglomerate and red and green shale passing upward into limestone that in turn grades upward into the St. Alban Formation of 345 feet of grey argillaceous limestone and greenish grey and maroon calcareous shale. The St. Alban overlaps the Griffon Cove beds to lie unconformably on the Ordovician rocks, and is gradationally overlain by the Cape Bon Ami Formation. In the west near Témiscouata, about 1,200 feet of sandstone, limestone, and minor dolomite make up the Lac Croche Formation of Pridoli age which there lies at the northwestern limit of Pridoli and Lower Devonian strata and is a more sandy correlative of the lower part of the Cape Bon Ami to the south and east.

The Grande Grève Formation conformably overlies the Cape Bon Ami and consists mainly of well-bedded, dark to light grey, hard siliceous and cherty limestone, calcareous siltstone, and minor sandstone. Silica occurs as chert nodules, chert beds, and in siliceous limestone interbedded with purer limestone. On Forillon Peninsula the formation is about 900 feet thick and thickens westward to 2,000 feet for some distance and then to a possible 9,000 feet northeast of Témiscouata Lake. In central Gaspé Peninsula the formation is about 3,000 feet thick. Oil occurs locally in cracks and vugs. A dozen thin intercalated beds of bentonite contain sanidine and biotite that yielded K–Ar dates of 388 and 382 m.y., respectively.

The Gaspé Sandstone (McGerrigle, 1950; Boucot, *et al.*, 1967) lies conformably on the Gaspé Limestone; it is composed mainly of three, dominantly non-marine, gradationally conformable formations in ascending order: York River, Battery Point, and Malbaie. The three formations were originally described in eastern Gaspé Peninsula, but only the lower two have been recognized in central and western Gaspé. The York River and Battery Point Formations are distinguished largely by the colour of strata and detrital feldspars. Variations in lithology led to definition of other formations in central and western Gaspé, and it seems probable that the boundaries are diachronous.

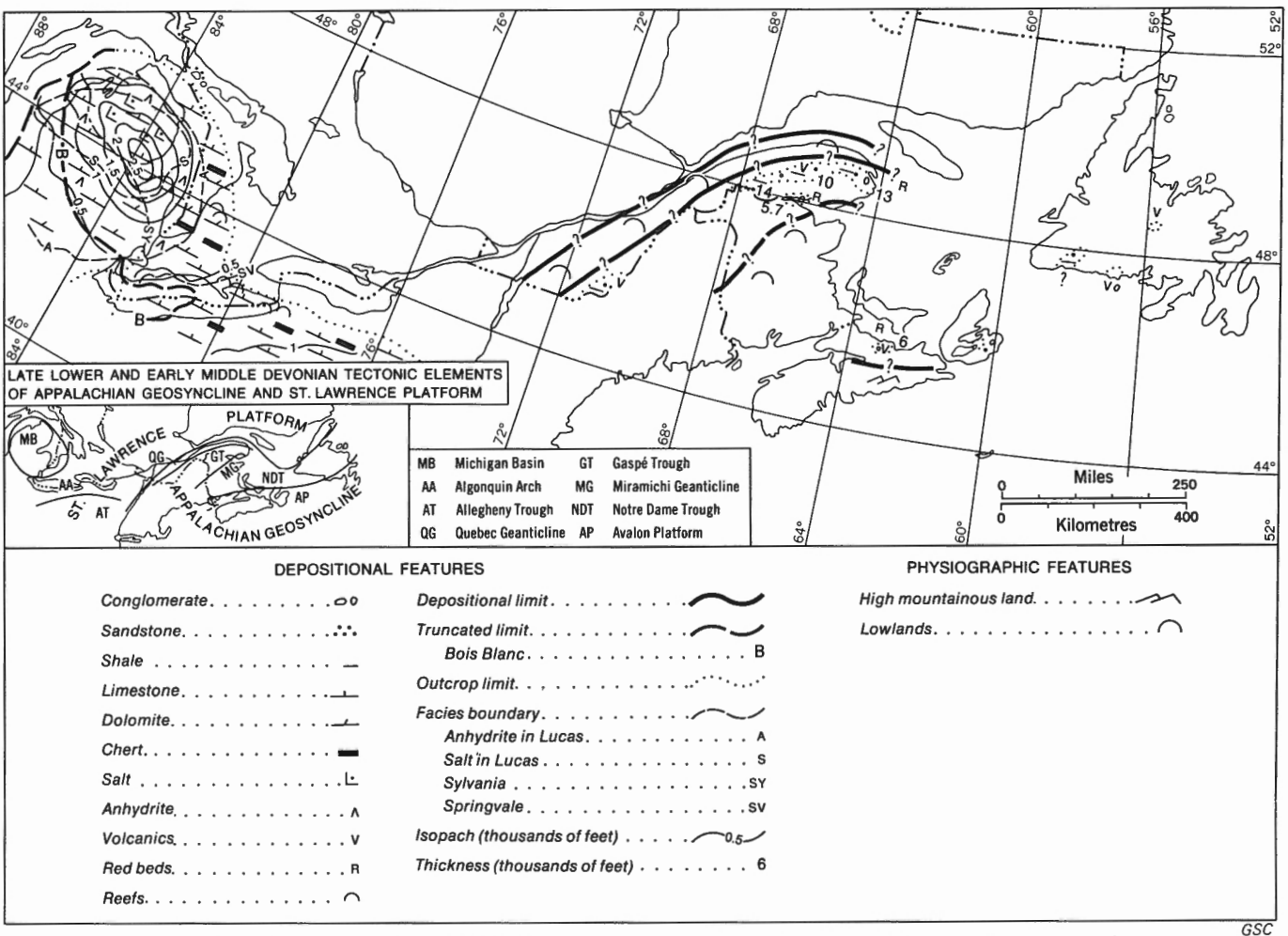


FIGURE VI-19. Late Lower and early Middle Devonian sedimentation, volcanism, and tectonism in Southeastern Canada.

The York River Formation consists of green to dark grey feldspathic sandstone and shales containing plant detritus and interbeds of calcareous sandstone and sandy limestone with a rich marine shelly fauna. The formation is 1,300 feet thick in northeasternmost Gaspé and thickens southward along the coastal region to about 6,000 feet; in western Gaspé it is 14,000 feet or more thick. The northern limit of the York River Formation in central Gaspé has a basal transitional facies, some 2,000 to 4,000 feet thick, of interbedded sandstone and shale of the York River type and limestone of the underlying Grande Grève type; this succession has been referred to as the York Lake Formation. In western Gaspé, sandstone beds similar and equivalent to the York River Formation were named the Heppel Formation; the upper part is a red sandstone facies called the Lake Branch Formation, about 4,000 to 5,000 feet thick. In central Gaspé, flows, tuffs, and breccias are intercalated in the York River Formation and its equivalents at horizons within the formation that vary from place to place. Locally, volcanic sequences are 2,000 feet thick and more. Most of the rocks are andesite and basalt, generally porphyritic (feldspar) and amygdaloidal; others are rhyolite and porphyritic rhyolite containing phenocrysts of quartz or feldspar.

The Battery Point Formation consists of a lower part of greenish grey feldspathic sandstone and an upper part of red and brown feldspathic sandstone, shale, and conglomerate. It contains a sparse marine fauna and abundant land plants in the lower part. Most detrital feldspar is pink, and like the York River consists of both potash feldspar and plagioclase. Clasts in the Battery Point consist of quartz, a variety of rocks, granite, and syenite. Ripple-marks and crossbeds are common in both formations. The Battery Point is 5,000 to 7,000 feet thick in eastern Gaspé and about 10,000 feet thick in central Gaspé.

The Malbaie Formation, the youngest unit of the Gaspé Sandstone, occurs only in easternmost Gaspé Peninsula. It comprises some 2,000 feet of grey to reddish brown pebble-conglomerate, sandstone, and minor green and red shale. Pebbles are mainly of limestone of the Ordovician and Silurian White Head and Lower Devonian Grande Grève Formations and sandstone of the underlying Battery Point Formation; other pebbles consist of quartz, chert, volcanics and rarely of granite, syenite, and gneiss. The Malbaie has been faulted and gently folded, and is unconformably overlain by the Carboniferous Bonaventure conglomerate. The entire Gaspé Sandstone succession comprises over 10,000 feet of clastic strata that were deposited in a narrow trough during a very short interval of time, perhaps as little as 10 to 15 million years. The trough probably developed as the area to the south, the Miramichi Geanticline, became uplifted during the early stage of Acadian Orogeny. The detritus was probably derived from the southern uplifted areas, but in the York River beds in central and northern Gaspé southwestward and westward longitudinal current flow is indicated by crossbedding. The Gaspé Sandstone records an increase in size of detritus; a decrease in abundance of marine fauna; an increase in abundance of land plants and spores to the middle of the Battery Point; a general change of strata colour from grey and dark grey to green, red, and reddish brown; and a change from mixed non-marine–marine strata to entirely non-marine continental strata. Oil seeps are common in the York River and Battery Point beds.

Between eastern Gaspé Peninsula and northwestern New Brunswick thin-bedded grey siltstone, slate and minor sandstone, limestone, and conglomerate constitute the Lower Devonian Fortin and Témiscouata Formations. These rocks form a belt that continues southwesterly through northern Maine into the southeast part of the Eastern Townships where they are included in the Seboomook Formation, St. Juste and St. Francis Groups, previously described. The Fortin and Témiscouata Formations represent a deeper water marine facies equivalent to the Gaspé Limestone and probably the lower part of the York River Formation. The facies belt lay between the carbonates to the north and the volcanic belt, including the Dalhousie Formation, to the south. In eastern Gaspé the Fortin Formation apparently pinches out northward between the underlying Grande Grève limestones and the overlying York River sandstones. The basal few feet of the Fortin is a conglomerate containing clasts of Grande Grève limestone. At its easternmost exposures the Fortin is about 5,000 feet thick; it is sandy and grades imperceptibly upward into the overlying York River Formation. Farther west, in east-central Gaspé, the Fortin Formation consists of an uncertain thickness, perhaps several thousand feet, of dark grey silty and fine sandy limestone and slate, minor feldspathic sandstone, and conglomerate. Much of the strata resembles the dark grey limestone, siltstone, and slate of the Cape Bon Ami Formation. In this region it rests on the Silurian and Lower Devonian Mount Alexander Group. Still farther west, in southwestern Gaspé and northern New Brunswick, dark grey slate, siltstone, and minor sandstone become dominant lithologies, and the formation is bounded by strike faults. Andesite and breccia of unknown thickness are intercalated in the Fortin Formation along the north fault-boundary in western Gaspé. In northwestern New Brunswick the formation rests on Upper Silurian strata; the basal strata are plant-bearing sandstones. E. Mencher reported that farther west, in northern Maine, the rocks lie with regional unconformity on Ordovician and Silurian rocks. In the Témiscouata region of adjacent parts of Quebec the Fortin-type rocks are referred to the Témiscouata Formation.

Lower Devonian sedimentary and volcanic strata in the southeastern part of Gaspé Trough lie conformably on the Silurian. Positively dated early Middle Devonian is lacking. In northern New Brunswick, at Dalhousie (Alcock, 1936; Greiner and Potter, 1966), the Dalhousie

Formation consists of an estimated 4,600 feet of interbedded sedimentary and volcanic rocks; among these the character and proportions of the rock types vary from place to place. A lower division is mainly of sediments, and a thick upper division is mainly of volcanic rocks. The sediments contain a rich marine fauna, now considered to be late Gedinnian (New Scotland). Sedimentary rocks consist of greenish calcareous and sandy shale, siltstone, and grey and buff shaly limestone, and boulder-conglomerate rich in volcanic clasts. Volcanic rocks are basalt and andesite flows, in part amygdaloidal and porphyritic (feldspar), breccias, tuffs, and palagonite tuff. The coarse nature of some volcanic breccias, in which some fragments are over 2 feet long, suggests that the volcanic vents were nearby. Similar rocks, but with different proportions of sediments, and volcanic rocks occur as a belt several miles wide and 100 miles long bordering Miramichi Geanticline on the northwest. Rhyolitic and trachytic flows and pyroclastic rocks accompany the volcanics. The sedimentary rocks are more pelitic and sandy, in places contain fine carbonaceous plant debris, and lack the limestones and abundant calcareous cement characteristic of the Dalhousie succession. Plant-bearing conglomerate and sandstone occur in one place at the southwestern end of the belt. During early Devonian the geanticline had a subdued topography and possibly was only partly emergent.

At the head of Chaleur Bay (Alcock, 1936) grey and green sandstone, conglomerate, and mudstone with a thin coal seam, probably several thousand feet thick and tentatively correlated with the Gaspé Sandstone, overlie the Dalhousie Formation probably disconformably. Clasts of mafic and silicic volcanic rocks probably derived from the Dalhousie are common. Plant fossils, fossil fish, and some marine invertebrates have been recovered. The conformably overlying Pirate Cove Formation is composed of 660 feet of non-marine red and green conglomerate and shale. These strata lack megafossils but contain spores similar to the Battery Point Formation of late Emsian and possibly younger age. The clasts, as large as a foot in diameter, consist of Matapedia limestone, Silurian sandstone and limestone, and minor igneous rocks. Coarse crossbeds and ripple-marks indicate current flow from the northwest, and together with the lithology and the occurrence of mudcracks, suggest an alluvial fan or flood plain environment of deposition.

In southeastern Fredericton Trough, along the coast of southern New Brunswick, early Lower Devonian rocks have been recognized but not clearly defined (Smith, 1966). They form the upper part of the mainly Silurian Mascarene Group, and consist of an unknown thickness of grey to brown sandstone, siltstone, and argillite with minor red shale and basalt; these strata are cut by rhyolite and andesite plugs of similar age. Deformed black shale, sandstone, and volcanic-breccia conglomerate contain plant fossils. Some 30 miles northwest of the Mascarene volcanic rocks are thin-bedded, greenish grey sandstones, siltstone, and slate, with abundant detrital muscovite, forming part of the Pale Argillite Division of the Charlotte Group. The presence of undated plant fossils in pregranite rocks suggests Lower or Middle Devonian age.

In Notre Dame Trough in Newfoundland, Lower and early Middle Devonian rocks definitely occur at one locality and possibly at two others; almost everywhere the youngest pre-Acadian stratified rocks are Wenlockian terrestrial sandstones. At La Poile Bay (Cooper, 1954), the Bay du Nord Group comprises an estimated 10,000 feet or more of slate, quartzite, conglomerate, and minor limestone. It passes laterally into schist and gneiss and is cut by Devonian granites. Relationship to the nearby, undated, possibly Silurian La Poile Group is obscure. The Bay du Nord contains plant fossils. The plants and the limestone suggest mixed marine and non-marine environments of deposition. The group may be equivalent to the Gaspé Sandstone assemblage: the two groups are probably the same age and both lie near the northwest boundary of the Notre Dame Trough, consist predominantly of a very thick clastic assemblage, and are marine and non-marine. In central and southern Newfoundland there is several hundred feet of flow-banded rhyolite tuff and agglomerate. At Cape la Hune on the south coast 50 feet of boulder-conglomerate and arkosic sandstone underlies the rhyolites and contains clasts of granite, gneiss, diorite, and other rocks; underlying rocks are not exposed. Rhyolites and conglomerates are cut by mid-Devonian Ackley-type granite. In central Newfoundland, near Stoney Lake, the rhyolites apparently overlie folded sandstones of the Botwood Group, but the contact is not exposed and the structural relations are uncertain. The Cape la Hune and Stoney Lake rocks are tentatively correlated, despite the lack of fossils, and are presumed to be Lower Devonian.

On the Avalon Platform, Lower and early Middle Devonian strata of a platformal nature occur in northern Nova Scotia and on Cape Breton Island. None has been recognized in southeastern Newfoundland and southern New Brunswick. In northern Nova Scotia (Williams, 1914) the Stonehouse Formation at the top of the Arisaig Group consists of about 1,300 feet of marine, greenish grey calcareous siltstone, shale, slate, sandstone, and lenses of limestone. The conformably overlying Knoydart Formation of probable late Gedinnian and Siegenian age consists of about 1,000 feet of non-marine, red and green sandstone, siltstone, and shale, locally with limestone nodules, and was deposited in a fluviatile and deltaic environment. The Knoydart contains fragments of fossil fish and some plant material. Current directions on crossbeds indicate flow toward the northwest apparently from an area with Ordovician rocks uplifted during an early stage of the Acadian Orogeny. Rocks similar to the Stonehouse and Knoydart Formations occur in eastern Cobequid Mountains. There, the probable Knoydart is transitionally overlain by andesite flows believed to be correlative with the lower part of the River John Group a few miles to the northeast. The group consists of more

than 6,000 feet of non-marine strata deposited in a fluviatile environment. The lower 1,500 feet comprises mainly grey and red siltstone, sandstone, shale and basalt flows with minor conglomerate, and petroliferous limestone. Albertite veins occur. The succeeding 4,500 feet consists of red-brown conglomerate, sandstone, and siltstone. Spores from the top of the lower part are probably Middle Devonian. Clasts are as much as 5 feet in diameter, and some are fossiliferous limestone derived from the Stonehouse Formation. The River John sandstones contain very little feldspar; the source area evidently did not include much granitic terrane, whereas the sandstone of the unconformably overlying Carboniferous strata contains abundant potash and plagioclase feldspar and quartz–feldspar rock fragments. Like that of the Knoydart, the source of River John detritus probably lay not far to the south, probably the Silurian strata of the eastern Cobequid Mountains.

In southeastern Cape Breton Island (Weeks, 1954) the probable Middle Devonian McAdam Lake Formation rests unconformably on Cambrian strata and possibly on granitic rocks of probable Ordovician age. It consists of an unknown thickness of folded grey conglomerate, arkose, plant-bearing shale, and some tuff, probably deposited under fluviatile conditions. The conglomerate contains pebbles of quartz and sandstone and some boulders of granite, rhyolite, and diorite.

In Meguma Trough the Lower Devonian Torbrook Formation (Smitheringale, 1960) conformably overlies the Ludlovian Kentville Formation in western Nova Scotia and is cut by Devonian granites. The Torbrook consists of about 5,000 feet of slate, siltstone, quartzite, and petroliferous limestone. Intercalated quartzitic iron-formation beds were mined. It has a rich brachiopod fauna. The lowest beds are unfossiliferous, and the uppermost contain some plant debris. The Torbrook represents a normal shallow-water marine deposit, but the iron-formation and plant debris suggest near-shore conditions. Marine deposition in the Meguma Trough ended during the early Devonian, about the same time as in Notre Dame Trough and on Avalon Platform.

St. Lawrence Platform (Figs. VI-18, 19). Lower and early Middle Devonian strata are preserved at three widely separated localities—on St. Lawrence Platform in southwestern Ontario, in a Cretaceous diatreme breccia in Quebec Basin, and in southwest Newfoundland. Most of the platform was probably submerged during much of late Early Devonian and periodically throughout the Middle and Late Devonian as the seas extended across the southern Canadian Shield to connect with those of Hudson Platform.

During the Gedinnian southwestern Ontario was undergoing erosion. Salt was leached from the Upper Silurian Salina Formation. The oldest Devonian rocks of the region, the Siegenian Oriskany Formation, are preserved as erosional remnants on the Niagara Peninsula and beneath adjacent parts of Lake Erie (Sanford, 1968). They consist of light grey, medium- to coarse-grained orthoquartzitic sandstone as much as 20 feet thick disconformably overlying the Upper Silurian Bertie Formation. The orthoquartzitic sandstones in Ontario and western New York were possibly derived from the crystalline rocks of Adirondack Mountains or the Canadian Shield, whereas those in the southeastern Allegheny Trough were derived from Ordovician and Silurian rocks in the Appalachian Geosyncline uplifted during an early phase of the Acadian Orogeny. The Emsian Bois Blanc Formation disconformably overlies the Oriskany or the Upper Silurian Bass Islands and Bertie Formations and is a correlative of the Scoharie Formation in eastern New York State (Oliver, 1966). In the Niagara Peninsula and beneath eastern Lake Erie the Bois Blanc consists of 15 to 40 feet of bluish grey, finely crystalline and silty limestone containing abundant grey nodular chert that in places exceeds the carbonate content. In this region the Bois Blanc commonly has a lower unit, the Springvale Member, consisting of light grey orthoquartzitic sandstones possibly derived in part from the subjacent Oriskany Formation. The sandstone may occur at the base of the Bois Blanc or alternate with cherty limestones at various horizons higher in the formation. The Bois Blanc grades laterally northwestward to grey and tan, medium granular dolomite bordering Lake Huron, where it increases to 150 feet in thickness.

Early Middle Devonian (Eifelian) limestone, dolomite, and minor sandstone disconformably overlie the Bois Blanc Formation. They reach a thickness of 600 feet near southern Lake Huron and thin to 250 feet beneath Lake Erie (Fig. VI-21). The early Eifelian Amherstburg Formation consists of brown, sucrosic, and locally bituminous dolomite as thick as 300 feet. Interdigitating with the lower beds in the Lake St. Clair region are strandline deposits of the Sylvania Member, as much as 80 feet of well-rounded and sorted orthoquartzitic eolian sandstone. The Amherstburg dolomite thins eastward and grades to grey, coarse-grained crinoidal and coral limestone of the Edgecliff Member, which forms the basal 75 to 100 feet of the Onondaga Formation in Niagara Peninsula and beneath eastern Lake Erie. Bioherm reefs occur in the Amherstburg and Edgecliff. Best known is the Formosa reef near Lake Huron; it covers an outcrop area of 65 square miles and has a relief of 90 feet. The Lucas Formation conformably overlies the Amherstburg, and in central Michigan Basin consists of salt and anhydrite that thins and grades laterally southeastward in Ontario, to 300 feet of light tan, finely crystalline dolomite with thin beds of anhydrite near southern Lake Huron, and thence to finely crystalline and sublithographic, high-calcium limestone of the Anderdon Formation bordering northern Lake Erie. During a late Eifelian epeirogenic uplift the Anderdon and much of the Lucas were removed by erosion from Niagara Peninsula and beneath eastern Lake Erie. The succeeding late Eifelian Dundee Formation, 100 feet of coarse crinoidal and finely crystalline

limestone, lies disconformably on the Lucas in Michigan Basin. It grades southeastward across the erosional edge of the Lucas into 200 feet of the brown, aphanitic, and cherty limestone of the Seneca–Moorehouse Members of the Onondaga Formation that rest disconformably on the Edgecliff.

In Quebec Basin, Lower Devonian limestone occurs as blocks in a Cretaceous diatreme breccia related to the Monteregian Intrusions on St. Helen's Island near Montreal. Fossils in the blocks are Gedinnian and Siegenian. The blocks preserved in the breccia provide the only evidence that post-Upper Ordovician seas encroached on Quebec Basin of the St. Lawrence Platform.

In Anticosti Basin (Fig. VI-7) the Clam Bank Formation occurs on the west shore of Port au Port Peninsula and has recently been traced in sporadic subsea outcrops a few miles offshore for about 80 miles to the northeast. The formation consists of about 1,500 feet of mainly terrestrial, coarse-grained and pebbly, crossbedded red sandstone, shale, and minor limestone (Rodgers, 1965). Clasts are mainly quartz and feldspar. The formation lies probably disconformably on the Middle Ordovician Long Point Formation. Brachiopods in the upper Clam Bank are Pridoli; some lower and upper strata may be Ludlovian and Gedinnian. The Clam Bank sediments were probably derived from an uplift to the east formed during an early stage of the Acadian Orogeny. The formation presumably grades westward to shale and limestone beneath the Gulf of St. Lawrence.

Acadian Orogeny

Regional metamorphism. During Acadian Orogeny, in the late Middle and early Late Devonian, most of the pelitic strata that were unmetamorphosed at the onset were regionally metamorphosed to argillite and slate. Other strata were metamorphosed to schist and gneiss of the greenschist and almandine amphibolite facies.

In southern Nova Scotia (Taylor and Schiller, 1966), biotite quartzite, schist, and slate pass southward into schist and metaquartzite containing oligoclase, andesine, almandine, staurolite, and sillimanite. Andalusite and cordierite are also common in these rocks and in pelitic schists in eastern Nova Scotia and may have formed during thermal metamorphism by nearby Devonian granites. In southernmost Nova Scotia, paragneiss and migmatite locally border the granites. Biotite from schist in southernmost Nova Scotia yielded K–Ar dates of 338 and 383

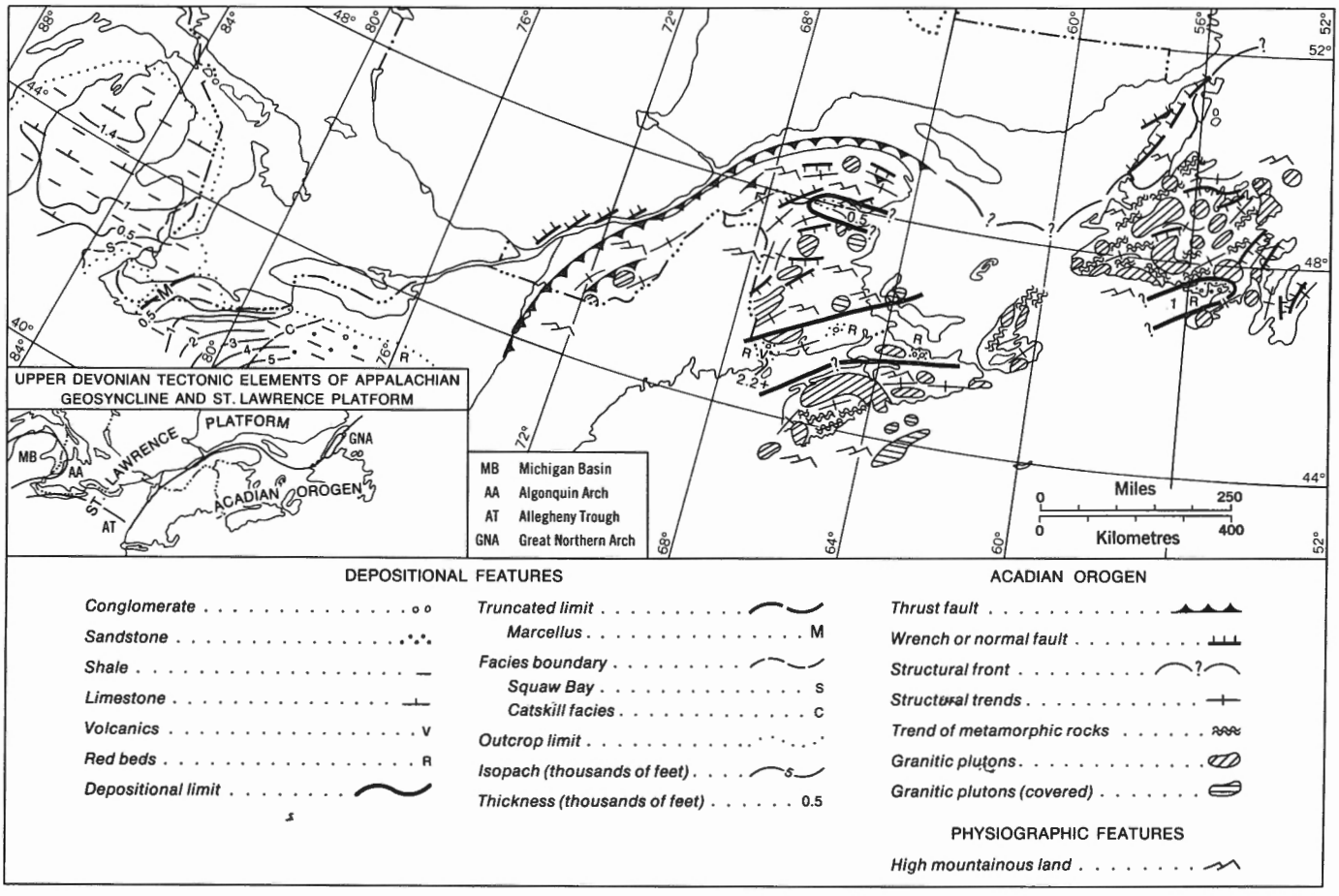


FIGURE VI-20. Late Middle and Upper Devonian sedimentation, volcanism, and tectonism in Southeastern Canada, and features of Acadian Orogeny.

m.y., both interpreted as indicating Devonian, pre-granite metamorphism.

In southern New Brunswick, most progressively regionally metamorphosed rocks with grades higher than greenschist facies probably developed in pre-Acadian times. Some basic schists, developed from Coldbrook Group volcanics, belong to the quartz–albite–epidote–biotite subfacies of the greenschist facies. A K–Ar date on lineated actinolite in amphibolite is 369 m.y., strongly suggestive of Acadian metamorphism.

On Cape Breton Island, and St. Paul Island 15 miles offshore, schists and gneisses have developed from the Precambrian George River Group and rocks correlated with it (Neale, 1963; Phinney, 1963). Progressive regional metamorphism possibly occurred in three periods of orogeny—the Avalonian, Taconian, and Acadian—but the effects of each episode in various parts of Cape Breton are not clear. Acadian regional metamorphism certainly occurred in northern Cape Breton and on St. Paul Island. Biotite from staurolite schist on St. Paul Island yielded a K–Ar date of 360 m.y., similar to the dates from recognized Acadian granites in the Appalachian Region. The rocks have been metamorphosed to greenschist and almandine amphibolite facies, and locally along some major Carboniferous faults amphibolite has been retrograded to greenschist. In the higher grade rocks, pelitic schists and gneisses contain a variable assemblage of staurolite, kyanite, almandine-rich garnet, hornblende, oligoclase, biotite, muscovite, and quartz. Metavolcanic rocks are schistose and gneissose amphibolites with hornblende, oligoclase, and garnet. Associated with the high-grade rocks are large areas of mixed granitic gneiss, pegmatite, paraschist and gneiss, and amphibolite.

In Newfoundland (Williams, 1969) regionally metamorphosed rocks with grades as high as almandine amphibolite facies form almost a quarter of the Acadian Orogen. They consist of paraschist, paragneiss, and migmatite. Metamorphism is probably Devonian but may also be late Silurian and older than the crosscutting Devonian granite stocks and batholiths. Metamorphism in western Newfoundland may be partly pre-Silurian. A belt of migmatite, schist, and gneiss extends from northeastern Newfoundland to near La Poile Bay on the south coast. Biotite in augen schist on the south coast yielded a K–Ar date of 393 m.y., a typical Acadian date. The rocks, which developed from lower Gander Lake Group, Love Cove Group, and equivalents, consist of interlayered schist, gneiss, quartzofeldspathic rocks of granitic composition, and augen schist and gneiss. Large tabular porphyroblasts of pink potash feldspar form as much as 50 per cent of the rock. The schists and gneisses contain garnet, sillimanite, kyanite, andalusite, and micas, an assemblage representing the almandine amphibolite facies. Westward and northward, the migmatite belt passes with decreasing metamorphic grade into biotite schist and gneiss, locally containing garnet, sillimanite, and kyanite. These rocks in turn grade through the biotite and chlorite zones of the greenschist facies to phyllite and slate. Eastward and southward, where not bordered by granite, the migmatites pass into greenschist facies and thence to phyllite and slate. The metamorphic rocks of much of south-central Newfoundland were developed from Ordovician and Silurian rocks. They extend southwestward to grade into Devonian strata near La Poile Bay and merge with the west-trending metamorphic rocks described above. The rocks are schist, gneiss and some granitic gneiss, migmatite, and amphibolite. Locally, they contain garnet, staurolite, and sillimanite typical of the almandine amphibolite facies, and cordierite and andalusite near granite plutons. In western Newfoundland a belt of metamorphic rocks extends from White Bay southwestward beneath Carboniferous cover to Grand Lake and thence to the southwest tip of Newfoundland near Port aux Basques. These rocks were probably metamorphosed during the Taconian and Acadian Orogenies. All the metamorphosed rocks are pre-Silurian except for Silurian volcanics at the northeast end that rest nonconformably on granodiorite. A Devonian age of metamorphism of the Fleur de Lys Group in the northeast is suggested by a K–Ar date of 355 m.y. on muscovite from a biotite–muscovite schist. On the other hand, a Taconian age is suggested by a K–Ar date of 452 m.y. on biotite from augen gneiss in the central part of the belt, and Middle Silurian ages of 420 m.y. on biotite in granite in central part and 415 and 420 m.y. on muscovite from pegmatites interlayered with schist, gneiss, and quartzite at the south end of the belt.

Near White Bay the Fleur de Lys Group (Neale and Kennedy, 1967) comprises complexly folded pelitic and psammitic schists and gneisses, minor amphibolite with mineral assemblages typical of the almandine amphibolite facies—staurolite, kyanite, garnet, and high greenschist facies. Biotite and garnet have been partly retrograded to chlorite. Near Grand Lake, the Grand Lake Brook Group and Mount Musgrave Formation consist of metamorphic rocks that increase in grade easterly from phyllite, marble, quartzite, and low-grade schists through biotite schist and gneisses of the high greenschist facies, interlayered with granitic rocks, to the lowest grade of the almandine amphibolite facies. The higher grade rocks contain biotite, garnet, sillimanite, cordierite, oligoclase, and quartz.

Structure. Almost the entire Appalachian geosynclinal belt was deformed by Acadian folds and faults, and in southern Quebec was thrust northwestward and northward onto the St. Lawrence Platform along Logan's Line. A narrow belt of Taconian folded rocks lying south of the thrusts was probably not refolded during the Acadian (Neale, *et al.*, 1961). Also, in the Conception Bay area of Avalon Peninsula, Newfoundland, the flat-lying Cambrian-Ordovician strata, underlain by Holyrood granite, were not folded although they were faulted, probably during Acadian Orogeny. Acadian folds generally have steep axial planes, and their axes plunge at angles varying from horizontal to steeply northeast and southwest. Slaty



PLATE VI-10

Recumbent Acadian folds in Ordovician slate and phyllite of Baie d'Espoir Group, formed by northwestward movement on subhorizontal axial plane cleavage, Baie d'Espoir, Newfoundland.

cleavage is generally parallel or subparallel to the axial planes. Phyllitic and fine-grained schistose rocks of various ages commonly exhibit an early isoclinal fold system which was later deformed by upright, more open folds. Even later, cross-folds and kink bands are common minor features.

Throughout the Appalachian Region, folds and faults form regional arcuate patterns that consistently curve from northeast to east to northeast in the shape of immense, open, right-lateral drag-folds (Map 1251A). Indeed, the boundary between the Appalachian Geosyncline and the St. Lawrence Platform traces out the same pattern. Major transcurrent faults parallel the belts. These patterns seem to be related to the trends of pre-Acadian tectonic elements.

On Avalon Peninsula, folds and faults in Ordovician and older rocks and faults in Devonian granites are presumed to be Acadian (McCartney, 1967). Folds are upright and vary from gentle open folds to tightly compressed; the Conception Bay area was apparently not folded during the Acadian. Several east-directed thrusts with net slips of a few thousand feet and less cut Cambrian and older rocks in the eastern half of Avalon Peninsula. The thrusts are cut by and perhaps synchronous with northwest- to northeast-trending steep faults, many of which have several miles of horizontal displacement and a few thousand feet of vertical displacement. In Trinity Bay the tear faults have both right-lateral and left-lateral displacements. Many of the major northerly trending, steep faults were apparently active in late Hadrynian time and were reactivated during Acadian Orogeny. In eastern Newfoundland, northwest of Avalon Peninsula, northerly trending, upright folds and long, steep faults cut Ordovician and older rocks. All folds and some faults are truncated by Devonian granite, but some faults cut the granite. Some Cambrian-Ordovician strata have been overturned to the west and some to the east. Complexly folded phyllites and siltstone occur at Baie d'Espoir, and probably extend northward through Gander Lake. Rocks with open folds and axial planes dipping about 55 degrees northwest are deformed farther south by recumbent shear folds with subhorizontal cleavage dipping 12 degrees southeast (Pl. VI-10). In northeast and central Newfoundland, folds in Ordovician strata are generally upright or northwesterly overturned and tightly compressed, whereas those in Silurian conglomerate and sandstone are

more open and locally overturned to the northwest. All faults are steep; some are intruded by granite and others cut granite. In Notre Dame Bay (Kay, 1967) several northeast-trending slices of northwesterly overturned Ordovician and Silurian strata, several thousand feet thick and dipping homoclinally southeast, are bounded by transcurrent faults.

Schists and gneisses of the Fleur de Lys Group (Neale and Kennedy, 1967) in western Newfoundland are complexly folded. The prominent, steeply dipping schistosity is parallel to the axial planes of tight and isoclinal folds in which locally an older isoclinal fold system is apparent. Steep strain-slip schistosity, which parallels the axial planes of the younger tight folds, cuts the prominent schistosity. Still later are flexure-slip folds and kink bands.

In central and eastern Nova Scotia (Fyson, 1966) three generations of Acadian folds and faults were developed. The oldest and most prominent folds are upright and low-plunging, and have a steep axial-plane cleavage. Younger north-trending cross-folds associated with a steep fracture cleavage are small and developed only locally. Both fold systems are older than the cross-cutting granite plutons. The youngest deformation is represented by northwest-trending kink bands that pass into left-lateral faults; these folds and faults are both pre- and post-granite. In southeastern Cape Breton Island, folds believed to be Acadian are upright and north-trending; associated Acadian faults are steep and right-lateral. In northern Nova Scotia the main Acadian folds are both upright and overturned southeasterly. They were deformed by minor cross-folds and by major northeast-trending steep faults and minor north- to northwest-trending faults. In New Brunswick, Acadian folds are generally upright and in places overturned to the northwest and southeast. They vary from open to tightly compressed and have axial-plane cleavage. Folds and cleavage with some variations trend northeasterly and are truncated by granite. Major strike faults and minor cross-faults cut the folded rocks; some are pre-granite and some are post-granite.

In the Eastern Townships of southern Quebec, all Silurian-Devonian strata southeast of Sherbrooke are overturned to the southeast in an extraordinarily large homocline (St-Julien, 1967). These rocks, with some unconformably underlying Cambrian-Ordovician rocks, have been thrust northwestward upon a belt of Middle and Upper Ordovician and the unconformably overlying Upper Silurian rocks. The Upper Silurian rocks themselves, near Lake Memphremagog, have upright tight and open folds, in places slightly overturned to the northwest, with a well-developed axial-plane cleavage. The northwestern limit of Acadian folding is uncertain, but it appears to extend no farther than the string of Taconian ultramafic bodies bordering the anticlinal Sutton–Bennett belt on the southeast. In the Témiscouata–Matapedia area folds in Silurian-Devonian rocks are generally broad and open and locally tight. Some beds are overturned to the northwest. Major faults are steep; some are clearly northwest-directed thrusts whereas others are normal with southeast side down. In Gaspé Peninsula (McGerrigle, 1950) Acadian folds are broad and open and they plunge east and west with low-angles to produce doubly plunging regional synclines and anticlines and elongate domal structures in northern Gaspé. In southern Gaspé the incompetent Fortin pelitic strata and Matapedia – White Head – Pabos thin-bedded limestones and shales are commonly closely folded and crumpled. Some strata and the limbs of some major folds are overturned northerly. Beds in southern Gaspé dip more steeply than those in the north, except near the unconformably underlying Mictaw Formation in southern Gaspé. Northern Gaspé rocks are cut by a few, steep, north-directed thrust faults and normal faults with the south side down. Horizontal displacement on the thrusts is presumed to be less than a few thousand feet. Longitudinal, steep, normal faults, extending for many miles, have vertical displacements of several thousand feet; some are downthrown on the south and others on the north. Other faults in eastern Gaspé trend northwesterly, more northerly than the fold and longitudinal normal and thrust faults, and transect folds and possibly the longitudinal faults. These oblique faults are probably of normal type with northeast side down, or they may be transcurrent. The location of the northerly limit of Acadian folding north of the Silurian-Devonian strata in the Témiscouata–Gaspé region is uncertain, but was probably somewhere south of St. Lawrence River.

The thrust faults comprising Logan's Line along the boundary between Appalachian Geosyncline and St. Lawrench Platform in southern Quebec bring Cambrian-Ordovician geosynclinal pelitic and psammitic strata northwestward and northward upon Middle and Upper Ordovician limestone and pelitic strata of the platform between Lake Champlain and along St. Lawrence River as far east as Anticosti Island. The folded and thrust-faulted pelitic strata of the St. Germain Complex in the Eastern Townships occur within the zone of Logan's Line. The thrusts mark the northerly limit of Paleozoic deformation. Some thrusting occurred during the late Taconian Orogeny but Acadian thrusting probably produced most of the present shortening, for if the thrusts of Logan's Line were late Taconian, coarse detritus derived from highlands in the upper plate would probably have been deposited on nearby Anticosti Island. On Anticosti Island, however, only carbonate and shale were deposited continuously from the Ordovician to the Silurian. Southwest of Quebec City, along the north side of the St. Lawrence River valley, are normal faults along which the Precambrian crystalline basement was raised against Paleozoic strata to the southeast. The faults may be Acadian or younger.

Along the west coast of Newfoundland and southeast coast of Labrador, several northeast-trending faults cut the Precambrian basement, flat-lying to gently folded Paleozoic cover strata and rocks of the Taconian klippen. The faults are steep and trend northeast. Some are

normal and some are reverse; in most the side adjoining the Gulf of St. Lawrence is down, producing a graben-like structure. On western Port au Port Peninsula the Lower Devonian Clam Bank Formation is cut by faults and locally overturned northwesterly. Many or all of these folds and related faults are probably Acadian, but some may be Carboniferous or younger.

Plutonic rocks. Acadian granites intrude strata of all ages as young as Lower Devonian in Appalachian Geosyncline except parts of southern Quebec, northwestern New Brunswick, and Avalon Peninsula. About 60 potassium–argon and a few rubidium–strontium isotopic age determinations have been made on minerals and whole-rock samples. The dates range from 420 to 250 m.y.; many are concentrated in the 350 to 375 m.y. range, about middle and late Devonian. Most New Brunswick granites provide dates on micas in the 380 to 395 m.y. range, which is from 20 to 40 m.y. older than the 350 to 380 m.y. range common to many Nova Scotia granites. Several dates between 400 and 450 m.y. were obtained from dioritic and gabbroic border phases of some Newfoundland granites, whereas the more granitic interiors have yielded dates of 350 to 400 m.y., even though the composite plutons cut Middle Silurian rocks (*ca.* 420 m.y.).

Acadian plutonic rocks of New Brunswick, Nova Scotia, and central Newfoundland reach large batholithic proportions, whereas those of southern Quebec and southeastern Newfoundland are small batholiths and stocks. Granite with abundant quartz is the most common plutonic rock; diorite and gabbro are much less common. Where granites and basic rocks occur together the granites intrude or grade into the basic rocks. Almost all are regionally discordant bodies with sharp contacts and steep walls and have hornfelsic thermal aureoles.

Diorite and gabbro form large bodies in western and southwestern New Brunswick and on Cape Breton Island. They form either discrete bodies or border phases of granite bodies, and consist of dark coloured rocks of various grain sizes and amounts of hornblende, pyroxene, and/or biotite. Small amounts of late-phase granodiorite and quartz diorite accompany the diorites and gabbros. Most of these basic rocks are chloritized and saussuritized.

Most granitic rocks are granodiorite and quartz monzonite with biotite; others contain hornblende and muscovite, generally accompanied by biotite. Younger phases of the granitic sequence (Poole, 1963) are commonly muscovitic, buff weathering, and either equigranular or porphyritic; many are associated with such minerals as fluorite, tourmaline, beryl, cassiterite, wolframite, and molybdenite, and some of these locally have been altered to greisen. Such granite bodies in central New Brunswick have steep walls and undulatory tops, and have been only partly unroofed. Granites in northern Gaspé Peninsula and southeastern Cape Breton Island are associated with bodies, sills, and dykes of rhyolite and porphyry. In northern Gaspé and northern New Brunswick skarns contain deposits of base metal sulphides, whereas in southern Nova Scotia gold–quartz veins, some with accessory scheelite, developed in Meguma strata. The granitic rocks of northern Cape Breton Island (Neale, 1963) comprise a suite much more varied in texture and composition; they have been emplaced in a metamorphic terrane of schist and gneiss interlayered with granitic rock and partly feldspathized. The schist and gneiss also form feldspathized inclusions in once-mobile granitic rock. Many such rocks are migmatites. Some coarse-grained granite contains microperthite porphyroblasts as long as 3 inches. The youngest granite bodies grade into the metamorphic rocks and consist of a composite of granodiorite, granite, and syenite, with biotite, muscovite, and hornblende. Some phases are gneissic.

Acadian plutonic rocks in Newfoundland comprise about a third of the outcrop area of the Appalachian geosynclinal belt and consist of several varieties ranging from gabbro to alkaline granite. The ultramafic bodies in Newfoundland are believed to be Ordovician, but one body of serpentinite, serpentinized peridotite, and layered gabbro, about 10 miles in diameter, appears to have intruded the Silurian Botwood Group, and, if so, could be Acadian. The mafic to silicic plutonic rocks comprise four general groups. Those of the presumably oldest group have intruded the Middle Silurian Botwood Group and have yielded biotite dates in the 350 to 450 m.y. range, suggesting a Middle Ordovician to late Devonian age. Most of the bodies are composite: dominantly massive to locally layered augite gabbro, hornblende diorite, and biotite–hornblende quartz diorite have been intruded by biotite, hornblende granodiorite, and quartz monzonite, with intergradational minor hornblende syenite and monzonite. Contacts with stratified host rocks are sharp and locally consist of agmatite; thermal aureoles are relatively narrow. Perhaps belonging to this group are the foliated hornblende–biotite granite, diorite, and gabbro of Burin Peninsula. The second group of plutons yielded K–Ar dates on micas from 350 to 370 m.y., about Middle and Upper Devonian. The plutons consist of a distinctive, uniformly massive, coarse-grained, pink to grey porphyritic biotite quartz monzonite commonly referred to as the Ackley granite type. Phenocrysts are tabular pink perthite crystals as long as 4 inches. The granites have sharp contacts and narrow thermal aureoles. The third group of plutons is least abundant. It occurs among the second group in eastern and southern Newfoundland, and consists of garnetiferous muscovite leuco-quartz monzonite. One K–Ar date on muscovite was 360 m.y. The rocks are massive, generally medium grained and equigranular, and contain disseminated pink garnets and abundant white potash feldspar. Associated with them are white pegmatites with accessory garnet, tourmaline, and beryl. The presumably youngest group of plutons occurs east of the main areas of granite intrusion and within the western part of the Cambrian-Ordovician Avalon Platform. The

plutons consist of massive, alkalic granite stocks and small batholiths on Burin Peninsula and near Treytown, Bonavista Bay. They are light pink to red, leucocratic, medium to coarse grained, equigranular, and massive. Miarolitic cavities are common in plutons of Burin Peninsula. The granites consist of abundant quartz and perthite and some albite, locally with riebeckite and fluorite. One sample analyzed about 11 per cent Al_2O_3 and over 8 per cent soda plus potash. Fluorite is mined from veins in the granites at St. Lawrence. Dykes and sheets of porphyritic rhyolite cut the Burin Peninsula strata and alkalic granites.

Diorite, diabase, and gabbro dykes cut Acadian folded rocks and plutons. Most are probably Devonian and others Mesozoic. They are generally not deformed, consist of plagioclase, hornblende, and augite, and some are slightly altered.

Late Middle and Upper Devonian

Appalachian Geosyncline (Fig. VI-20). On New Brunswick Platform, the Fleurant Formation along Chaleur Bay (Alcock, 1936) consists of 45 feet of fluviatile polymictic conglomerate that overlies the Pirate Cove Formation with a disconformity or low-angle unconformity. The conglomerate contains pebbles and boulders as much as 4 feet long, of Dalhousie volcanics, granite, Pirate Cove Formation, and a variety of other rocks. The Escuminac Formation conformably overlies the Fleurant and consists of 370 feet of non-marine grey shale and sandstone containing fish fossils and plants of early Upper Devonian age. The Escuminac is overlain unconformably by Carboniferous (Namurian?) conglomerate.

Along the northwest edge of the Fundy Epieugeosyncline in southern New Brunswick, the Perry Formation is composed of over 2,100 feet of terrestrial red sandstone, conglomerate, siltstone, and few basalt flows and tuff beds. Upper Devonian plants occur in nearby Maine. Clasts, as much as 6 inches in diameter, are of Mascarene volcanics, granite, and a variety of other rocks. Channels, trough crossbedding, and pebble imbrications indicate a southward current flow. Near Moncton the Memramcook Formation of non-marine red sandstone and conglomerate occurs at the base of a Mississippian sequence and recently has yielded Upper Devonian spores.

On Newfoundland Platform the Upper Devonian Great Bay de l'Eau, Terrenceville, and Pools Cove Formations consist of red conglomerate with arkosic sandstone, siltstone, and shale interbeds; they unconformably overlie Silurian and older strata. The first two formations contain Upper Devonian plants, but recently Terrenceville yielded spores that suggest an early Mississippian age. The Pools Cove is cut by a granite, clasts of which appear in the Great Bay de l'Eau. The formations are 1,000 to 3,000 feet thick and are tilted and gently folded. Clasts were derived from local older formations and granites. Ripple-marks, mudcracks, rain prints, and plant fossils suggest a terrestrial, probably fluviatile environment of deposition.

Deformational and plutonic events attributable to the late Devonian part of the Maritime Disturbance are difficult to distinguish from those that occurred during the Carboniferous and Permian. In most of the Appalachian Geosyncline the oldest, post-Middle Devonian strata are middle Mississippian or younger. Tilting, uplift, and erosion of Upper Devonian Escuminac Formation in Chaleur Bay was followed by deposition of Carboniferous (Namurian?) Bonaventure Formation. In Fortune Bay of southern Newfoundland the Upper Devonian Terrenceville Formation is tilted and gently folded and has been overthrust by Precambrian rocks from the southeast, possibly during the late Devonian. Forty miles to the west the Upper Devonian Great Bay de l'Eau Formation is tilted and broadly folded and has been intruded by the Belleoram granite stock. The stock is a massive, medium-grained hornblende–biotite quartz monzonite, the biotite yielding a K–Ar date of 342 m.y., very near the Devonian–Mississippian boundary. The roof of the stock is nearly planar and horizontal.

St. Lawrence Platform (Figs. VI-20, 21). In southwestern Ontario deposition of late Middle and Upper Devonian sediments continued during the Acadian Orogeny and the late Devonian interval of the Maritime Disturbance in the Appalachian Geosyncline (Sanford, 1968). The former distribution of these strata on the platform is unknown, but similar strata occur on the Hudson Platform, suggesting that the two regions were probably connected.

The Givetian Hamilton Group has a thickness of 300 feet near southern Lake Huron bordering the Michigan Basin, and thins to 200 feet beneath central Lake Erie in northern Allegheny Trough. The lower beds, the Marcellus Formation, consist of 50 feet of black bituminous shale and minor argillaceous limestone lying conformably on the Seneca limestone beneath central Lake Erie. During subsequent uplift and erosion the presumably more widespread Marcellus shales were stripped off. The succeeding Arkona shale and Ipperwash limestone, 150 feet thick beneath Lake Erie, transgress the erosional surface of the Marcellus and Dundee Formations and grade westward into 300 feet of alternating shale and limestone bordering Michigan Basin. Near Lake Huron the Hamilton Group consists of several formations which in ascending sequence form the Bell shale (60 feet), Rockport Quarry limestone (10 to 20 feet), Arkona shale (120 feet), Hungry Hollow limestone and shale (6 feet), Widder shale and limestone (45 feet), and Ipperwash limestone (94 feet).

Upper Devonian strata underlie the Chatham Sag between the Algonquin and Findlay Arches in southwestern Ontario, and form the youngest strata of the Paleozoic succession (Sanford, 1968). They have a maximum thickness of 350 feet in Ontario, thicken northwestward to 985 feet in northwestern Michigan, and southeastward to 1,000 feet along the south shore of Lake Erie. The Kettle Point Formation disconformably over-

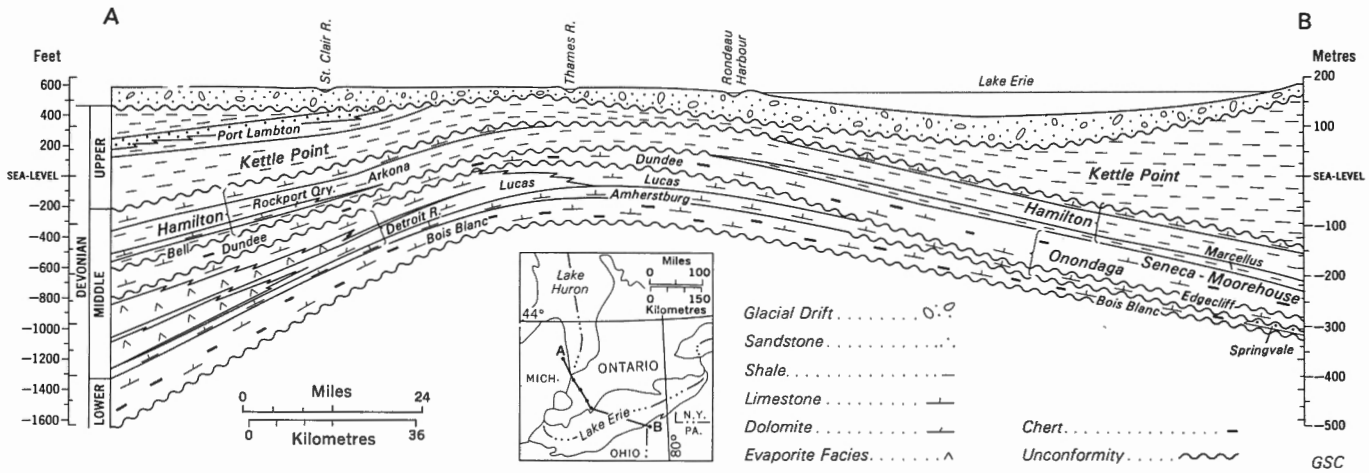


FIGURE VI-21. Cross-section through Devonian rocks of Algonquin Arch of southwestern Ontario (by B. V. Sanford).

lies the Hamilton Group and consists of black and dark brown bituminous shale with interbeds of greyish green shale. Where exposed on the shore of Lake Huron, the shales contain numerous calcareous concretions as much as several feet in diameter that are commonly referred to as "kettles" (Pl. VI-11). The Kettle Point is 250 feet thick near southern Lake Huron, and it thins southeastward to 100 feet or less in Chatham Sag then thickens to 1,000 feet along the south shore of Lake Erie. The Kettle Point is overlain conformably by three formations south of Lake Huron. These total 200 feet and thicken westward into Michigan Basin where they underlie non-marine Mississippian clastics not represented in southwestern Ontario. In ascending sequence they comprise the Bedford grey shale, Berea micaceous sandstone, and the Sunbury black shale.

Carboniferous and Permian

Tectonic Summary

Carboniferous rocks lie with regional unconformity on granites and pre-Middle Devonian strata deformed by the Acadian Orogeny, but locally lie conformably on post-Acadian Upper Devonian rocks. None occurs on St. Lawrence Platform in Canada.

Carboniferous sediments are almost entirely non-marine and were deposited in a series of connected troughs or intermontane basins within a regionally subsiding area, the Fundy Epieugeosyncline (Figs. VI-22 to 28). The area of greatest subsidence probably extended from the Bay of Fundy in southeastern New Brunswick to southwestern Newfoundland. It is within this area that the thickest and most complete sections of Carboniferous rocks are preserved. Although local unconformities are present within Carboniferous rocks (Pl. VI-12), sedimentation was probably continuous in one part or other of the epieugeosyncline until early Permian. The areas of deposition were, at most times, separated from one another by complementary positive areas from which detritus was derived, but were not always separated from one another throughout the entire Carboniferous and early Permian. At other times, what had been previously uplands and source areas became depositional basins.

A delicate balance in base-level between basins of deposition and source areas was maintained throughout much of Carboniferous time. This is inferred from the thousands of feet of coarse clastics fairly similar in composition, texture, and primary features that occur in some Carboniferous sections. The maintenance of source areas at a fairly consistent level relative to adjacent depositional basins was accomplished by continuous uplift of source areas, either as welts or by movement on bordering faults. Epeirogenic subsidence or uplift of the whole of the Carboniferous of the entire Fundy Epieugeosyncline and bordering platforms took place at the same time as folding or faulting within or along the boundaries of upland areas within it. These epeirogenic movements were mainly opposite to the local movements so that epeirogenic subsidence occurred at the same time as uplift of local source areas.

The combination of epeirogenic subsidence with local uplifts during Lower Carboniferous sedimentation resulted in onlap of each succeeding rock unit upon the flanks of positive areas. Such a pattern continued until late Lower Carboniferous, by which time much of the Acadian Orogen became covered by a shallow sea. During the preceding interval islands within the sea had been the source of non-marine deposits. Erosion and uplift of these islands continued during the marine deposition to produce coarse and fine marginal deposits. Epeirogenic uplift during the interval of marine deposition resulted in a withdrawal of the sea, first from the New Brunswick Platform and from some parts of the Moncton and Cumberland Basins, and eventually from the rest of the Appalachian Region. As the sea withdrew non-marine clastic sedi-

ments were deposited conformably above the marine sediments in the axial regions of the basins. By the end of the Lower Carboniferous, or early in the Upper Carboniferous, continental conditions had completely returned to eastern Canada.

During earliest Upper Carboniferous epeirogenic uplift continued. The area of deposition was smaller than that of the earlier marine deposition and the sediments generally have offlap relationships. Local areas were uplifted more than the regional uplift and continued as source areas for the alluvial sediments of the adjacent basins.

The combination of epeirogenic subsidence with local uplifts occurred again, probably from late Namurian to early Westphalian C. By the end of Westphalian B, or early in Westphalian C, local tectonic activity in the bordering uplands had ended with the result that succeeding Carboniferous sediments were essentially controlled by epeirogenic movements. From Westphalian D to early Permian either the supply of sediments was depleted or epeirogenic uplift again resulted in a restriction of deposition to eastern New Brunswick and Prince Edward Island, northern Nova Scotia, and western and eastern Cape Breton Island. These areas were possibly part of one, larger basin.

The non-marine sediments of the Carboniferous basins were mainly gravels, sands, silts, and muds deposited in fluvial and flood plain environments; and sands, silts, and muds laid down in lacustrine and paludal environments. The various facies occur as interfingering lenses and are commonly time–transgressive. These interdigitating facies were further complicated by the intercalations of alluvial gravels and sands. Fairly extensive coal-forming swamps were present in various parts of the basins throughout the Upper Carboniferous. Coal has been mined from strata of Westphalian A, B, C, and D ages. The marine sediments include fine-grained red beds, limestone, and evaporites that commonly grade laterally into fine- to coarse-grained red beds in the vicinity of uplands. Thin basalt, andesite, and rhyolite flows occur locally and at different stratigraphic levels within the Carboniferous.

The Carboniferous tectonic activity referred to as the Maritime Disturbance occurred mainly in the basin areas of Fundy Epieugeosyncline (Fig. VI-29). It is manifested by local unconformities, open folds of various amplitudes, and high-angle reverse and strike-slip faults that occur mainly along the border between basins and uplands. Also, small granitic stocks and dykes were emplaced at one locality during the Carboniferous. On the platforms evidence of the disturbance is non-existent or limited to very broad open folds and local faults. There the main tectonic activity was epeirogenic subsidence and uplift.

Unconformities in the epieugeosyncline are mainly local, adjacent to upland areas. Their differences in age at various locations indicate that movement of the uplands was not synchronous throughout, but varied somewhat from place to place. Also, the contact between two time-

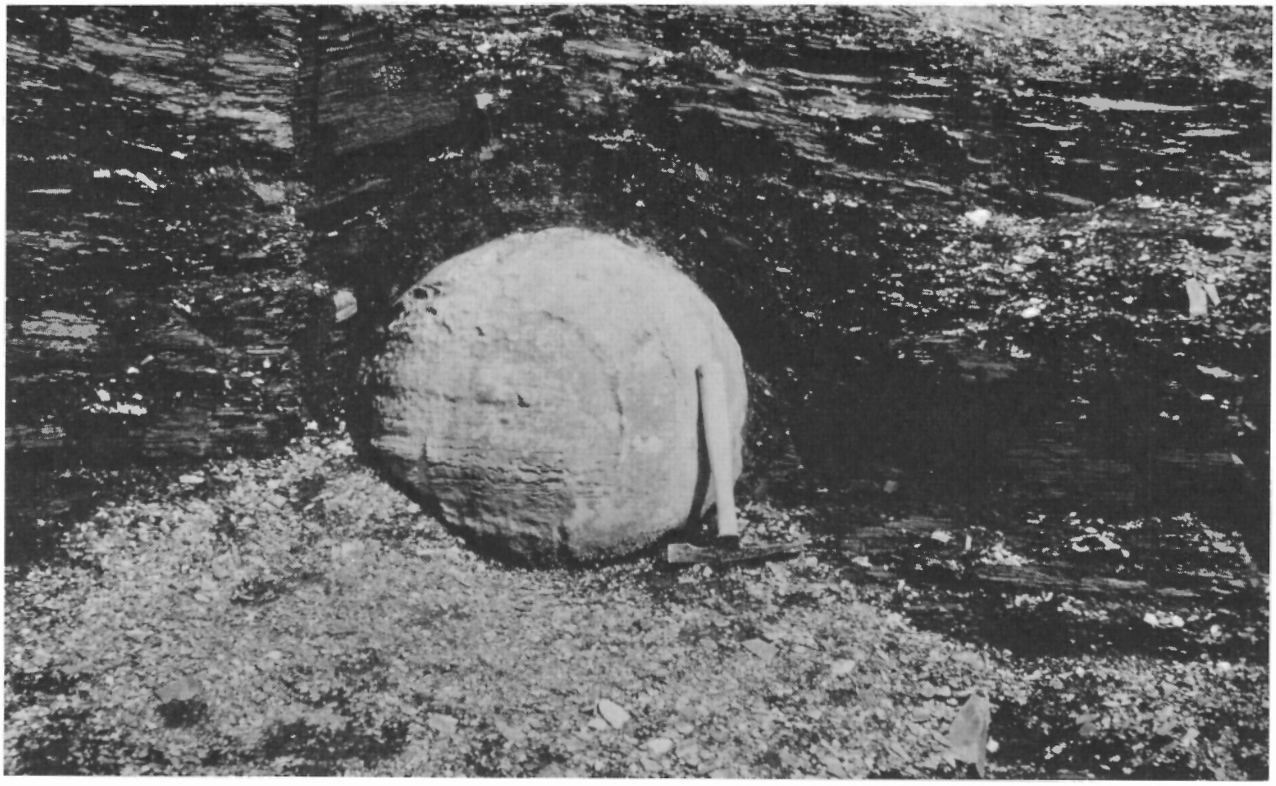
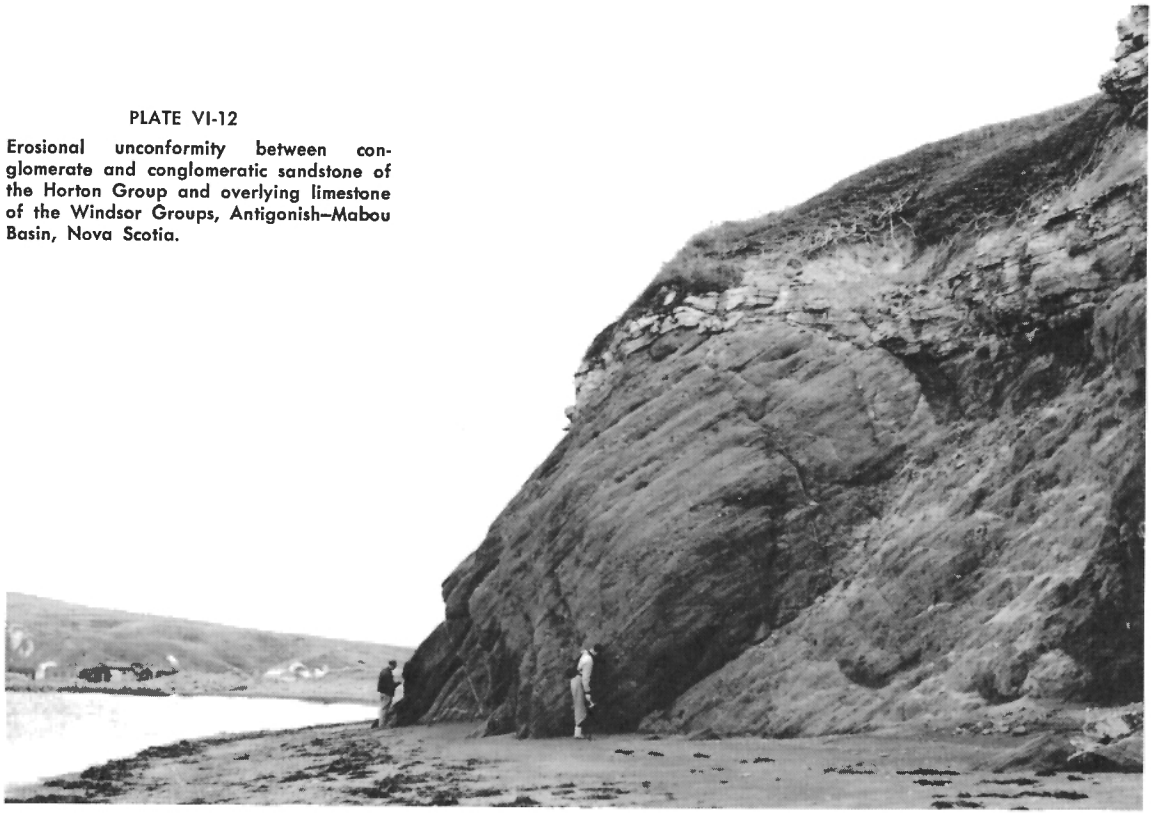


PLATE VI-11. Limestone concretion called a "kettle" in Upper Devonian black shale of Kettle Point Formation, southern Lake Huron, Ontario.

PLATE VI-12

Erosional unconformity between conglomerate and conglomeratic sandstone of the Horton Group and overlying limestone of the Windsor Groups, Antigonish–Mabou Basin, Nova Scotia.



stratigraphic units may be conformable adjacent to one upland but unconformable adjacent to another. Each basin had its own unique development, but sedimentation was continuous throughout the Carboniferous.

Westphalian A strata in the Moncton Basin, and mainly Westphalian C or younger strata elsewhere, extensively blanket all earlier Carboniferous rocks. The blanket is conformable to unconformable in the area of epieugeosynclinal activity and disconformable on the platform areas. These Carboniferous rocks have gentle dips except in the vicinity of faults; where folded, they are characterized by very broad open folds that are probably due to the terminal pulses of the Maritime Disturbance or to salt doming.

Stratigraphic Nomenclature

The subdivision of Carboniferous rocks most commonly used is the essentially time-stratigraphic subdivision of Bell (1958). His most recent subdivisions (Table VI-1) are but a slightly modified version of those he proposed in 1927. Although the major units are called groups and can be defined partly by lithology, they are primarily defined by their age as indicated by fossils. Strict adherence to Bell's subdivision on a time-stratigraphic basis has required the placing of "group" boundaries where no lithological change is present.

For many years Bell's subdivisions remained unchanged, but in recent years there has been a considerable increase in the basic knowledge of Carboniferous stratigraphy, sedimentation, and tectonics that has led to changes in interpretations. For example, Bell's groups were originally considered to be separated by intervals of non-deposition, except locally between the Windsor and Canso Groups. It is now recognized that the contacts between most, if not all, of the groups are conformable at some localities and that some are diachronous.

TABLE VI-1 *Carboniferous strata in the Maritime Provinces (after Bell, 1958)*

Group	Age
Pictou	Westphalian C and D
Local unconformity and disconformity	
Cumberland	Early Westphalian B
Local unconformity and disconformity	
Riversdale	Westphalian A
Local unconformity and disconformity	
Canso	Early Namurian and (?) Late Viséan
Conformity and disconformity	
Windsor	Viséan
Rare local unconformity and disconformity	
Horton	Tournaisian
Unconformity	

Attempts have been made to correlate Carboniferous rocks on the basis of unconformities, but the unconformities are now known to indicate local movement and their effects disappear over short distances. Furthermore, the unconformities occur at various stratigraphic levels in different areas. A possible exception that may have regional significance is the unconformity beneath the uppermost Carboniferous Pictou Group. The group is commonly unconformable on either older Carboniferous or pre-Carboniferous rocks in western Cape Breton Island and in most parts of the Cumberland Basin. In two areas of the Cumberland Basin the Pictou Group, as a time-stratigraphic unit, is possibly conformable with the underlying Cumberland Group, whereas in most basins the Pictou Group overlaps older strata without any angular discordance and with only a minor gap in the sedimentary record.

The Lower Carboniferous, Horton, and Windsor Groups are now used solely as rock-stratigraphic units. The Horton Group is essentially the same as defined by Bell, except for removal of the Tournaisian time restriction. The partly to entirely marine Windsor Group is also restricted in use to a rock-stratigraphic unit and is only applied to those strata that include marine limestone and/or evaporites. In the Upper Carboniferous the problem of sorting out stratigraphic units largely remains, although some progress has been made (Table VI-2). In the text to follow the word "group" is used in the same way as proposed or used by earlier workers, that is with a biostratigraphic or time-stratigraphic connotation. Without quotation marks group is used as a rock-stratigraphic term.

Lower Carboniferous

The Horton Group (Fig. VI-22). This is a sequence of mainly Tournaisian (early Lower Carboniferous) continental sedimentary and volcanic rocks that rest unconformably on rocks as old as Precambrian and are overlain conformably or disconformably, and rarely unconformably, by the partly marine strata of the Windsor Group. The type locality is in the southern part of Minas Basin where it is divided into two formations: a lower Horton Bluff Formation and an upper Cheverie Formation. These formations are not recognized beyond the type area. The term Horton Group has been used throughout Nova Scotia and New Brunswick, but lithologically comparable rocks in Newfoundland in the same stratigraphic position are called Anguille Group. The Horton Bluff Formation consists of 1,000 to 3,400 feet of siltstone, shale, and minor ferruginous limestone; the Cheverie consists of 600 feet sandstone, siltstone, shale and arkose. The fine-grained lithology of the Horton Bluff and Cheverie Formations appears to be limited to the Minas Basin and to the subsurface in Cumberland Basin.

In the Moncton, Antigonish–Mabou, St. George's, and White Bay Basins the Horton and Anguille Groups are typically composed of red or red and grey sandstone and conglomerate, succeeded by grey siltstone, sandstone, and shale, then by red to red and grey sandstone, siltstone, and conglomerate (Table VI-3). This sequence is more typical of Horton Group rocks than those in the area of the type section. Volcanic and minor sedimentary rocks occur at the base of the Horton in parts of the Antigonish–Mabou Basin. Sandstone, conglomerate, and siltstone of the Craignish Formation are areally the most common lowest Horton formation on Cape Breton Island, however. The Albert, Strathlorne, and Snakes Bight Formations appear to be correlative because they are similar in lithology and stratigraphic position, but fossil spores suggest that the Albert is older than the Strathlorne.

Earliest sedimentation of the Horton Group probably began in Moncton Basin and was soon followed by deposition in Antigonish–Mabou Basin. In New Brunswick, spores of probable Upper Devonian age have been obtained from the Memramcook Formation, the lowest part of the Horton Group. On Cape Breton, earliest Carboniferous spores (there is a slight possibility they may be late Devonian) have been obtained from the Fisset Brook Formation.

Continental sedimentation continued uninterrupted longer than elsewhere into the Viséan, in Moncton and Sydney Basins, and probably also in Cumberland Basin. Continental sediments of the Hillsborough and Grantmire Formations were being deposited at the same time as marine limestones of the Windsor Group.

Deposition of Horton sediments appears to have commenced in isolated basins with the source nearby. In the Antigonish–Mabou and Moncton Basins each succeeding rock unit was not only deposited upon, but overlapped the older rocks to rest directly on pre-Carboniferous rocks. Consequently, the size of the depositional basins continually increased, first by sediments joining

TABLE VI-2 *Carboniferous stratigraphy in the Maritime Provinces*

Group	Age
Pictou Local unconformity, disconformity, and conformity	Westphalian C to Early Permian
Cumberland Unconformity, disconformity, and conformity	Westphalian A, B, and ? early C
Riversdale Conformity, disconformity, and local unconformity	Late Namurian and Westphalian A
Canso Conformity and disconformity	Late Viséan to late Namurian
Windsor Conformity, disconformity, and rare local unconformity	Viséan to early Namurian
Horton Unconformity	Late Devonian, Tournaisian, and early Viséan

around the ends of the upland areas and later by covering the uplands themselves, the sediments then being derived from more distant sources. There is no evidence to suggest that Cumberland Basin was separated from Minas Basin during Horton deposition. Cobequid Upland does not appear to have been a source area for Horton sediments in Minas Basin. Antigonish Upland, however, was uplifted during Horton sedimentation, as suggested by onlap of coarse conglomerate whose source rocks appear to have been the upland and by the presence of a local unconformity within the group.

Because the Horton Group changes to a coarse facies near the margin of some basins, the adjacent upland areas are presumed to have existed throughout Horton sedimentation. For example, the Albert Formation of Moncton Basin grades southwest into a conglomeratic unit, and the Strathlorne Member in Antigonish–Mabou Basin also grades easterly into a conglomeratic unit. The conglomeratic units of both areas are similar in composition, texture, and other primary features to conglomerates both below and above the Strathlorne Member and Albert Formation. The widespread vertical and lateral distribution of polymictic conglomerate in the Horton Group with phenoclasts from pre-Horton rocks is most easily explained by periodic availability of source material that may have resulted from either uplift in the source areas (some of which were local uplands), or downsinking in the basin area, or by a combination of the two. The nature of the movement is unknown. It only locally affected Horton strata already deposited, however, as only local unconformities resulted. Uninterrupted sedimentation took place a few miles from the area of disturbance.

The Windsor Group (Figs. VI-23, 24). This group is partly or entirely composed of marine strata that conformably to unconformably overlie the Horton Group

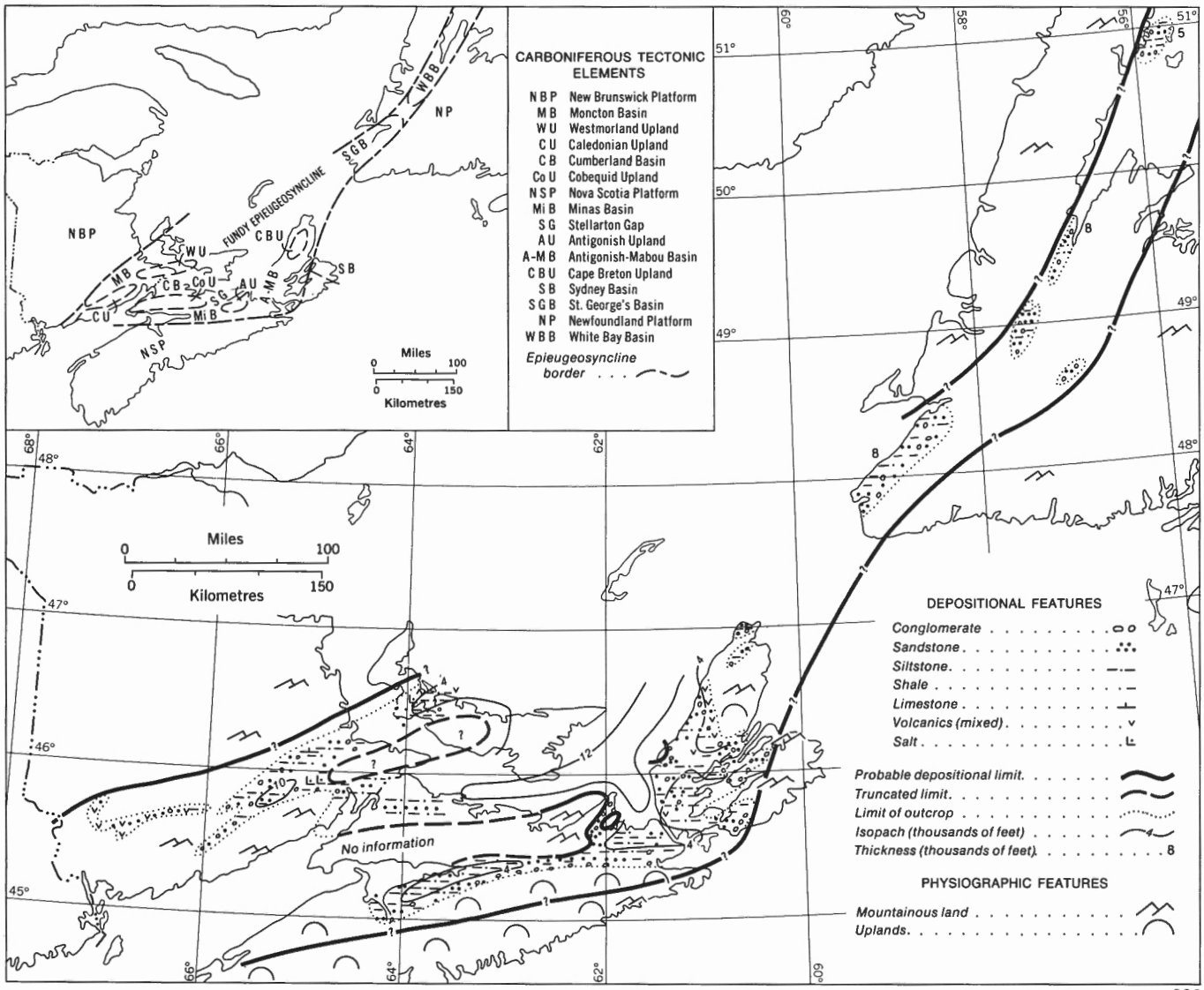


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FIGURE VI-22. Carboniferous, mainly Tournaisian (Horton Group), sedimentation, volcanism, and tectonism in Southeastern Canada.

TABLE VI-3 | *Subdivisions of the Horton and Anguille Groups*

Rock-stratigraphic unit		Lithology	Thickness (feet)
		NEW BRUNSWICK (Moncton Basin)	
Moncton Formation	Hillsborough Member	Red sandstone and conglomerate; basal ash bed?	±2,900
	Weldon Member	Red siltstone, shale, sandstone, and conglomerate	±5,000
Albert Formation		Dark grey, in part bituminous, shale and sandstone; oil shale; some limestone and salt	±5,500
Memramcook Formation		Red shale, sandstone, and conglomerate	±7,000
		CAPE BRETON ISLAND (Antigonish–Mabou Basin)	
Strathlorne–Ainslie Formation	Ainslie Member	Red and grey sandstone, siltstone, conglomerate; minor limestone and intraformational conglomerate	340–1,820
	Strathlorne Member	Grey sandstone, siltstone, shale; minor limestone	75–1,690
Craignish Formation		Grey arkosic sandstone and conglomerate, red siltstone, sandstone, and conglomerate; minor grey siltstone and sandstone	±65–8,000
Fisset Brook Formation		Andesite and minor rhyolitic rocks; red siltstone; minor conglomerate, and sandstone	800–1,500
		WESTERN NEWFOUNDLAND (St. George's Basin)	
Seacliffs Formation		Grey massive, feldspathic sandstone; grey thinly bedded, siltstone and shale; red and grey arkose; red siltstone	±1,700–4,000
Snakes Bight Formation		Thinly laminated, black, fine-grained argillaceous siltstone and dense, grey sandstone	±1,500
Cape John Formation		Grey siltstone, sandstone, and conglomerate; red siltstone	>1,500

and overlap pre-Carboniferous rocks. The type locality is near Windsor, Nova Scotia, partly in Minas Basin and partly on Nova Scotia Platform. There, the Windsor Group is 1,500 feet thick and divisible in ascending order into Subzones A, B, C, D, and E. The Windsor Group occurs in New Brunswick and on Magdalen Islands, but in Newfoundland correlative marine strata are included in the Codroy Group (*Frontispiece*). Continental strata equivalent to the Windsor Group are included with either the underlying Horton Group or the overlying Canso Group.

The Windsor Group consists of thick members of massive to poorly bedded, red or red and greenish grey, mottled siltstone, shale, and sandstone intercalated with gypsum, anhydrite, salt, and thin, tabular sequences of limestone and dolomite. Marginal facies of Windsor strata also commonly include polymictic conglomerate whose phenoclasts, like the Horton, are mainly pre-Carboniferous rocks. On Magdalen Islands, brecciated basalt, tuff, and agglomerate occur in the basal part of the Windsor, and in southern Antigonish–Mabou Basin a trachyte porphyry flow is present at the top of the basal Subzone A limestone. At Stellarton Gap, on the flank of Antigonish Upland, two thin flows of amygdaloidal basalt occur below Subzone E limestone, the lowest Windsor strata in that section. In Antigonish–Mabou Basin a few inches of bentonite are interbedded with basal Subzone A limestone. In St. George's Basin, above the Windsor-like section, but apparently correlative with Windsor strata of other areas, is 1,600 feet of dark grey shale, siltstone, and sandstone with thin beds of argillaceous limestone. These grey beds may be partly marine but they resemble

the lower part of the Canso Group that conformably overlie Subzone E limestone in Antigonish–Mabou Basin.

The oldest marine strata of Subzone A were limited to a long, narrow trough that embraced the Minas, Antigonish–Mabou, and St. George's Basins (Fig. VI-24). The basal member is a thinly laminated, fine-grained limestone 30–60 feet thick that is an important time-stratigraphic marker bed.

Subzone B sediments are the most widespread of the Windsor Group. During their deposition the sea covered all areas underlain by Subzone A strata and transgressed into the Cumberland and Moncton Basins and to Magdalen Islands. Marine strata were also deposited on the New Brunswick Platform during Subzone B time, but how the present isolated remnant was connected with Moncton Basin is unknown; possibly, the Windsor sea covered much of New Brunswick. However, the coarse clastics that occur along most present outcrop margins of the Windsor Group, which may be correlative with the Windsor Group but not necessarily with Subzone B, suggest that the maximum extent of the Windsor sea was not much greater than its present distribution. Subzone B deposits in St. George's Basin were probably more extensive, but rapid facies changes are common. Siltstone, limestone, evaporites, etc., outcrop along the present coast of Newfoundland and apparently grade to coarse clastics a few miles inland. Clastic rocks considered to be correlative with the Windsor Group, which are present in White Bay Basin, limit the extent of the Windsor sea in that direction.

Subzone C marine sediments apparently represent offlap relationships. The maximum extent of the Windsor sea during Subzone C time was less than during Subzone B time. Only one bed of Subzone C limestone is known in the Cumberland Basin, and above it are red clastics correlative in part with upper Subzone C or higher. Complete sections of Subzone C limestones are present in parts of Minas, Antigonish–Mabou, Sydney, and St. George's Basins and probably on Magdalen Islands.

Very few exposures are known of Subzone D limestones. They are probably more restricted than those of Subzone C, although present in Minas, Antigonish–Mabou, Sydney, and St. George's Basins.

Subzone E strata, generally the youngest of Windsor strata, have been found on Magdalen Islands and in St. George's, Sydney, Antigonish–Mabou, and Minas Basins,

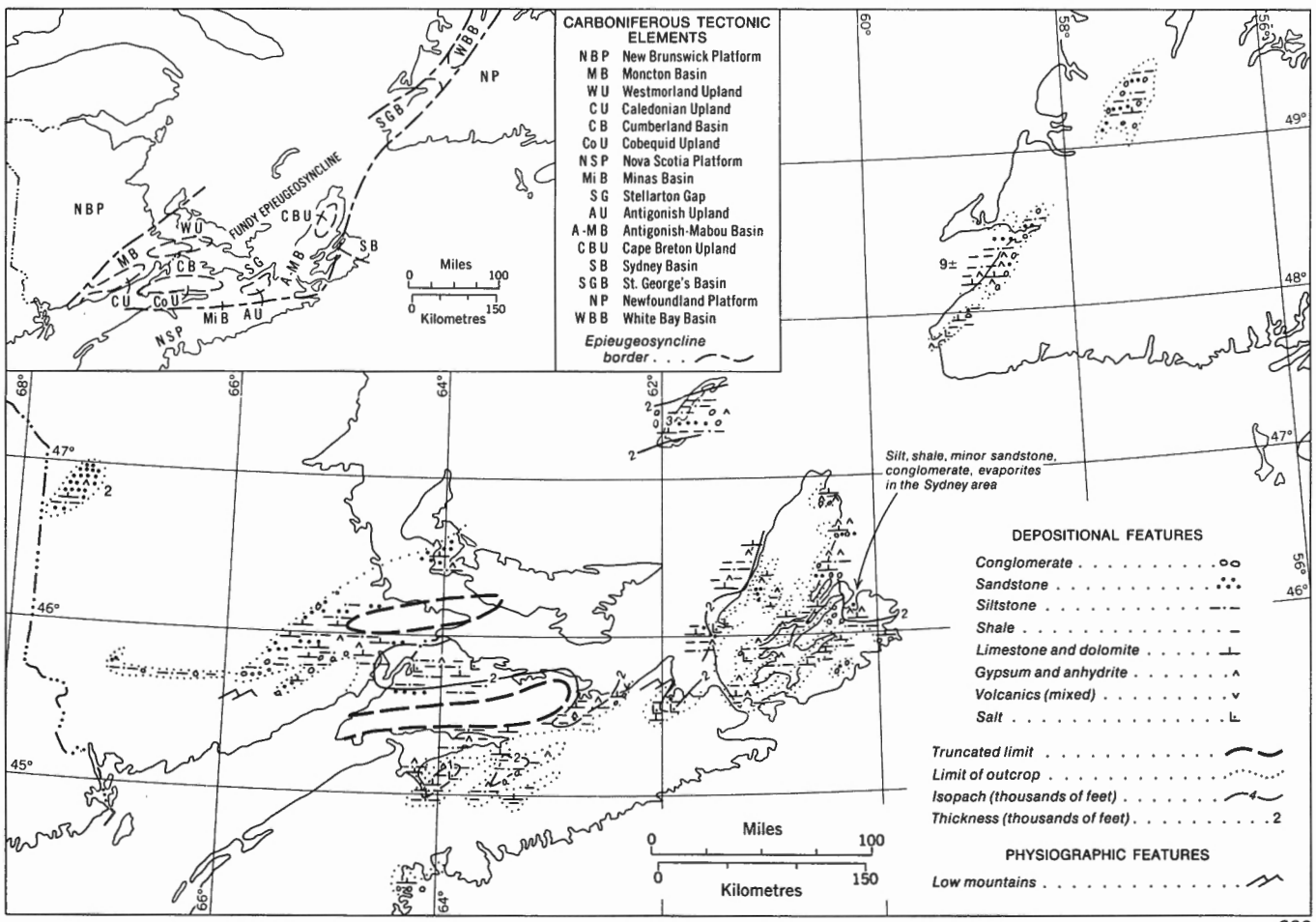

FIGURE VI-23. Carboniferous, mainly Viséan (Windsor Group), sedimentation, volcanism, and tectonism in Southeastern Canada.

at Stellarton Gap, and on Nova Scotia Platform. Subzone E limestone most commonly rests on a conformable Windsor section that includes all earlier subzones. In rare cases, however, it rests disconformably on Subzone B, as on Nova Scotia Platform, or unconformably on pre-Carboniferous rocks, as along the southeastern edge of Sydney Basin.

A single outcrop of limestone containing an assemblage of fossils assigned to Subzone F is present in a fault block near the northern boundary of Minas Basin. There is no evidence to support an age assignment younger than Subzone E, however. Withdrawal of the Windsor sea from the region soon after deposition of Subzone E limestone is apparent from the transitional contact between marine Windsor Group rocks and overlying partly brackish water rocks of the Canso Group.

Upper Carboniferous and Permian

The Canso Group (Fig. VI-25). This group is defined as the non-marine, red and grey shales that overlie marine Windsor Group or non-marine rocks of equivalent age and which, if fossiliferous, contain only early Namurian fossils. At the type locality along the eastern shore of the Strait of Canso in Antigonish–Mabou Basin the base of the Canso section is faulted against Windsor strata, and the top was arbitrarily chosen at a concealed interval. Later, the upper contact was extended upward to the base of a thick sandstone. The Canso, either as defined or redefined, is not a mappable unit, except possibly in the immediate vicinity of the type locality. The Canso "Group" in Nova Scotia consists of interfingering grey and red facies. The rocks are mainly fine-grained, red and grey sandstones, siltstone, and shale. Minor non-marine limestone occurs in the lower part of the group.

In Moncton Basin the upper part of the Hopewell Group, which consists of sandstones, siltstone, and conglomerate, has been correlated with the Canso "Group"; the lower part of the Hopewell Group has been correlated with the upper part of the Windsor Group. The Hopewell conformably overlies Lower Windsor strata of the Viséan age and is conformably to unconformably overlain by formations mainly of Westphalian A age.

In central New Brunswick Platform, 100 to 150 feet of conglomerate, tuff, and basalt are tentatively correlated with the Canso "Group". The basalt is commonly underlain by a few feet of coarse to fine red clastics that rest unconformably on pre-Carboniferous rocks. In the southwest the coarse clastics unconformably overlie red beds

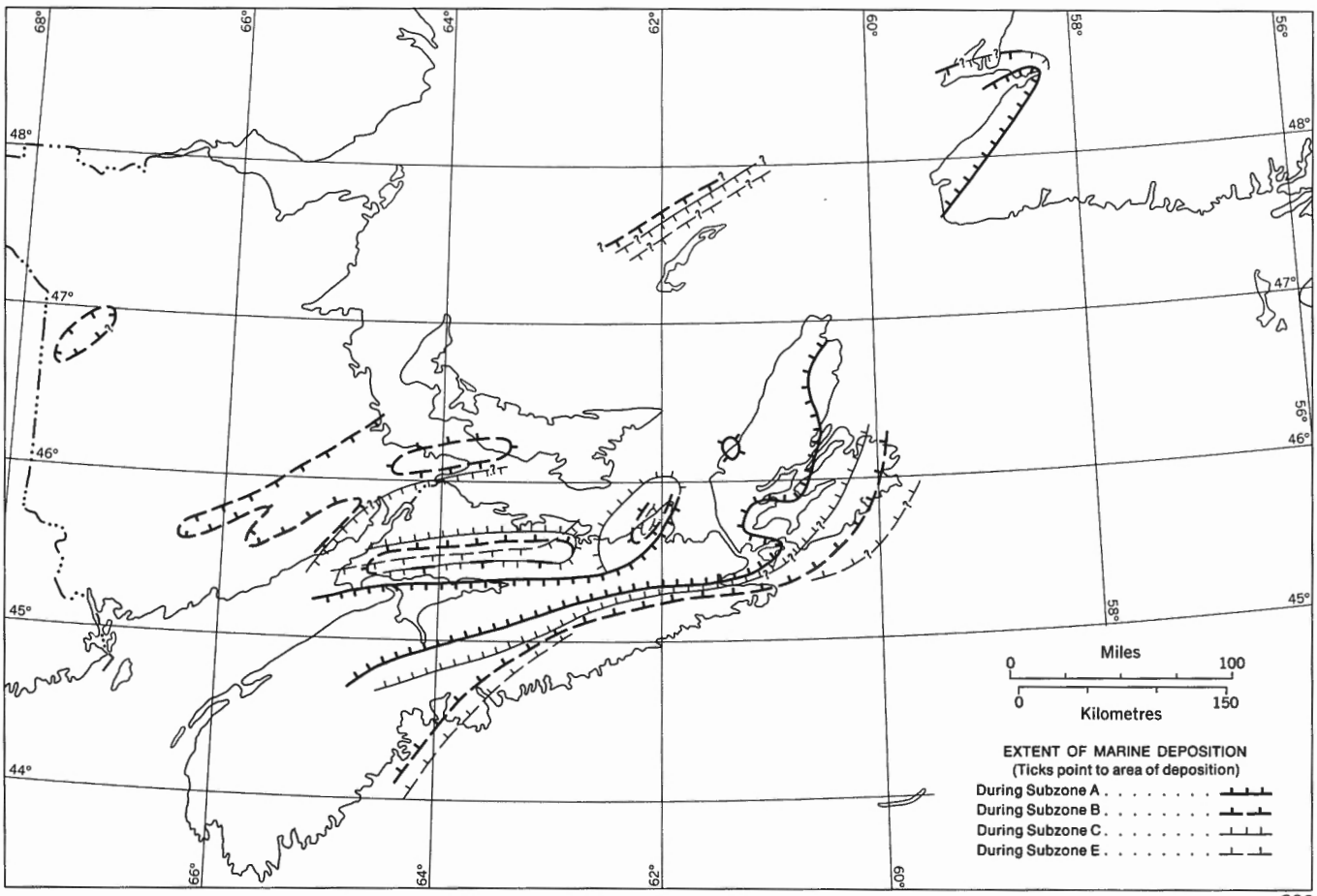

FIGURE VI-24. Extent of marine deposition of subzones of Windsor Group.

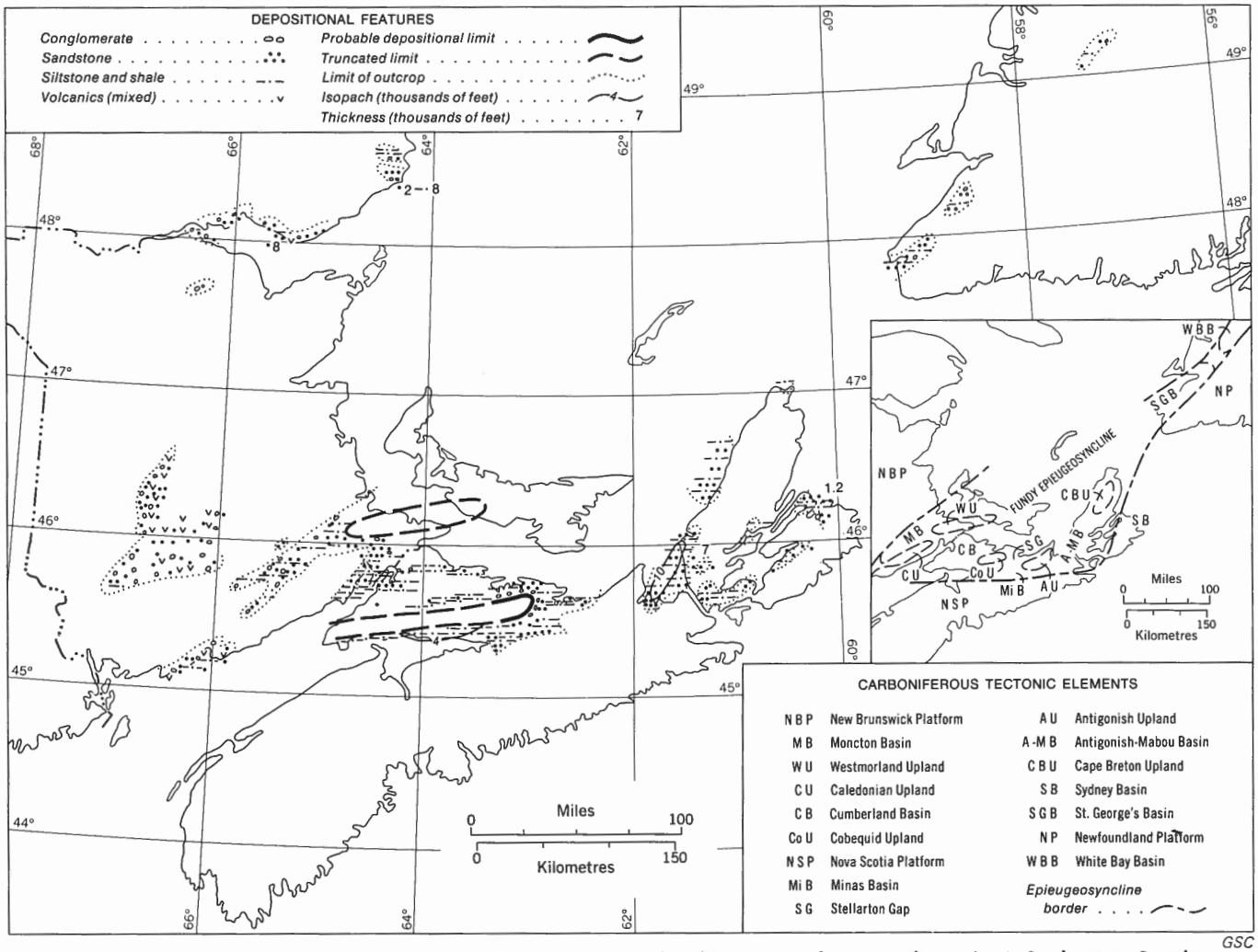

FIGURE VI-25. Carboniferous, Viséan to early Westphalian (Canso Group), sedimentation, volcanism, and tectonism in Southeastern Canada.

and rhyolite. These latter red beds are considered equivalent to a lower part of the Hopewell Group and the marine Windsor strata, and the rhyolite is correlated with the Horton Group. In northern New Brunswick and southern Gaspé Peninsula, conglomerate, sandstone, and shale with minor basalt and limestone of the Bonaventure Formation, and sandstone, shale, and conglomerate of the Cannes de Roches Formation are correlated with the Canso because of probable Namurian spores obtained from the Cannes de Roches. The Bonaventure Formation is probably the same age as the Cannes de Roches, but may have been deposited in a separate basin.

At the western end of Cumberland and Minas Basins in southern New Brunswick, the Mispek Group consists of conglomerate, sandstone, shale, tuff, basic and acidic intrusive and extrusive rocks. The Mispek is overlain by the Lancaster Formation, which contains an early Westphalian flora.

In St. George's Basin of Newfoundland, predominantly grey sandstone, shale, siltstone, and argillaceous limestone in the upper part of the Codroy Group, and conformably overlying predominantly grey sandstone, conglomeratic sandstone, siltstone, conglomerate, and minor silty limestone of the lower part of the Barachois Group have a flora similar to that of the Canso.

Deposition during the early Namurian in Minas, Sydney, and Antigonish–Mabou Basins appears to be mainly offlap sedimentation in areas of low relief whereby each succeeding bed is of increasingly restricted extent, relative to immediately underlying beds. Uplift of part of Cobequid Upland is indicated by conglomerate around most of the eastern end of the present Cobequid Mountains. Along the northern flank fine-grained red and grey sandstone and siltstone overlie a few feet of conglomerate that rests on pre-Carboniferous rocks.

The Riversdale Group (Fig. VI-26) includes the continental red and grey sandstone and shale that lie stratigraphically above the Canso "Group" and below the Cumberland Group. All strata correlative with the Riversdale "Group" are assumed to have the same age limits as the Riversdale. It is of late Namurian to early Westphalian

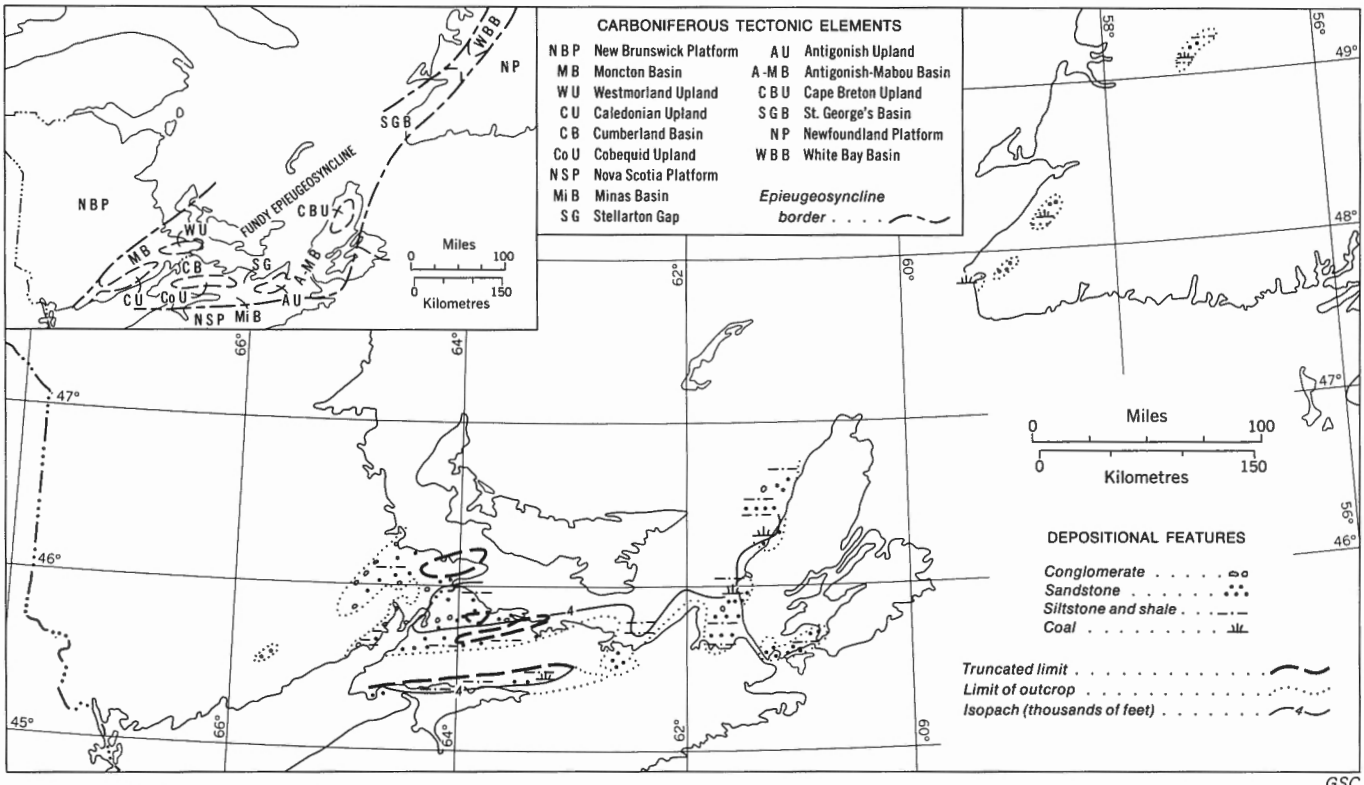

FIGURE VI-26. Carboniferous, mainly early Westphalian (Riversdale Group), sedimentation and tectonism in Southeastern Canada.

A age and in places is separated from older and younger strata by a disconformity or unconformity. Correlation of rocks with the Riversdale is based solely on the floral or faunal assemblage. The type locality is near Riversdale in Minas Basin, and the section exposed there is faulted so that neither the top nor the base is known. The Riversdale has not been divided into formations in the type area, but Riversdale or correlative strata have been subdivided in Cumberland and Moncton Basins. The lower unit is the Claremont Formation in Nova Scotia or the Enrage Formation in New Brunswick, and the upper is the Boss Point Formation. Local formational names, such as Port Hood Formation, have commonly been used interchangeably with the Riversdale. In St. George's and White Bay Basins, only part of the Barachois Group is correlative with the Riversdale.

Lithologically the Riversdale "Group" in the basins of Nova Scotia and New Brunswick consists of fine and coarse facies. In Minas Basin 4,000 to 5,000 feet of dark grey shale and siltstone with thin beds of sandstone and minor fine-grained red beds were originally assigned to the Riversdale. These clastics are similar to underlying Canso strata, and recent work has partially demonstrated that they belong to the same rock-stratigraphic unit as the Canso. Elsewhere the Riversdale consists of fine- to coarse-grained sandstone, shale, siltstone, and conglomerate with a local basal quartz-pebble conglomerate. The sandstone is commonly feldspathic, grey, yellow weathering, and in thick, massive beds that are commonly crossbedded and carry drift plant stems. Mineable seams of coal are present in the Mabou–Antigonish Basin, and thin seams are present in St. George's and White Bay Basins. The conglomerate at the base in Cumberland and Moncton Basins is apparently an alluvial fan that developed during uplift in either the late Namurian or early Westphalian A. The coarser facies of the Riversdale suggests rejuvenation of the source areas. The type of contact and spatial relationship between Riversdale strata and pre-Carboniferous rocks of upland areas are unknown, except in New Brunswick where the Enrage Formation was deposited on pre-Carboniferous rocks on the southeast side of Caledonian Upland. The Enrage and overlying Boss Point Formations are conformable where both are present in axial parts of Moncton and Cumberland Basins; in the axial part of Cumberland Basin they lie conformably on the Canso "Group". Where the Enrage is missing, in Moncton Basin, the Boss Point lies unconformably on all earlier Carboniferous rocks. The general coarsening of Riversdale lithology is therefore probably caused by local uplifts and rejuvenation of source areas.

The Cumberland Group (Fig. VI-27). This group was first defined to designate strata that lie either unconformably or disconformably above the Boss Point Formation of the Riversdale Group and unconformably below the Pictou Group. In the type section at Joggins in western Cumberland Basin over 9,000 feet of strata is present.

At Stellarton Gap the Cumberland "Group" is dominantly red sandstone but includes a basal fanglomerate that lies unconformably on Namurian strata.

In western Cumberland Basin the group consists of a basal, coarse, predominantly conglomeratic facies; a lower, fine-grained facies of grey shale, sandstone, and mineable coal seams; an upper, coarse conglomerate and sandstone facies; and fine clastic facies of red sandstone and siltstone at the top of the section. In southwestern New Brunswick red and grey sandstone, shale, grit, and conglomerate occur.

The floral assemblages of the Cumberland Group have been correlated with late Westphalian A or early Westphalian B. Megaflora and megafauna in the type area are mostly from the lower, fine-grained facies that contain the economically important coal seams. The age of strata in the upper part of the type section of Cumberland Group is possibly the same as that of the lower part of the Pictou "Group".

Deposition in Cumberland Basin was apparently very rapid in some places as suggested by the preservation of 30-foot trees in an upright position. Conglomerate was derived from the Cobequid Upland, the earliest indication that the western end of the Cobequid Mountains was a source area. The eastern end had been a source for a fanglomerate, probably in Namurian time, and the eastern Cobequids or Antigonish Upland was also the source for the Cumberland fanglomerate. It appears likely that Moncton Basin was static during late Westphalian A or early Westphalian B time as the Cumberland or equivalent strata in the Petitcodiac Group are thin or absent.

The Pictou Group (Fig. VI-28). In the type locality along River John in Nova Scotia the Pictou Group unconformably overlies rocks of Namurian age and embraces the highest beds in the axis of a syncline in Cumberland Basin. Its use, however, has been based on floral content, and although diagnostic flora were not present in the type section, which is predominantly red beds, the name was applied to all strata containing the characteristic flora. The flora is divided into three zones: the *Lonchopteris* zone of late Westphalian B to early Westphalian C, the *Linopteris obliqua* zone of Westphalian C to early Westphalian D, and the *Ptychocarpus unitus* zone representative of Westphalian D. Fossil spore age assignments are in agreement, and also add two additional zones representative of the Stephanian and early Permian. The Pictou Group in Cumberland Basin consists of about 8,000 feet of mainly red sandstone and siltstone, shale with arkosic grit, and minor grey sandstone, siltstone, and shale. Red beds are found at Stellarton Gap, along with grey sandstone and shale with oil shale and mineable seams of coal. These beds have been called the Stellarton Group.

Strata in Sydney Basin correlative with the Pictou are referred to the Morien Group. The group includes older strata than those at the Pictou type section and lies

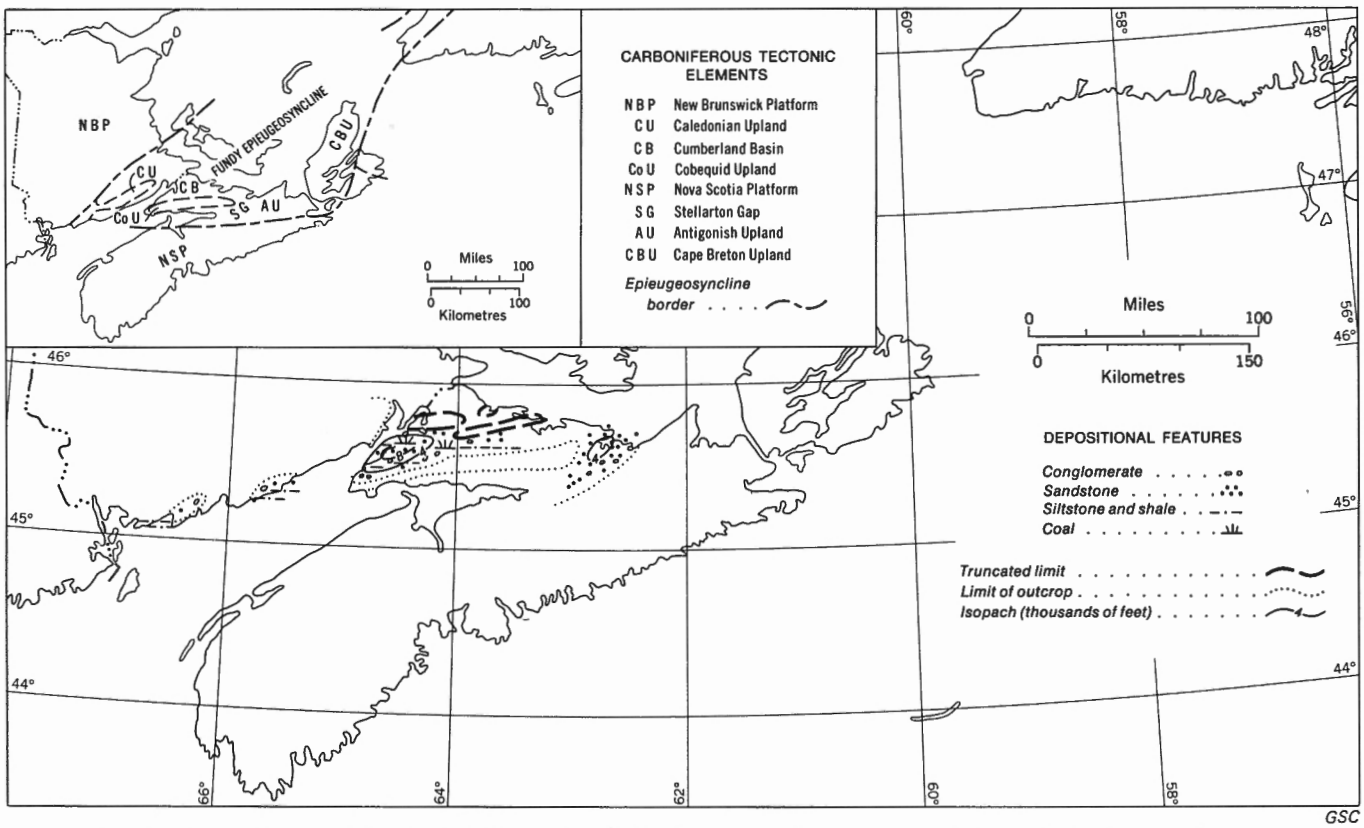

FIGURE VI-27. Carboniferous, early Westphalian (Cumberland Group), sedimentation and tectonism in Southeastern Canada.

disconformably on strata of Namurian and Viséan age, the Point Edward Formation and the Windsor Group. The group consists of 5,000 to 6,500 feet of grit, sandstone, conglomerate, siltstone, shale, and mineable coal seams. It thins to 2,000 to 2,500 feet in western Cape Breton Island. The coal seams are the most productive in the Atlantic Provinces.

Pictou strata in Moncton Basin and on New Brunswick Platform are 1,000 to 3,000 feet thick and consist predominantly of red and grey arkosic sandstone or mudstone, conglomerate, and minor shale. They are oldest in the southwest part of the platform and youngest in the southeast part near the present coastline. Their accumulation was mainly a lateral accretion or building outward of sediments toward the east or northeast, although not necessarily as a delta.

In two small areas in Antigonish–Mabou Basin Pictou strata unconformably overlie Windsor and Horton Groups and overlap pre-Carboniferous rocks.

The red beds of Prince Edward Island have generally been considered part of the Pictou even though they are partly Permian. The red sandstone on Magdalen Islands is tentatively considered part of the same sequence. In western Newfoundland, grey sandstone, siltstone, and conglomerate of Westphalian C or D age occur in a small area and are mainly drift covered.

Pictou strata, in general, constitute a transgressive sequence that spread laterally over all older rocks. The earliest Pictou strata (*Lonchopteris* zone) are probably present in all basins except possibly the Minas. The earliest deposition may have commenced without break in sedimentation at Stellarton Gap and in western Cumberland Basin. An unconformity between the Cumberland and Pictou Groups is present at one locality on a piercement salt structure in Cumberland Basin. It is interpreted to be of local significance only, and is related to the salt intrusion. No alluvial fan-type deposits have been recognized in Pictou strata except in Cumberland Basin where alluvium derived from Cobequid Upland may have continued into Westphalian C time. Otherwise there is no evidence of tectonic activity in the upland areas during Pictou deposition. Pictou strata are structurally conformable on earlier Carboniferous rocks on the platform areas and in Moncton and Sydney Basins.

Maritime Disturbance

Structure. Open folds and high-angle faults, whose latest movements were in Carboniferous and post-Carboniferous time, are the most common structural features in the region. Reverse and strike-slip faulting accompanied by folding occurred within the Fundy Epieugeosyncline during Carboniferous sedimentation as a result of the interaction between pre-Carboniferous basement rocks and rocks in the Carboniferous basins (Fig. VI-29). It appears that Carboniferous structures were mainly a response to deformation in the pre-Carboniferous basement.

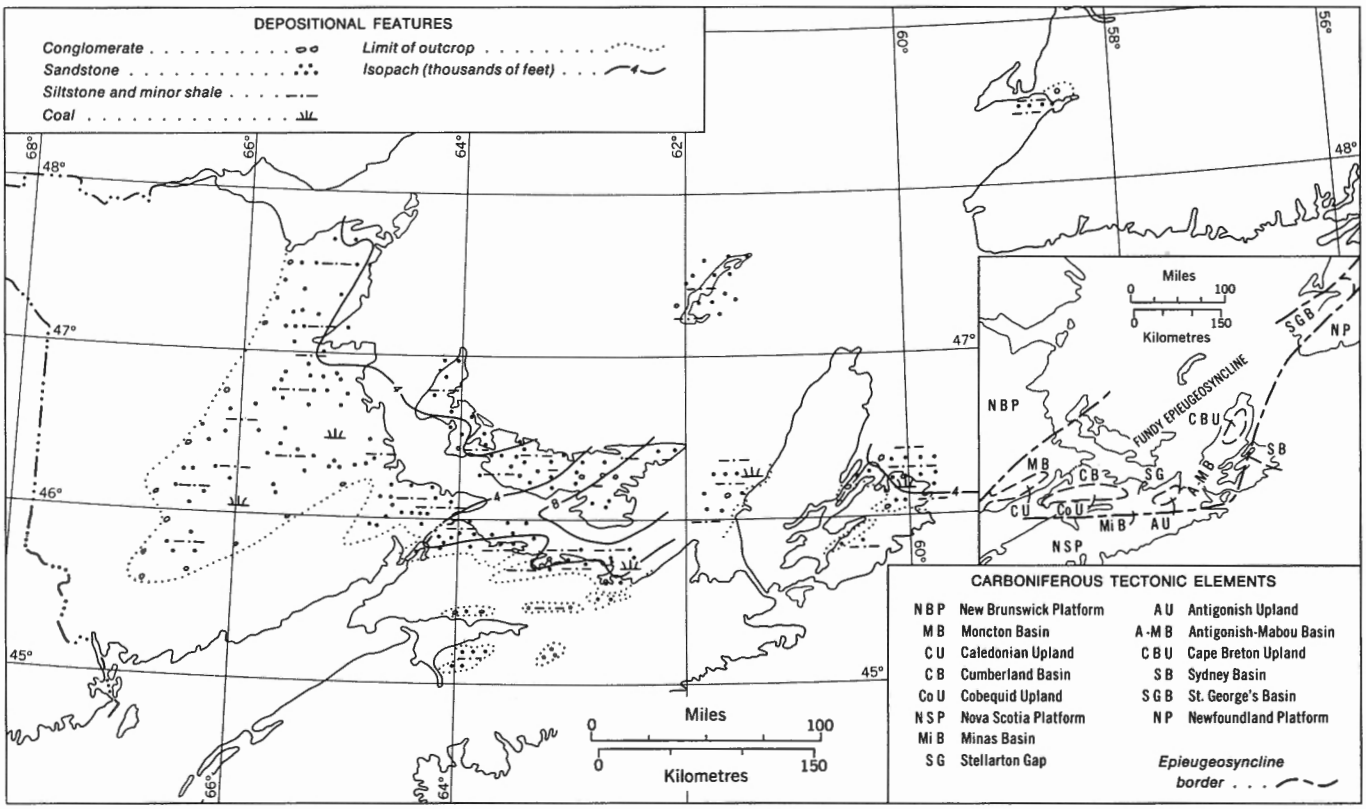

FIGURE VI-28. Carboniferous and Permian, late Westphalian to early Permian (Pictou Group), sedimentation and tectonism in Southeastern Canada.

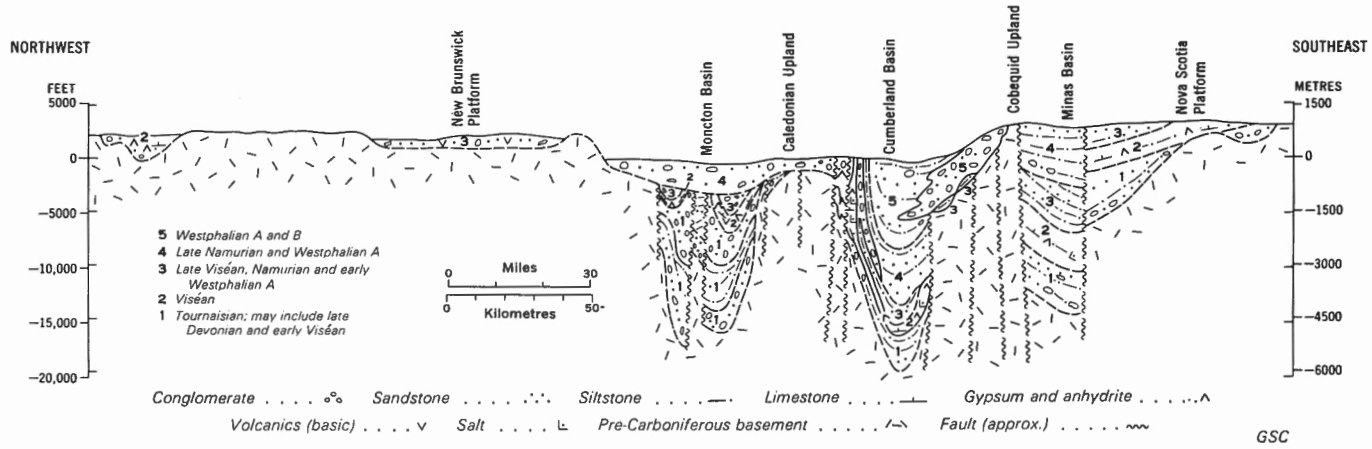

FIGURE VI-29. Diagrammatic restored section of Carboniferous rocks at end of Westphalian B time, New Brunswick and Nova Scotia (by D. G. Kelley).

It has been suggested that the dominant folds are due to gravity sliding, and they may be in part, but evidence of a similar stress pattern in Carboniferous and pre-Carboniferous rocks along the northern boundary of Minas Basin suggests that regional compression played a major role in the development of folds.

The trends of the very broad open folds in Pictou rocks, which essentially blanket earlier rocks, are parallel to the dominant northeast to east-northeast trends in the more intense folds of earlier Carboniferous rocks. Some are overturned. North-trending secondary structures, including folds, were impressed on earlier Carboniferous folds. These cross-folds trend at high angles to the earlier folds and locally are dominant.

A diapiric rise of salt accompanied and accentuated folds that were developed during the Upper Carboniferous. Seismic data tend to confirm that faulting was also a factor in the development of salt structures in the epieugeosyncline.

Metamorphism and igneous intrusion. Metamorphism of Carboniferous and Permian rocks is negligible except for those in a narrow, intermittent zone that extends from the southwest end of Cumberland Basin in southern New Brunswick, east along the southern boundary of Cobequid Upland, and continues out to sea along the same trend. In several places along this zone the Carboniferous rocks have a fracture cleavage. At Cape Spencer near Saint John, New Brunswick, and at several other places in the area underlain by the Carboniferous Mispek Group, there are several granitic stocks and sills. The granite, which is locally miarolitic, intrudes the volcanic and sedimentary rocks of the Mispek Group; both the granite and the intruded rocks are highly fractured, the deformation being late Carboniferous or younger. The granite is probably related genetically to the acidic volcanics of the Mispek Group.

In the Antigonish–Mabou Basin several small stocks and plugs of gabbro cut Lower Windsor strata and may be of Carboniferous age. There are also a number of diabase dykes and sills that cut the lower part of the Horton Group with little visible effect on the enclosing rocks. These dykes are probably Tournaisian.

MESOZOIC AND CENOZOIC

Tectonic Summary

The Mesozoic record of Southeastern Canada is scant (Fig. VI-30). Late Permian to Jurassic rocks are lacking on St. Lawrence Platform and the Appalachian Geosyncline, except for some Upper Triassic non-marine clastics and basalt in the taphrogenic Cape Split Trough of Nova Scotia and southern New Brunswick. The faulting and gentle warping of the trough rocks are referred to the Palisades Disturbance of Late Triassic and possibly younger age. Diabase dykes of Late Triassic and Permo–Triassic age were intruded in southern Nova Scotia and on Prince Edward Island, respectively.

During the Jurassic the entire region, including the Gulf of St. Lawrence and continental shelf, was uplifted and undergoing erosion. In the early Cretaceous and possibly late Jurassic, the basic and alkaline stocks and dykes of the Monteregian Intrusions were intruded along a west-trending zone from within Appalachian Orogen, across St. Lawrence Platform to the Canadian Shield. The platform and Shield were cut by normal faults of similar trend, and some Paleozoic rocks were preserved as outliers in the downfaulted blocks. A diabase dyke intruded lower Paleozoic strata on Anticosti Island, and lamprophyre dykes intruded Acadian-folded

rocks of northeastern Newfoundland and the Canadian Shield at Kirkland Lake.

During the Cretaceous the Paleozoic cover rocks on the Canadian Shield were eroded to about their present distribution. Lower Cretaceous non-marine clastics were deposited unconformably on the Paleozoic strata of Hudson Platform and on Aphebian rocks of Labrador Fold Belt near Schefferville. In the Appalachian Region, Lower Cretaceous fluviatile and lacustrine clastic sediments and lignite, unconsolidated and undeformed, were deposited in a deep river channel cut into an older, possibly Jurassic, upland surface. Upper Cretaceous sediments were deposited on the Scotian Shelf, and possibly also on other parts of the Atlantic Continental Shelf. Marine seismic surveys seem to detect two troughs separated by a ridge—one beneath the shelf and the other beneath the slope and rise. The age of the basement is unknown, but may be Acadian deformed rocks and granites.

The Tertiary record in Southeastern Canada is as meagre as that of the Mesozoic. Tertiary strata have not been recognized, except possibly a conglomerate in Nova Scotia. The entire region may have been low land with thick Tertiary sediments deposited in the troughs of Scotian Shelf and probably also other parts of the continental shelf.

During the Quaternary (*see* Chapter XII), Pleistocene ice sheets advanced southward over the Canadian Shield, St. Lawrence Lowlands, Appalachian Region, and onto the Scotian Shelf, most of which was exposed by the eustatic fall of sea level of at least 400 feet. The island of Newfoundland supported its own ice cap which flowed radially toward the sea. From a glacial maximum about 20,000 years ago, the glaciers began retreating and the sea advanced upon the land, reworked the glacial materials and deposited marine silt and clay along some major river valleys and glacial lake silt and clay in the Great Lakes region. Eustatic fall of sea level and glacial rebound brought Southeastern Canada to its present elevation. Within the last few thousand years, parts of coastal Nova Scotia have subsided.

Triassic

Upper Triassic rocks in the Cape Split Trough outcrop extensively in northern Nova Scotia and southern New Brunswick bordering Bay of Fundy, and locally in Chedabucto Bay. They underlie much of Bay of Fundy and extend southwestward beneath the Gulf of Maine. They are covered by a discontinuous layer of

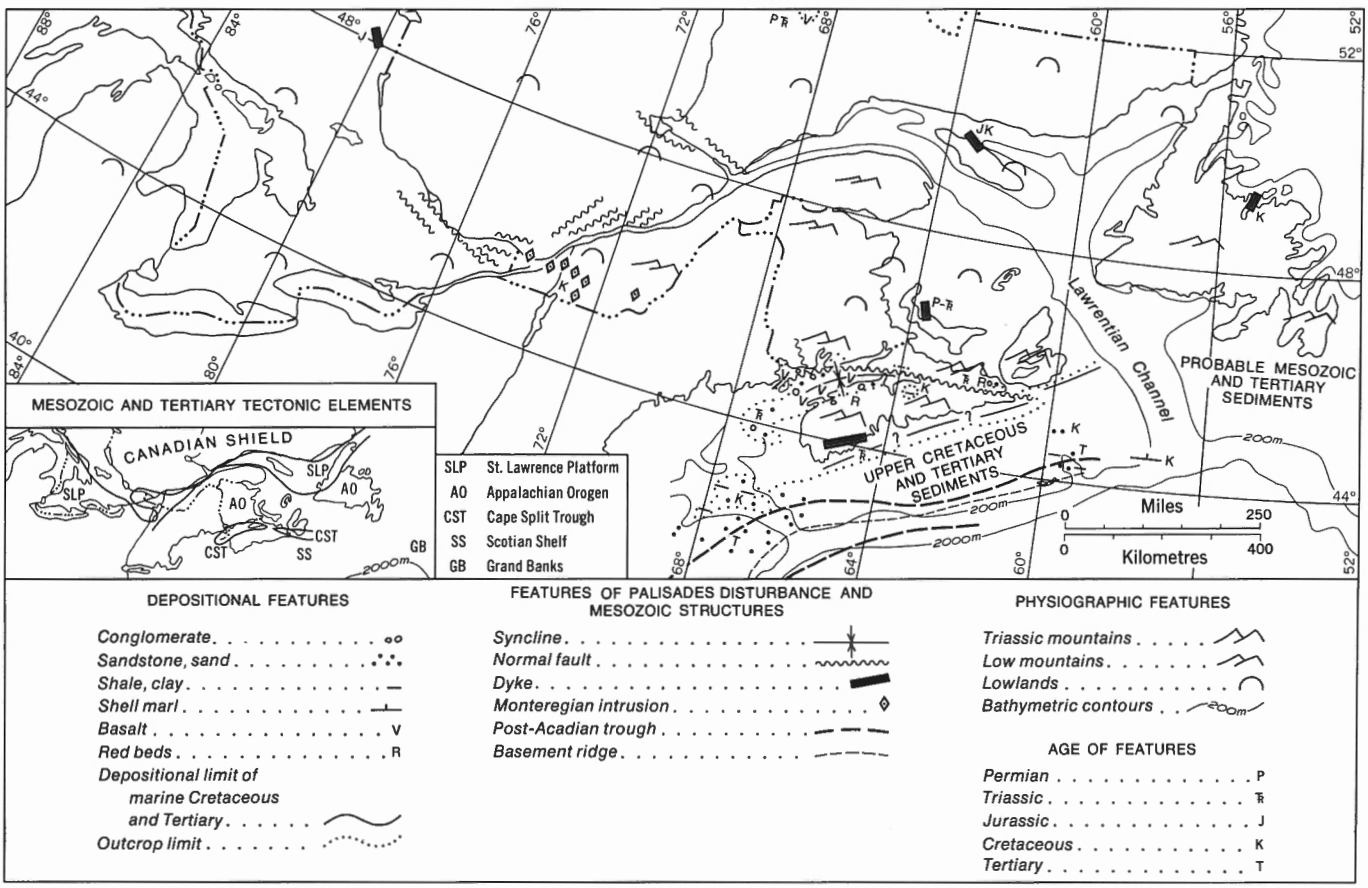

FIGURE VI-30. Mesozoic and Tertiary sedimentation and volcanism in Southeastern Canada, with features of late Triassic Palisades Disturbance and other structures.

Quaternary deposits as much as several hundred feet thick. East of Chedabucto Bay marine seismic and gravity studies suggest that some of the 3,700 feet of consolidated sedimentary rock with a compressional wave velocity about 4.3 km per sec. lies in a trough above possible lower Pennsylvanian strata. Triassic sedimentary rocks may occur in the troughs on Scotian Shelf and elsewhere on the Atlantic Continental Shelf.

In northern Nova Scotia, Upper Triassic rocks of the Fundy Group comprise two sedimentary units separated by basalt flows (Klein, 1962). The lower sedimentary unit, formerly called the Annapolis Formation, consists of the Wolfville and Blomidon Formations. The Wolfville overlies Carboniferous and older rocks with angular unconformity and is made up of 200 to 2,500 feet of red and brown pebble and boulder polymictic conglomerate, greywacke, orthoquartzitic sandstone, and shale. It is of fluviatile origin and exhibits cut-and-fill structures, crossbedding, ripple-marks, current lineation, and pebble imbrication. The conformably overlying, and in part laterally equivalent, Blomidon Formation consists of 25 to 1,200 feet of even-bedded, reddish and greenish grey sandstone and thin-bedded mudstone. Groove casts, salt casts, rain prints, ripple-marks, mudcracks, crossbedding, and tree trunks suggest a mainly lacustrine facies that intertongues with the Wolfville fluviatile facies. The conformably overlying North Mountain Basalt consists of 875 feet or more of tholeiitic flows containing amygdules filled with a varied assemblage of zeolites. Five K–Ar whole-rock dates average 198 m.y., that is, latest Triassic. The upper sedimentary unit, the Scots Bay Formation, occurs at the water's edge of the Bay of Fundy. It consists of 23 feet of grey and brown sandstone, purple and grey mudstone, and limestone with chert and jasperoid nodules, probably of lacustrine origin. Fossils from all the sedimentary formations are non-marine forms, including conchostracan shells, reptillian bones, vertebrate teeth, fish scales, and ferns; they are late Triassic, probably Karnian and Norian. Red sandstone and conglomerate, lithologically similar to the Wolfville, outcrop on Chedabucto Bay and in southern New Brunswick. Basalt outcrops in Minas Basin and on Grand Manan Island.

Two diabase dykes are late Permian or Triassic. The one on Prince Edward Island is a gently dipping dyke or sill, at least 10 feet thick, intrusive into red sandstone of the Pictou Group. The K–Ar whole-rock age is 207±8 m.y. (N. J. Snelling), about Middle Triassic, whereas paleomagnetic pole positions suggest a probable late Permian age. Another diabase dyke, in southern Nova Scotia, is 200 to 500 feet wide and 70 miles long. It has a whole-rock K–Ar date of 197±32 m.y., on or near the Triassic–Jurassic boundary. The paleomagnetic pole position is similar to that of North American Triassic rocks including the North Mountain basalt lavas to which the dyke has been related.

Jurassic and Cretaceous

Northeast-trending lamprophyre dykes cut folded Silurian sedimentary rocks and Devonian granite at Notre Dame Bay, Newfoundland. K–Ar age determinations on biotite and hornblende range from 115 to 144 m.y. and average 133 m.y., which corresponds to earliest Cretaceous. A northwest-trending diabase dyke cuts Upper Ordovician limestone on Anticosti Island and yielded a K–Ar whole-rock date of 138 m.y., that is on or near the Jurassic–Cretaceous boundary.

The Monteregian Intrusions cut folded rocks of the Appalachian Orogen, the gently deformed to flat-lying strata of the St. Lawrence Platform and Precambrian crystalline rocks of the Grenville Province of the Canadian Shield. With related dykes they occur along a west-trending line for about 150 miles. Most are roughly circular plugs with steep walls, but others are considered to be laccoliths. Near Montreal there are diatreme breccias comprising various proportions of igneous material and inclusions of Precambrian to Lower Devonian country rock. The Devonian limestone fragments must have dropped down within the diatreme because no strata younger than Ordovician outcrop in Quebec Basin. The intrusions and dykes consist of a great variety of alkaline rocks among which are nepheline, syenite, essexite, nordmarkite, pulaskite, yamaskite, and rougemontite. Monzonitic and granitic rocks form part of the easternmost intrusion near Lake Megantic. The Oka Intrusion contains abundant carbonatite and is mined for the niobium-bearing mineral pyrochlore. Several K–Ar and Rb–Sr age determinations obtained from minerals and whole-rock samples range from 84 to 123 m.y. with many dates falling in the 100 to 115 m.y. range corresponding to the late Lower Cretaceous. A northwest-striking kimberlite dyke at Kirkland Lake may be related to the Monteregian Intrusions. A K–Ar age determination on phlogopite yielded 151±8 m.y., corresponding to late Jurassic.

Cretaceous sediments occur in Nova Scotia and on the Scotian Shelf. Non-marine Lower Cretaceous sediments, probably pre-Albian, consisting of about 60 feet of red, brown, grey, and black interbedded clay, sand, silt, and lignite containing pollen and spores, are exposed along Musquodoboit River and in an open pit near Shubenacadie in central Nova Scotia (Stevenson, 1959). The sediments are probably of lacustrine and fluviatile origin and fill a valley cut more than 200 feet below the upland surface. They probably lie unconformably on the Windsor Group and are overlain by glacial deposits. The clay has been used in the manufacture of pottery and buff firebrick.

Upper Cretaceous sediments may underlie much of Scotian Shelf and Grand Banks, and perhaps the continental shelf off northeastern Newfoundland and Labrador. They are less likely to occur in Gulf of St. Law-

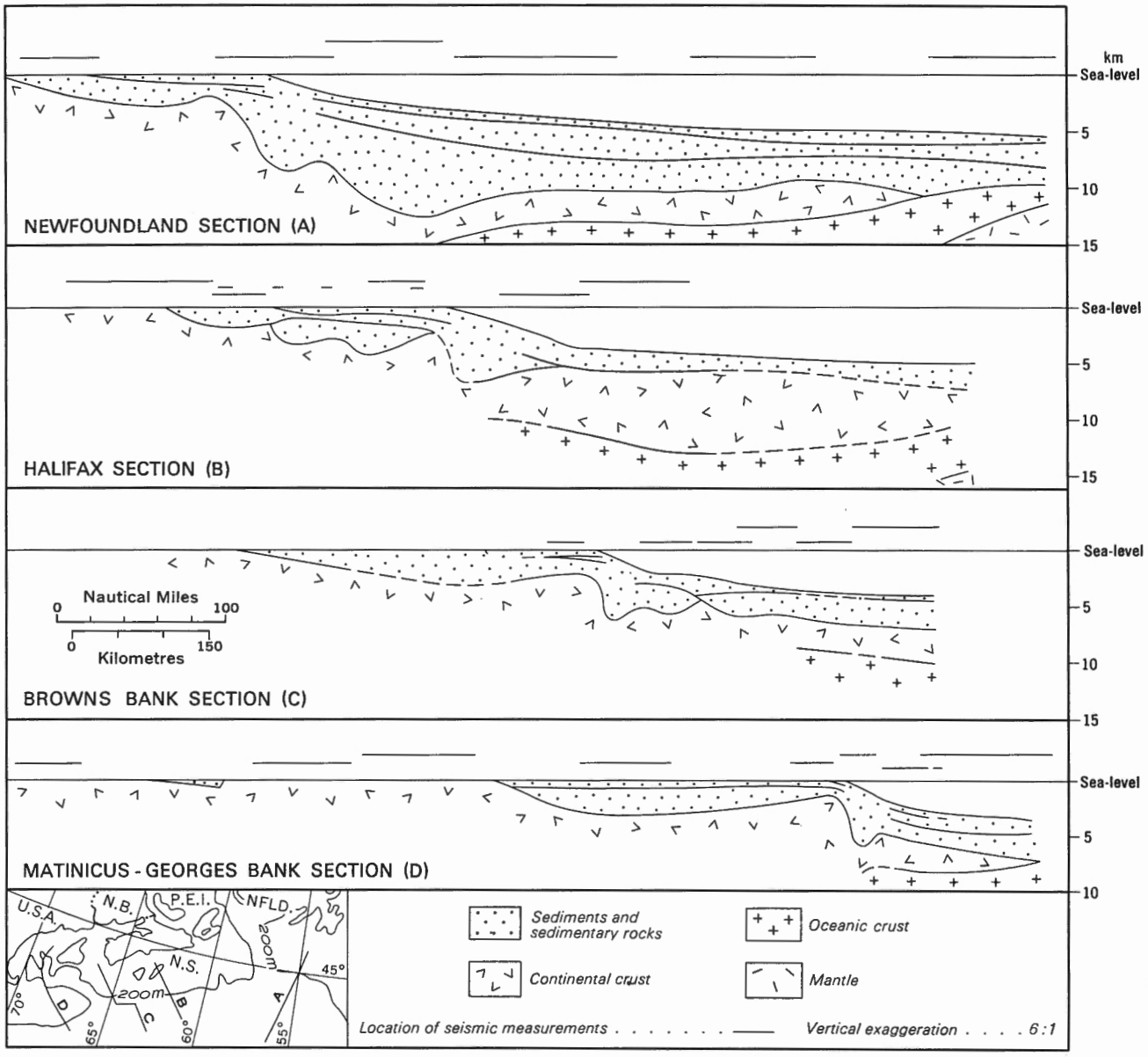

FIGURE VI-31. Structure-sections from seismic refraction measurements, Atlantic Continental Shelf (Drake, et al., 1959).

rence as much of the sea floor between Magdalen Islands and Prince Edward Island is underlain by red sandstone that is probably Carboniferous and Permian. On eastern Banquereau Bank of Scotian Shelf several fragments of indurated, highly fossiliferous, finely arenaceous shell marl were lifted from 200 fathoms of water. The fragments were considered to have fallen from upslope subsea outcrops. The fauna are marine and of Upper Cretaceous age, probably Cenomanian. A greenish grey, impure sandstone was dredged from a subsea peak on central Scotian Shelf north of Sable Island (L. H. King). The fauna are probably Cenomanian (J. A. Jeletzky). The marine late Upper Cretaceous micaceous and glauconitic sandstones dredged from Gulf of Maine probably extend northeastward to the Scotian Shelf and Grand Banks. The Upper Cretaceous could be several thousand feet thick if the strata form part of the troughs that occur beneath the continental shelf, continental slope, and continental rise (Fig. VI-31).

Tertiary

Tertiary sediments have been dredged and cored at 425 to 488 fathoms from the sides of The Gully at the edge of the Scotian Shelf (Marlowe, 1967). They consist of dark brown, hard and fissile silt with foraminifera of late Tertiary, probably Miocene age. A cobble of coarse glauconite sandstone of probable Pliocene age was recovered from the bottom of The Gully. There may be

Tertiary sediments on other shelf areas, including parts of Gulf of St. Lawrence, as the outline of the present land areas seems to be similar to those that could have existed in the Tertiary.

Possible Tertiary stratified grit and gravel deposited by streams and firmly cemented by limonite outcrop in Nova Scotia at Maitland, Greenfield, and between Bridgewater and Halifax. They lie directly on pyritic Halifax slate, and although they underlie the glacial material, they could be Pleistocene outwash deposits.

Quaternary

Pleistocene moraines have been identified in the central part of Scotian Shelf, and abundant igneous and metamorphic clasts of probable glacial origin have been recovered from the southeastern shelf and slope and in canyons like The Gully. Similarly, the floor of Gulf of St. Lawrence is veneered by glacial materials. The Laurentian Channel is a pre-glacial valley that was modified by the glaciers. Retreat of the ice allowed the rising sea to invade main river valleys and deposit marine clays, and lacustrine clay was deposited in the Great Lakes region. In the past few thousand years many coastal parts of Nova Scotia and Prince Edward Island have been depressed, resulting in drowned forests.

In historical times there have been only a few earthquakes in Appalachian Orogen, generally of low intensity. The St. Lawrence Platform and adjacent Canadian Shield are seismically active near Quebec City and along a belt that extends through Montreal to Boston, Massachusetts (Fig. VI-32). The concentrations of earthquake epicentres are apparently not controlled by near-surface geological elements.

Beneath the Appalachian Orogen, approximately along the axial region, a belt of thick crust has been detected by marine refraction seismic surveys (Dainty, *et al.*, 1966). Within the belt (Fig. VI-33), the crust is about 45 km thick and overlies an upper mantle with

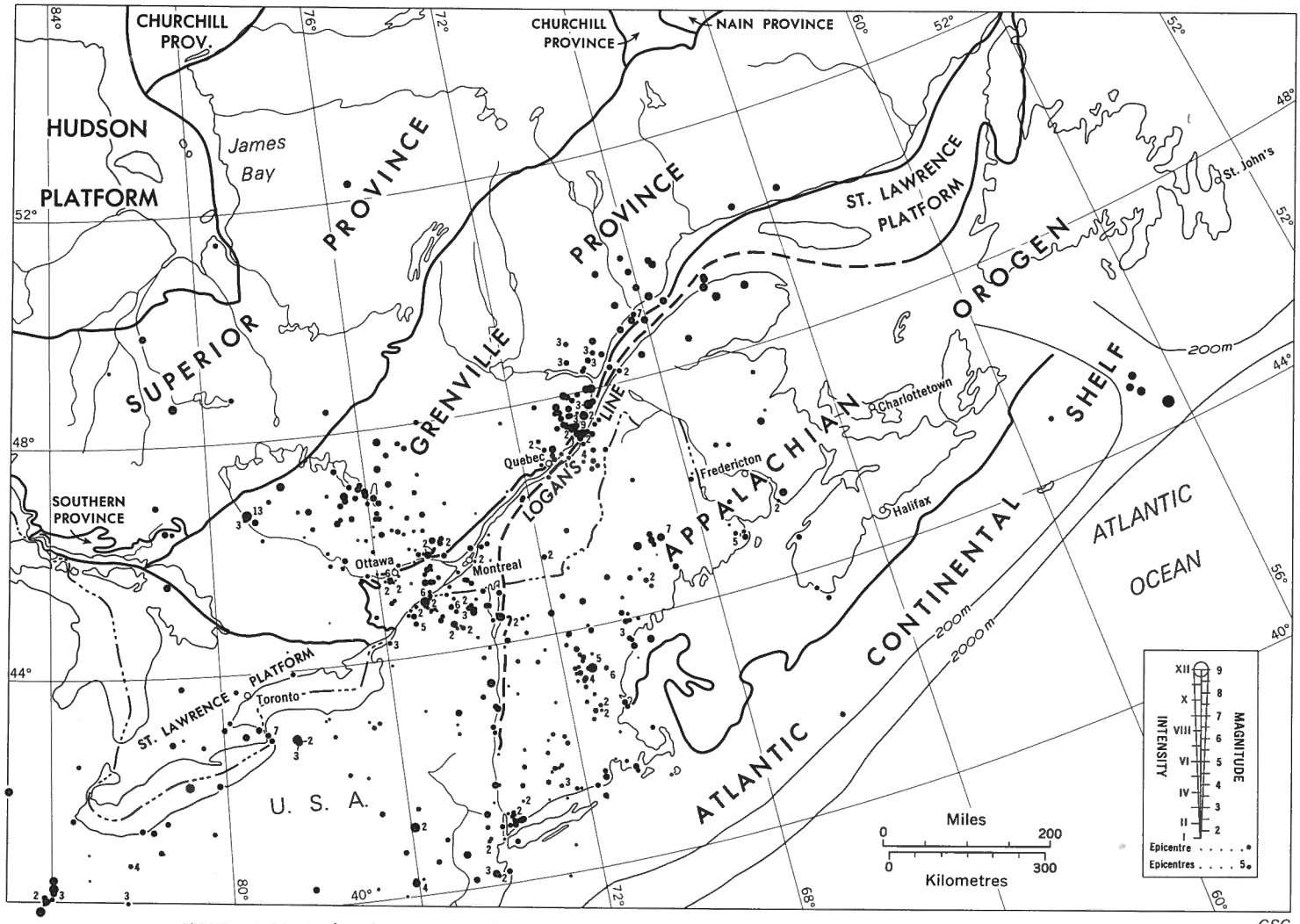

FIGURE VI-32. Earthquake epicentres of eastern Canada and adjacent areas, 1928–59 (modified from Smith, 1967).

high compressional wave velocities of 8.5 to 8.7 km per sec. In the lower part of the crust an intermediate layer has velocities of 7.3 to 7.5 km per sec. The crust flanking the thickened belt is 30 to 40 km thick. It lacks the intermediate layer, and overlies an upper mantle with the usual velocities of about 8.0 km per sec. The thick crustal belt corresponds closely to Notre Dame Trough, a persistently negative tectonic element of Appalachian Geosyncline within which thick eugeosynclinal sedimentary, volcanic, and ultramafic rocks accumulated, and which was intensely deformed during the climactic Acadian Orogeny.

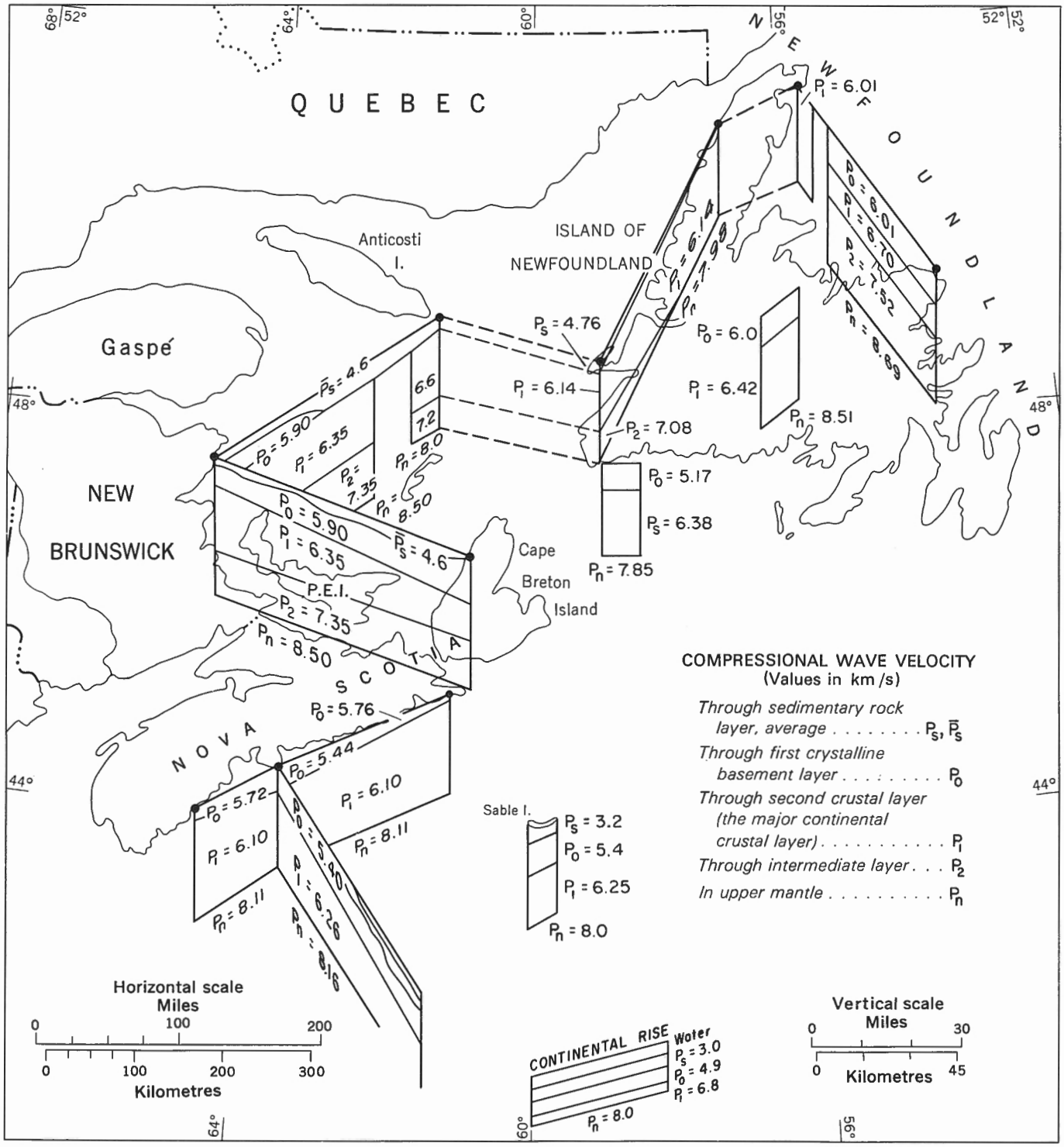

FIGURE VI-33. Fence diagram through crust, Appalachian Orogen (Dainty, et al., 1966).

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VII. Economic Minerals of Southeastern Canada

Introduction 306

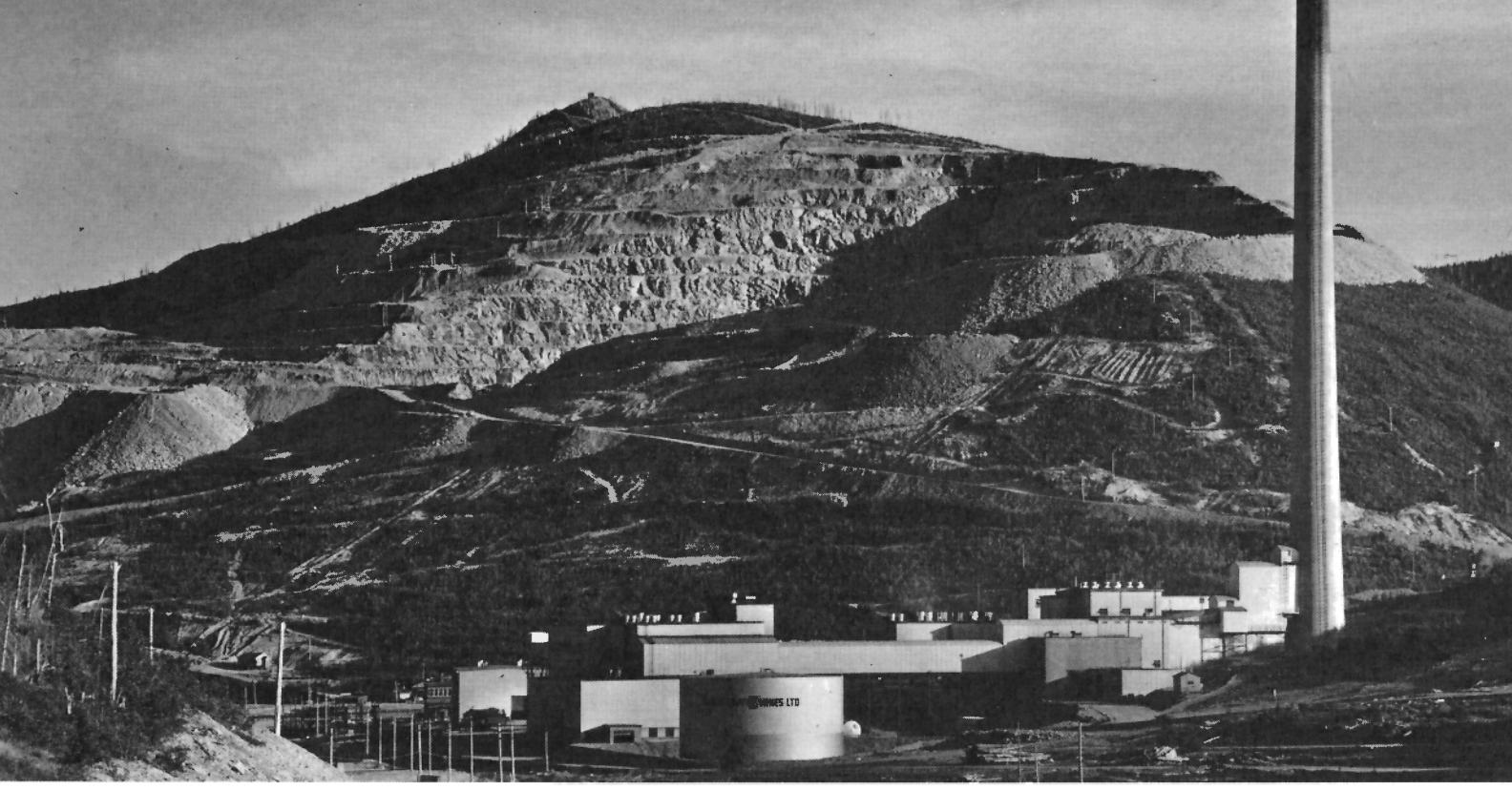
Mineral Provinces 309

Metalliferous Deposits 314

Industrial Mineral Deposits 336

Fuels 345

Selected References 362



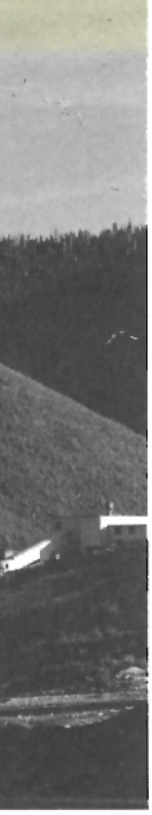
INTRODUCTION

The mineral industry of Southeastern Canada produces several metals, a variety of industrial minerals, structural materials, and fuels—oil, gas, and coal. Their value in 1966 was nearly $702 million, about 18 per cent of the value of the total mineral output of Canada. As shown in Table VII-1, the value of both structural materials and industrial minerals exceeded that of the metals produced. Asbestos was the most valuable economically with a production of 1,401,416 tons of fibre, 94 per cent of the Canadian total, and an estimated value of $152 million. It was followed in order of production value by cement, sand and gravel, stone, coal, zinc, copper, clay products, lead, salt, lime, gypsum, iron ore, silver, natural gas, oil, peat moss, niobium (columbium), fluorspar, gold, and barite; the values of these products ranged from $101 million for cement to $1.8 million for barite. Of the base metals zinc was the most valuable, with a production of 390,490,896 pounds, accounting for 20 per cent of the Canadian total and valued at $57.7 million. It was followed by copper, with 164,976,657 pounds at $40.7 million or 16 per cent of the Canadian total, and by lead, with 144,349,393 pounds at $21.5 million or 20 per cent of the Canadian total; this gives a total of 699,816,946 pounds of copper, lead, and zinc valued at $120 million for the year. Precious metals recovered from these ores contributed an additional $10 million from silver and gold, as well as $1.4 million from molybdenum, cadmium, and bismuth.

The principal mineral deposits are grouped into three main classes—metalliferous deposits, industrial mineral deposits, and fuels. In view of the complex geological history of the metalliferous deposits and of the uncertainties regarding their age, origin, and host rocks, the deposits have been grouped according to the dominant metals and minerals, thus avoiding subjective methods of classification, and enabling comparisons to be made and conclusions to be drawn regarding areas favourable for mineralization.

There is a marked association of mineral deposits with igneous rocks illustrated by the clustering of more than five hundred deposits in centres of igneous activity such as the serpentine belt of the Eastern Townships, the Notre Dame Mountains of Gaspé Peninsula, the central mineral belts of New Brunswick and Newfoundland, and in the roof rocks of the Devonian granite in Nova Scotia (Fig. VII-1). More specific associations are asbestos, chromium, and nickel deposits with the ultrabasic and basic rocks (Fig. VII-4); and gold, molybdenum, antimony, bismuth, tin, and tungsten deposits with granitic rocks (Fig. VII-5).

[1] Authorship responsibilities for sections of this chapter are as follows: Introduction, Mineral Provinces, Metalliferous Deposits, and Non-bedded Industrial Minerals by E. R. Rose; Natural Gas, Oil, and Bedded Industrial Minerals by B. V. Sanford; Coal by P. A. Hacquebard. When acknowledging, reference should be made to individual author and page number.



VII

Economic Minerals of Southeastern Canada

E.R. Rose, B.V. Sanford, and P.A. Hacquebard[1]

Gaspé Copper Mines, Limited, Murdochville, Quebec.

Résumé of Important Mineral Developments

Occurrences of valuable minerals in the early settled parts of Canada were first established by Champlain's discoveries of iron, silver, and copper in 1604 in what is now Nova Scotia. Also, the first coal mined by Europeans in the new world came from Newcastle Creek in the Minto field of New Brunswick. Massachusetts colonial records indicate that coal was shipped to Boston from the Grand Lake coal basin as early as 1639. The coal seams of Cape Breton, subsequently so important to the Canadian economy, were noted in 1672 by Nicholas Denys, a Governor of Acadia, and mined at Port Morien in 1720 to fuel the fortress of Louisbourg, and later the Citadel of Halifax. To date, nearly half a billion tons of coal have been produced.

Early discovery of bog iron in the Eastern Townships of Quebec was followed in 1823 by discovery of placer gold in the Chaudière River valley, and by discoveries of copper and gold in the bedrock schists of the Eastern Townships in 1841. In the period 1843–48, W. E. Logan, Alexander Murray, and James Richardson of the newly formed Geological Survey of Canada investigated the Eastern Townships, and the association of copper, asbestos, and chromite with rocks of the serpentine belt was noted. A number of small copper mines in this belt reached the production stage, stimulated by demands created by the American Civil War in 1865.

In Newfoundland, in 1857, the La Manche Lead mine was opened, and the returns from it are said to have helped finance the laying of the first transatlantic cables. In 1864 the first of Newfoundland's great copper mines, the Union mine at Tilt Cove, was opened with the guidance of Alexander Murray, newly appointed Director of the Geological Survey of Newfoundland. Murray's recognition of the similarity of the rock formations in the two producing areas led to prospecting and the development of many other copper deposits in the Notre Dame Bay area.

The first shipment of asbestos from the serpentine belt of the Thetford area of the Eastern Townships was made in 1876, the prelude to Quebec's subsequent eminence in this field. Asbestos has since become by far the most valuable mineral product of the Appalachian Region. In Ontario, bog iron deposits were discovered at Simcoe (Norfolk county) in 1800, followed by the discovery of gypsum near the present town of Paris in 1822. Crude oil was first produced from seepages in Enniskillen township of Lambton county and refined for use in illumination. This led to the drilling of the first well in 1858 in Lambton county and ultimately the discovery of Devonian oil at Oil Springs and Petrolia in 1861 (Pl. VII-1) and at Bothwell in 1862. Salt was discovered at Goderich in 1866, as a direct result of the search for oil. Natural gas was discovered at opposite extremities of the Niagara Peninsula in 1889 at the Kingsville and Welland fields.

In Nova Scotia iron deposits were developed at Torbrook in 1825, at East River in 1828, and at Londonderry in 1849. Albertite was found in New Brunswick in 1851. In the 1860s the discovery of gold-bearing quartz veins

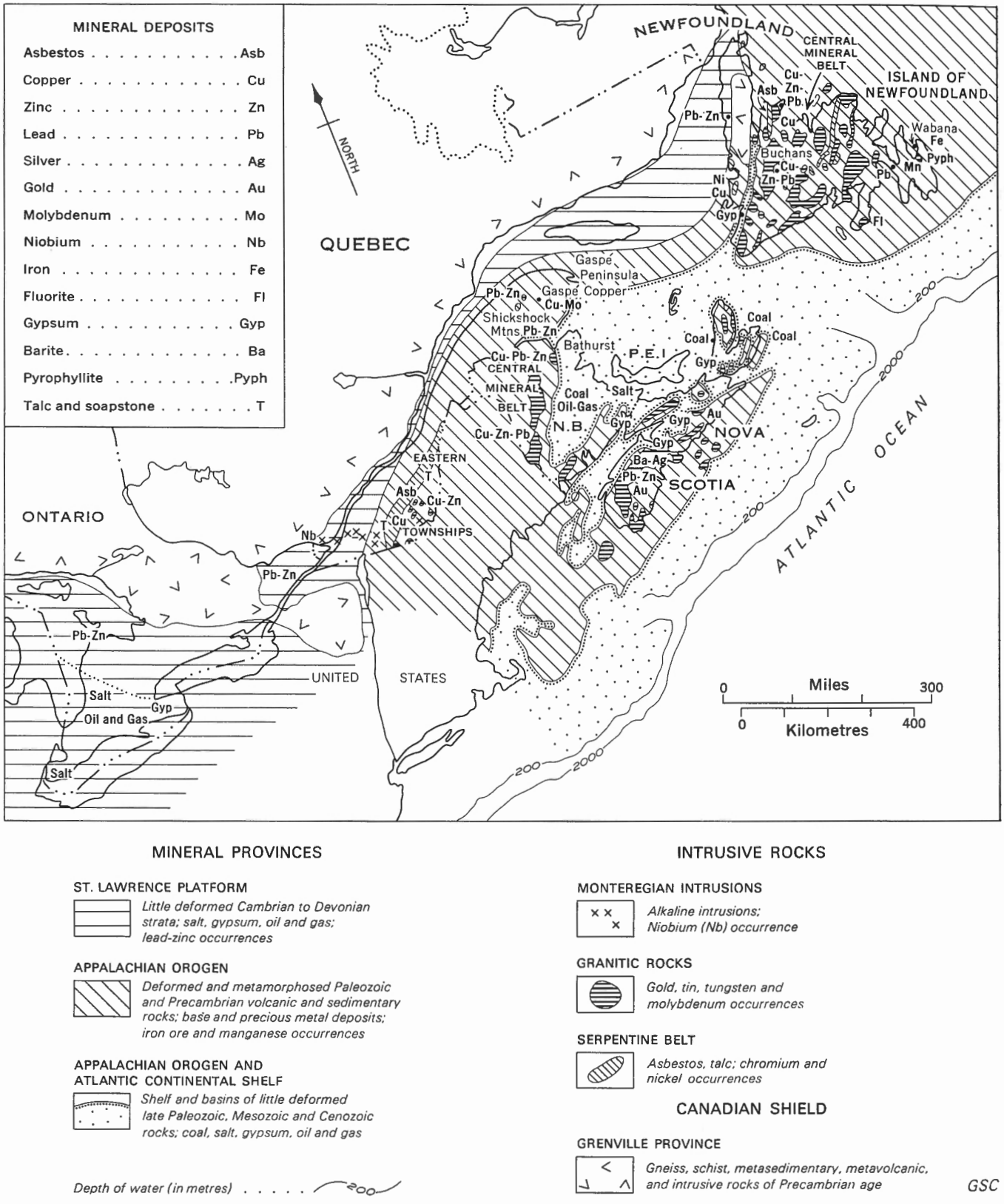


FIGURE VII-1. Mineral provinces of Southeastern Canada, showing principal mineral-producing areas and deposits (by E. R. Rose).

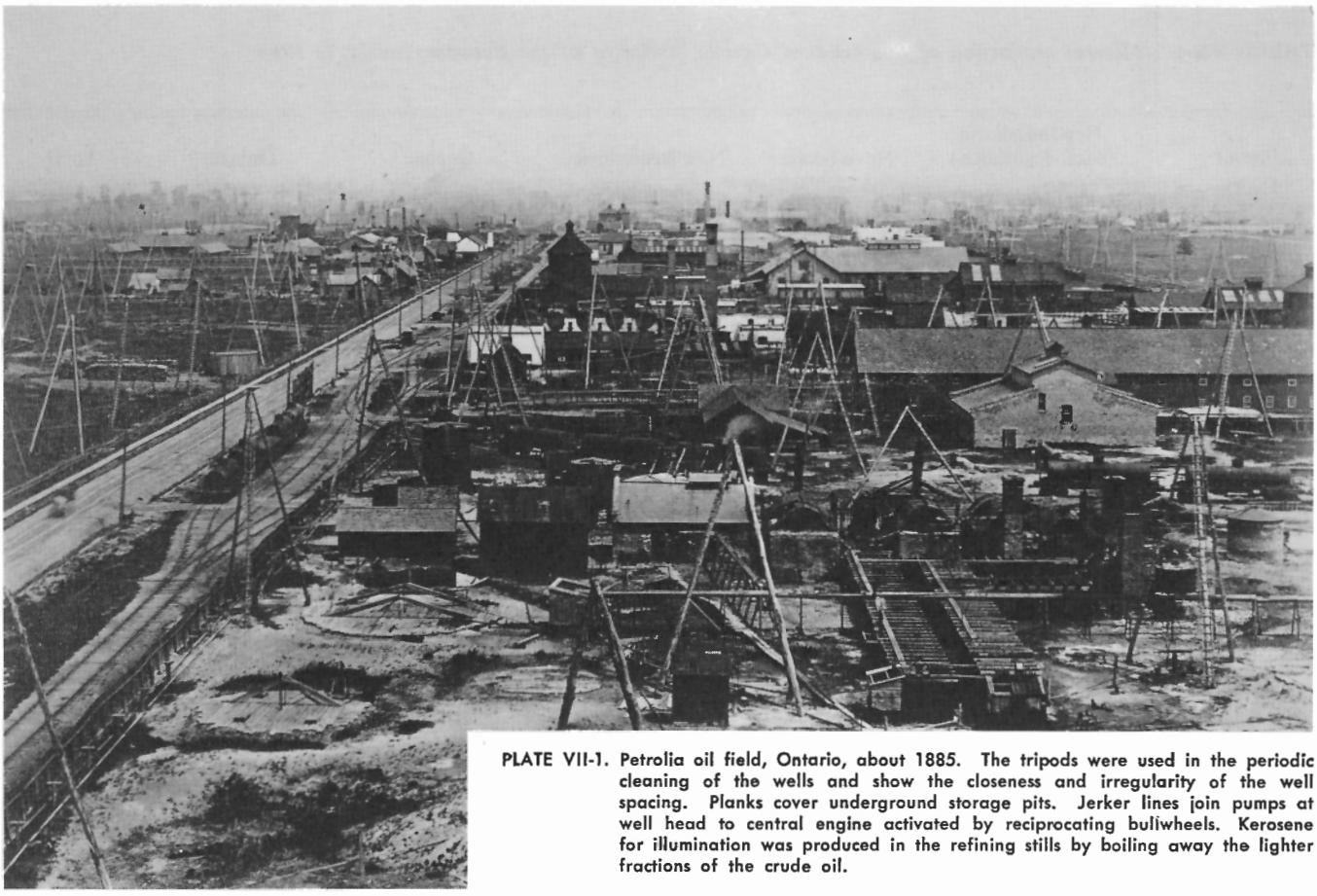


PLATE VII-1. Petrolia oil field, Ontario, about 1885. The tripods were used in the periodic cleaning of the wells and show the closeness and irregularity of the well spacing. Planks cover underground storage pits. Jerker lines join pumps at well head to central engine activated by reciprocating bullwheels. Kerosene for illumination was produced in the refining stills by boiling away the lighter fractions of the crude oil.

led to the development of an important gold mining industry in Nova Scotia. The nineteenth century closed with the discovery in 1892, and first production of iron ore at Wabana on Bell Island, Newfoundland, in 1895. During the next 70 years, until the mines were closed in 1966, the Wabana deposits yielded more than 70 million tons of iron ore, and they are by no means exhausted now. The present century of modern mineral development was ushered in with the establishment of the Dominion Steel and Coal Company steel mill at Sydney, Nova Scotia, in 1900.

Discoveries of this Century

The outstanding discoveries of the present century have been: the Buchans lead–zinc–copper deposit in central Newfoundland discovered in 1905 and brought into production in 1928; the Stony Creek gas and oil field in 1909 near Moncton, New Brunswick; the fluorite veins at St. Lawrence, Newfoundland, brought into production in 1933; the Walton barite deposit and associated lead–zinc–silver ore in Nova Scotia in 1940; the copper deposits in the Gaspé Peninsula discovered in 1921 and brought into production in 1955; natural gas in the Silurian at Tilbury in 1906, which marked the beginning of natural gas production and marketing on a major scale; oil in the Ordovician in 1917, broadening the exploration targets; drilling offshore in 1943, which resulted in the extension of several gas fields into Lake Erie; oil and gas in the Cambrian in 1958; the Advocate asbestos deposits near Baie Verte, Newfoundland in 1963; and recently, the zinc–lead–copper deposits in the Bathurst area of New Brunswick that now are coming into full production with completion of the smelter complex at Belledune.

MINERAL PROVINCES

Southeastern Canada is divisible into three geological provinces, the Appalachian Orogen, St. Lawrence Platform, and Atlantic Continental Shelf. They form parallel, northeasterly trending belts along the eastern margin of the continent and are flanked by the Canadian Shield on the northwest (Fig. VII-1). Differences in the geology and metallogenic history of these provinces permit comparable separation of the region into three mineral

TABLE VII-1 | *Mineral production of Southeastern Canada, exclusive of the Canadian Shield, in 1966*

Mineral	Newfoundland (excl. Labrador)	Nova Scotia	New Brunswick	Quebec	Ontario	Total
METALS						
Bismuth				36,913 lb. $156,880*		$ 156,880
Cadmium			38,902 lb. $93,365	138,202 lb. $380,481*		473,846
Columbium (Niobium)				2,600,000 lb. $3,150,000		3,150,000
Copper	37,338,361 lb. $16,764,924	583,170 lb. $261,843	13,106,548 lb. $5,884,840	113,948,578 lb. $17,800,000*		40,711,607
Gold	24,912 oz. $939,182		1,768 oz. $66,654	24,411 oz. $861,552*		1,867,388
Iron ore	792,000 tons $9,085,000*					9,085,000
Lead	42,223,742 lb. $6,308,227	3,251,914 lb. $485,836	95,998,000 lb. $14,342,101	2,875,737 lb. $420,726*		21,556,890
Molybdenum				531,598 lb. $823,976*		823,976
Silver	1,070,943 oz. $1,498,249	574,505 oz. $803,732	3,025,094 oz. $4,232,107	1,020,446 oz. $1,659,902*		8,193,990
Zinc	68,876,000 lb. $10,400,276	1,090,320 lb. $164,636	292,398,635 lb. $44,152,194	28,125,941 lb. $3,000,000*		57,717,106
Total metals	$44,995,858	$1,716,047	$68,771,261	$28,253,517		$143,736,683
NON-METALLICS						
Asbestos	64,850 tons $10,300,000			1,336,566 tons $141,559,725		$151,859,725
Barite		195,000 tons $1,811,300				1,811,300
Fluorspar	$2,187,500					2,187,500
Grindstone			5 tons $1,500			1,500
Gypsum	459,685 tons $1,173,401	4,502,836 tons $8,140,651	108,207 tons $413,074		565,185 tons $1,581,010	11,308,136
Iron oxide				300 tons $14,000		14,000
Peat moss		3,000 tons $105,000	53,000 tons $1,438,000	110,000 tons $2,500,000		4,043,000
Salt		474,981 tons $4,724,993			3,782,191 tons $15,243,791	19,968,784
Soapstone, talc, pyrophyllite	40,548 tons $608,220			15,100 tons $191,100		799,320
Total non-metallics	$14,269,121	$14,781,944	$1,852,574	$144,264,825	$16,824,801	$191,993,265

FUELS						
Coal		3,854,534 tons $51,518,674	898,315 tons $7,892,427			$59,411,101
Natural gas			97,601 Mcf $92,600	3,000 Mcf $450	15,537,157 Mcf $5,896,662	5,989,712
Petroleum			6,853 bbl. $8,700		1,323,781 bbl. $4,500,000	4,508,700
Total fuels		$51,518,674	$7,993,727	$450	$10,396,662	$69,909,513
STRUCTURAL MATERIALS						
Clay products	$194,102	$1,592,466	$588,625	$5,280,000*	$25,365,409	$33,020,602
Cement	55,441 tons $1,219,700	210,000 tons $3,549,000	260,000 tons $4,290,000	2,381,288 tons $39,488,812*	3,242,591 tons $52,680,630	101,228,142
Lime			3,723 tons $105,650	176,171 tons $1,884,153*	1,077,550 tons $12,210,576	14,200,379
Sand and gravel	3,647,355 tons $3,502,363	7,896,332 tons $10,573,115	6,054,873 tons $3,147,253	39,688,898 tons $18,488,984*	53,534,066 tons $38,612,235	74,323,950
Stone	86,000 tons $108,400	519,005 tons $1,719,465	2,888,579 tons $2,896,080	38,860,131 tons $44,117,616*	21,011,319 tons $24,709,496	73,551,057
Total structural materials	$5,024,565	$17,434,046	$11,027,608	$109,259,565	$153,578,346	$296,324,130
Grand total	$64,289,544	$85,450,711	$89,645,170	$281,778,357	$180,799,809	$701,963,591

Compiled from data supplied mainly by Mineral Resources Division, Dominion Bureau of Statistics, and provincial agencies
* Value calculated at current market price

provinces. The westernmost St. Lawrence Mineral Province includes the relatively undeformed Paleozoic sedimentary cover on the crystalline rocks of the Canadian Shield, and five of the eight known Cretaceous (?) alkaline Monteregian Intrusions. The Appalachian Mineral Province, or middle belt, includes the deformed Paleozoic and Precambrian rocks of the Appalachian Mountains, three of the Monteregian Intrusions, and several more or less undeformed sedimentary basins of Late Paleozoic and Triassic rocks. The rocks of this belt probably extend beneath the Gulf of St. Lawrence and the Atlantic Continental Shelf where younger Mesozoic and Cenozoic sediments occur as a cover on the folded rocks.

St. Lawrence Mineral Province

The St. Lawrence Mineral Province is characterized by unmetamorphosed sedimentary rocks—sandstone, shale, carbonate, and evaporites of Paleozoic age—that were deposited in the epeiric seas covering the Canadian Shield. Here oil and gas accumulated under favourable conditions, and deposits of limestone, salt, and gypsum were precipitated. The strata overlying the Precambrian crystalline basement are relatively undisturbed except along the Logan thrust zone and associated thrusts and tear faults, and near post-orogenic normal faults bordering horst and graben structures formed after the Appalachian orogenies. Igneous activity was mainly restricted to the emplacement of the Cretaceous–Triassic Monteregian alkaline intrusive complexes during which niobium, rare earth, and iron–titanium–vanadium occurrences were formed. Five of these intrusive plugs forming Mount Royal, St. Bruno, Rougemont, and Johnson intrude undeformed Paleozoic rocks in the St. Lawrence Province, and the related Oka stock, from which niobium is now being produced, pierces an inlier of the Precambrian basement complex nearby.

Appalachian Mineral Province

The Appalachian Mineral Province has a complex tectonic and geological history (Fig. VII-2) involving several orogenic disturbances; several episodes of igneous activity including the intrusion of ultrabasic, basic, intermediate, and acidic rocks of diverse ages on a large scale; many epochs of multiform volcanism ranging through the Precambrian, Cambrian, Ordovician, Silurian, and Devonian periods; as well as a host of metalliferous mineral deposits, a variety of industrial minerals, important coal deposits, and a minor amount of oil and gas. This region is particularly noted for copper–zinc–lead deposits (Fig. VII-3) and asbestos (Fig. VII-4), as well as iron, gypsum, salt, barite, fluorite, and coal. Many other metallic elements—antimony, arsenic, bismuth, cadmium, cobalt, chromium, manganese, molybdenum, nickel, tin, tungsten, uranium, vanadium and others—are known in numerous scattered occurrences, some of which have been mined from time to time (Figs. VII-4, VII-5). Bis-

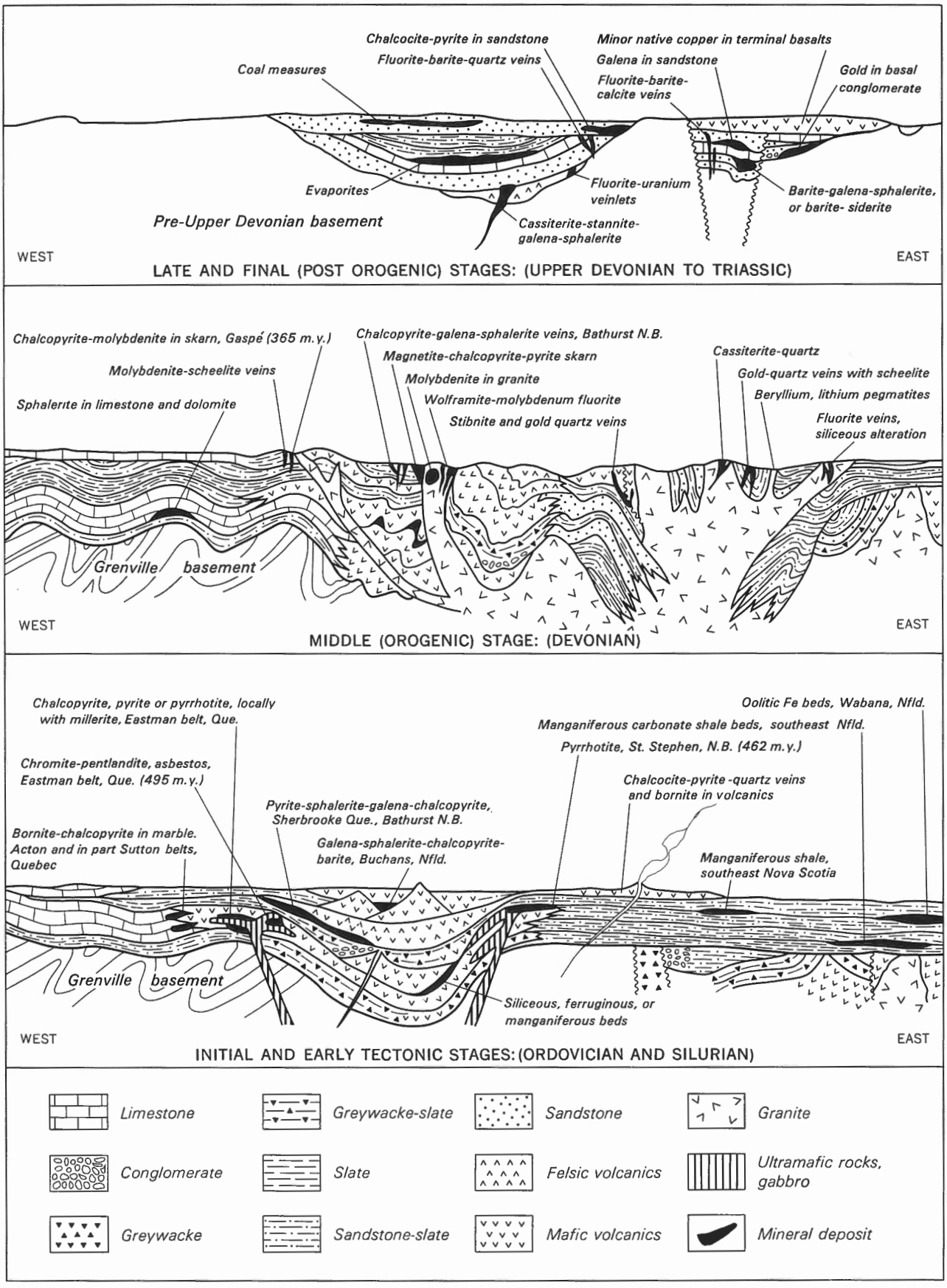


FIGURE VII-2. An idealized metallogenic model of the Appalachian Orogen depicting some known types of mineral occurrences in relation to their tectonic setting and the facies and approximate ages of their wall-rocks (by W. D. McCartney).

muth, cadmium, gold, molybdenum, selenium, silver, and tellurium are also now being recovered as byproducts of some of the mines. Within this province inliers of Precambrian crystalline rocks form the core of the Great Northern Peninsula of Newfoundland, parts of the Long Range Mountains of western Newfoundland, and the highlands of Cape Breton and southeastern New Brunswick. In eastern Newfoundland folded Proterozoic sedimentary and igneous rocks form a platform that probably extends under the Atlantic Continental Shelf. Extrusive and intrusive igneous rocks constitute a major part of the central mineral belts of Newfoundland and New Brunswick, as well as the Eastern Townships and Gaspé Peninsula of Quebec. There is a general association of ore deposits with these belts of igneous rocks.

In parts of the Appalachian Mineral Province Paleozoic and Precambrian strata have been deformed in four orogenies—the Ordovician Taconian Orogeny, a disturbance in the Silurian, the Devonian Acadian Orogeny, and the Appalachian Orogeny or Maritime Disturbance closing the era. The Taconian Orogeny is represented by local unconformities, thrust faults, and the emplacement of large plutons of ultrabasic rock in the Eastern Townships and central Newfoundland that have yielded asbestos, talc, soapstone, chromite, titaniferous magnetite, pyrrhotite, and chalcopyrite. The climactic Acadian Orogeny is represented by widespread folding and faulting, intrusion of granitic rocks, and metamorphism, particularly in southern Nova Scotia, central southern Newfoundland, central New Brunswick, and the Eastern Townships. Mineral deposits produced were gold in quartz veins in the metasedimentary rocks of Nova Scotia, and stibnite, molybdenite, cassiterite, stannite, wolframite, scheelite, bismuthinite, and fluorite in the Eastern Townships of Quebec. The emplacement of the granitic rocks has involved partial assimilation, metamorphism, recrystallization, and re-mobilization of pre-existing wall-rocks and contained mineral deposits in places; elsewhere it has merely resulted in a slight remetamorphism and mineralization. The effects of the Appalachian Orogeny are folding, thrusting, and metamorphism in the United States; in Canada these effects occur mainly in southwestern New Brunswick. Elsewhere, in western Newfoundland and northeastern Nova Scotia, wrench faulting and folding of the Carboniferous coal-bearing beds occurred, partly contemporaneously with their deposition.

Permian red beds unconformably overlie the Carboniferous and show little sign of deformation. Man-

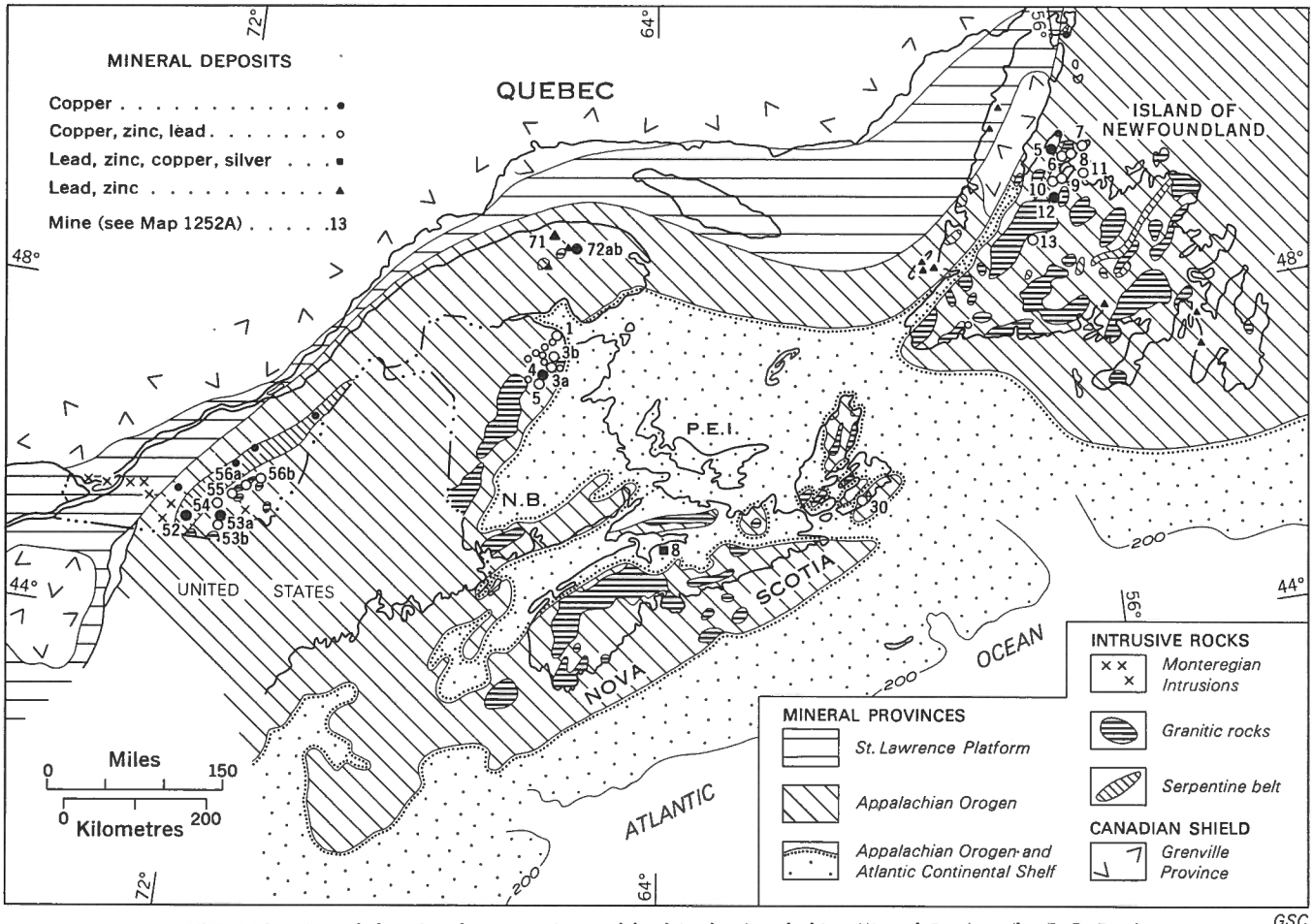


FIGURE VII-3. Principal deposits of copper, zinc, and lead in the Appalachian Mineral Province (by E. R. Rose).

GSC

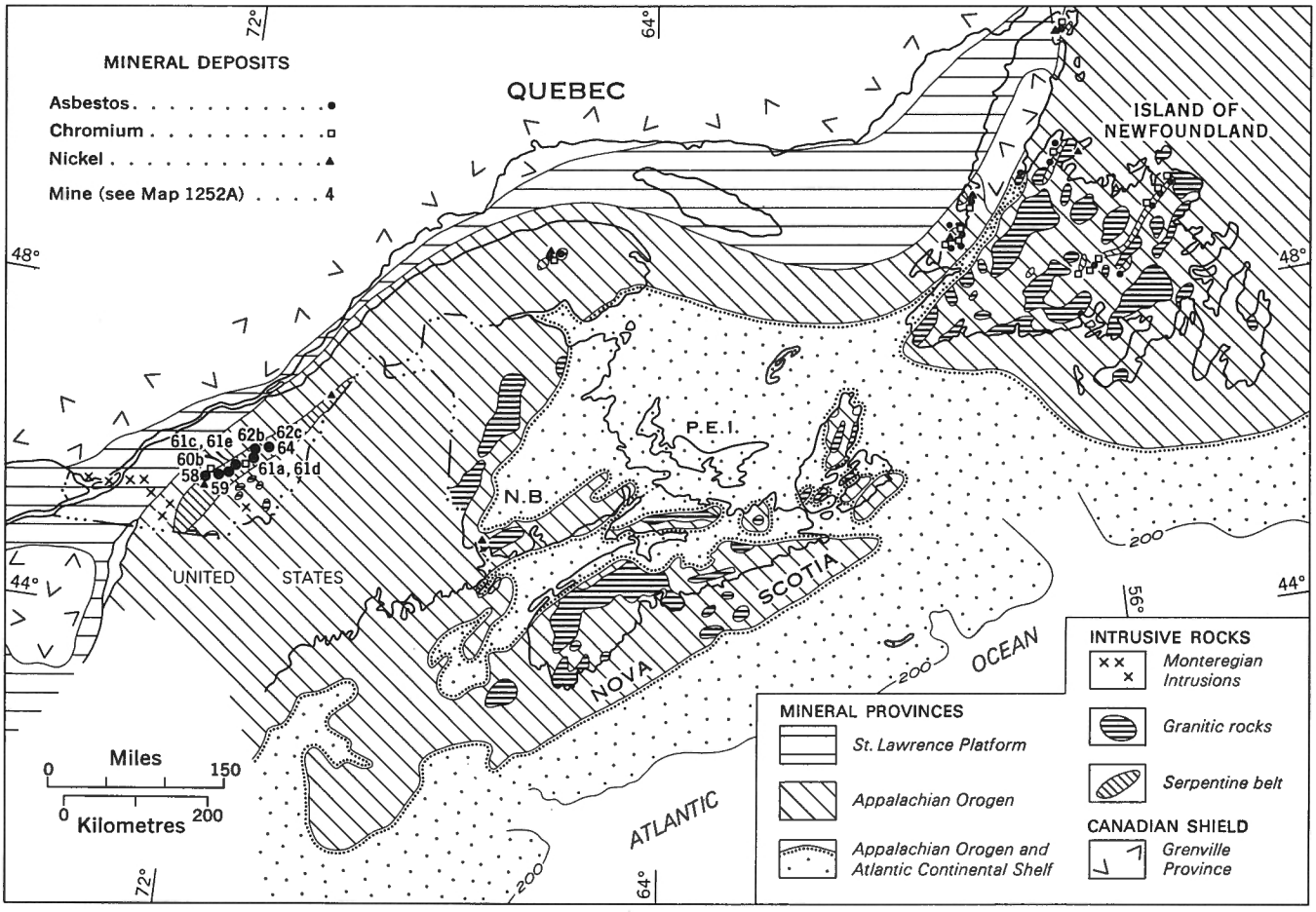


FIGURE VII-4. Principal deposits of asbestos, chromium, and nickel in the Appalachian Mineral Province showing their spatial relationship to ultrabasic and basic rocks (by E. R. Rose).

ganese oxide occurs in the weathered rocks at the unconformity. Igneous activity occurred again in the Triassic. The trap and related rocks produced carry titaniferous magnetite, native copper, amethyst, and zeolites in Nova Scotia and New Brunswick. Three of the Monteregian Hills, Mounts Yamaska, Shefford, and Brome, as well as the nordmarkite ring dykes of Mount Megantic, are formed by intrusion of ovate bodies of alkaline magma into the folded rocks of the Appalachian Province in the southern part of the Eastern Townships. These intrusions probably represent the most recent igneous activity in southeastern Canada.

Atlantic Continental Shelf

The Atlantic Continental Shelf extends seaward for more than 300 miles east of Nova Scotia and Newfoundland forming a belt on the eastern margin of the continent approximately equal in width to the Appalachian Orogen. Although little is known yet about the geology and mineral deposits of this vast submarine area, which undoubtedly has had a varied structural and geological history, it is probably partly underlain by thick deposits of sedimentary rocks ranging from Upper Devonian to Recent. Currently the Atlantic Continental Shelf and the Gulf of St. Lawrence are being investigated by sea- and air-borne geophysical surveys and by drilling. A sedimentary basin, believed to be more than 15,000 feet thick, underlies the Gulf of St. Lawrence, and it is anticipated that equally thick basins of Mesozoic and Cenozoic rocks occupy parts of the continental shelf, possibly including favourable reservoirs for recoverable oil and gas.

METALLIFEROUS DEPOSITS

Metalliferous mineral deposits of the Appalachian Mineral Province contain a variety of elements including the base metals (nickel, copper, zinc, lead, iron, and manganese), the precious metals (gold and silver), and a number of minor metals (antimony, bismuth, cadmium, molybdenum, tin, tungsten, chromium, titanium, and vanadium). The deposits are formed mainly of sulphide and oxide ore minerals, with the sulpharsenides, arsenides, tellurides, carbonates, and native metals being only locally abundant. Many deposits contain a variety of elements,

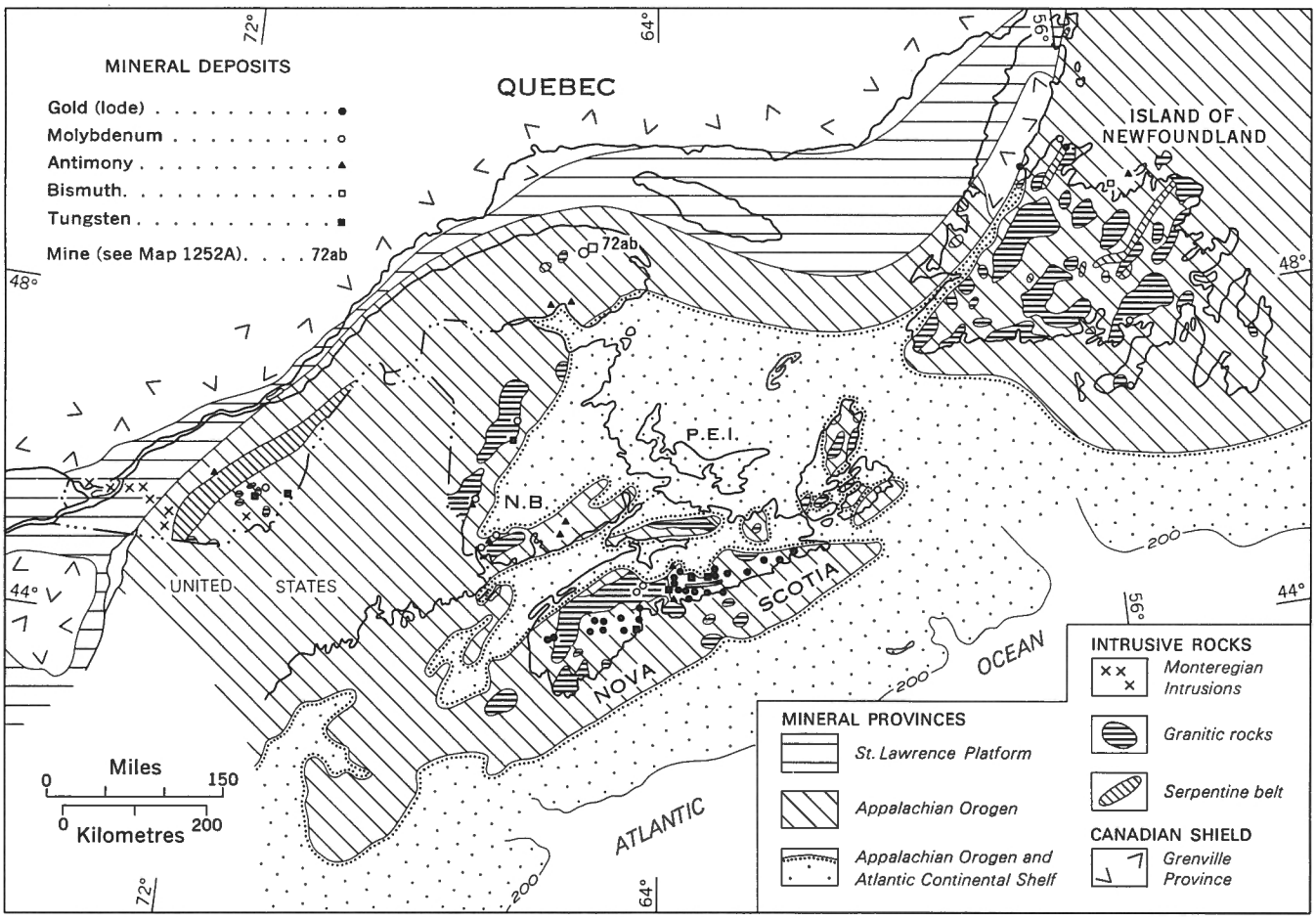


FIGURE VII-5. Principal deposits of gold, molybdenum, antimony, bismuth, tin and tungsten in the Appalachian Mineral Province showing their spatial relationships to granitic rocks (by E. R. Rose).

and commercial recovery of two or more of the contained metals or minerals as byproducts of the main mining operation is now common practice. Only a few deposits are large and rich enough to be mined. Although zinc was the most valuable metal produced in 1966, few mines produce zinc alone. In many, copper is commonly associated with zinc, lead, silver, gold, antimony, arsenic, bismuth, cadmium, and sulphur, and less commonly with nickel or molybdenum, any of which may be recovered. Gangue minerals include pyrite, pyrrhotite, quartz, carbonates, chlorite, and barite. Sulphide ore minerals are mainly chalcopyrite, sphalerite, and galena. They occur in varying proportions to form massive or disseminated sulphide deposits in a variety of host rocks and structures.

Pyrite is the dominant sulphide mineral in most of the base-metal deposits although pyrrhotite predominates in some and nickeliferous pyrrhotite and pentlandite occur in a few. Many of the nickeliferous pyrrhotite–pentlandite deposits are within or closely associated with mafic or ultramafic intrusions and are undoubtedly genetically related to them, as are some of the pyrrhotite–chalcopyrite occurrences. Many of the latter merge into pyrite–chalcopyrite occurrences with sphalerite and galena, forming base-metal sulphide deposits that are either replacements of altered basic to intermediate volcanic rocks, with or without intercalated metasediments, or of the interlayered metasediments. This general association with volcanic rocks suggests a common origin for the contained copper–zinc–lead deposits, many of which are in pre-Devonian rocks, probably Ordovician. According to some investigators these deposits may have formed with the volcanic and sedimentary rocks at or near surface, but according to others they appear to have been introduced at depth. In a few mines, such as Gaspé Copper and Nigadoo, base-metal deposits occur within granitic intrusions and their aureoles of skarn, and there seems little doubt of a genetic association with the intrusions. On the other hand, numerous lead–zinc occurrences are known in undeformed carbonate strata and veins seemingly unrelated to igneous activity.

Antimony

In the Eastern Townships of Quebec, at the *South Ham* mine, native antimony, stibnite (Sb_2S_3), kermesite (Sb_2S_2O), and valentinite (Sb_2O_3) occur as flakes and powder along cleavage planes in sheared arkosic sediments

and schist, and as a vein in a shear zone near or along the contact with an intrusion of serpentinized pyroxenite. The occurrence, about 18 miles east of Asbestos, was first described by Sir William Logan in 1863. It has been traced along strike for half a mile and explored by a shaft 100 feet deep and an adit 380 feet long. About 180 tons of concentrates have been produced. In Gaspé Peninsula stibnite occurs in quartz veins in Ordovician quartzite and conglomerate in New Richmond township and in small pockets in volcanic rocks and limestone in Carleton township.

Stibnite and native antimony occur in quartz veins cutting Ordovician(?) slate and quartzite near masses of diabase and Devonian granite at *Lake George,* about 25 miles west of Fredericton, New Brunswick. At *West Gore,* Nova Scotia, stibnite, kermesite, and native antimony occur with gold, arsenopyrite, and pyrite in small ore shoots in quartz–calcite veins in fissures cutting black slates similar to those of the Ordovician Halifax Formation. One vein has been traced for 2,700 feet and was worked sporadically from 1844 until 1942. Although the nearest exposure of intrusive Devonian granite is several miles distant, specimens from the mine show dykelets of quartz and feldspar cutting the slate (Stevenson, 1958).

In Newfoundland, at *Moretons Harbour* on New World Island, a quartz vein carries arsenopyrite, pyrite, sphalerite, chalcopyrite, galena, stibnite, and calcite. It averages 4 inches in width and has been traced for about half a mile along the sheared footwall contact with a thin rhyolite dyke that is intrusive into chloritized andesite of the Ordovician Moreton volcanics. An assay of a channel sample taken across one foot of vein material and 4 inches of footwall showed 5.39 per cent antimony, 4.39 per cent sulphur, 48.6 per cent silica, and 0.55 ounce of both gold and silver per ton. The deposit was mined sporadically around 1890, 1906, and 1915 (Baird and Snelgrove, 1953).

Bismuth

On *Pond Island* in the Bay of Exploits, Newfoundland, three quartz veins, each several inches wide, carry pyrite, bismuthinite (Bi_2S_3), chalcopyrite, sphalerite, tetrahedrite, and calcite; they follow joint planes in Silurian(?) granodiorite near a stock of rhyolite porphyry. A selected sample yielded 1.5 per cent bismuth, 13.92 per cent antimony, 0.5 per cent zinc, 3.65 per cent copper, 12.6 per cent sulphur, and 5 ounces per ton silver.

In 1966, at Gaspé Copper Mines Limited, Quebec, 36,913 pounds of bismuth valued at $156,880 was recovered from smelter flue dust. The native bismuth is a rare constituent of the complex copper ores mined at Copper Mountain and Needle Mountain.

Cadmium

In 1966, 38,902 pounds of cadmium valued at $93,365 was recovered from zinc concentrates produced from zinc–lead–copper ores in the *Bathurst* area of New Brunswick. The cadmium occurs in solid solution in sphalerite, and probably also as greenockite (CdS).

Nickel

Nickeliferous Pyrrhotite–Pentlandite

Nickeliferous pyrrhotite is a common constituent of relatively unaltered basic and ultrabasic igneous rocks of Appalachian Region, as at Mount Albert in Gaspé; St. Stephen, New Brunswick; and in Rope Cove Canyon, Gregory River, Hare Bay, and Chapel Cove all in Newfoundland. Pyrrhotite-rich sulphide disseminations, segregations, and veins are associated with basic and ultrabasic intrusions. Nickel occurs as strong traces, and in some, pentlandite may be intergrown with nickeliferous pyrrhotite and chalcopyrite. Not all the pyrrhotite is nickeliferous, particularly that in deposits within highly altered rocks, such as those at Gullbridge mine in Newfoundland and Huntingdon mine in the Eastern Townships. On Sugarloaf Island in Bay of Exploits, Newfoundland, however, nickeliferous pyrrhotite, pyrite, magnetite, and chalcopyrite form part of a recrystallized carbonate lens that is intercalated with altered basic Ordovician volcanics and shale.

South of *Bay of Islands,* Newfoundland, a zone of disseminated nickeliferous–pyrrhotite–chalcopyrite 10 feet wide was discovered in 1936 near the base of the Ordovician(?) Bay of Islands Complex (Cooper, 1936). The sulphides occur interstitially to silicate minerals in a zone of interbanded troctolitic anorthosite, serpentine, and diallagic pegmatite. Pyrrhotite, pentlandite, and chalcopyrite, with secondary marcasite and violarite(?), average about 2 per cent of the zone, and representative samples show as much as 1 per cent nickel and 0.32 per cent copper. The mineralized zone was observed at two places a quarter mile apart. In the *Hare Bay* area of northern Newfoundland, dissemination of nickeliferous pyrrhotite in hornblendite and in peridotite near hornblendite intrusions is reported. Pyrrhotite, pentlandite, chalcopyrite, magnetite, and ilmenite, along with biotite and chlorite, occur interstitially to silicates. Pentlandite grains are veined by fine-grained pyrrhotite. At Goose Cove, also on Hare Bay, a copper mine was opened to a depth of 100 feet in 1908 on small sulphide deposits in Goose Cove schist of Ordovician age. Pyrite, pyrrhotite, chalcopyrite, and small amounts of sphalerite, magnetite, and copper–nickel–iron arsenides occur in a gangue of hornblende, epidote, calcite, and quartz as irregular lenticular bodies along the crests of drag-folds in the schists. The ore ranged from 2 to 12 per cent copper.

Nickeliferous Pyrite–Millerite

About 400 tons of high-grade nickel ore was produced prior to 1876 from lenses within talc–carbonate rocks associated with serpentinized peridotite in the Tilt Cove copper mine. The deposit is described in the section on copper–zinc–lead. Minor occurrences of a similar nature have been reported from time to time in talc–carbonate rocks elsewhere in the Appalachian serpentinite belts.

The *Eastern Metals* deposit was found in 1949 in Rolette township, Quebec, in carbonate–talc schist along the contact of serpentinite and slate, both of which are intruded by granodiorite dykes. A small tabular body of massive pyrite averaging 0.91 per cent nickel occurs within a larger mineralized zone that averages 0.15 per cent nickel and 1.52 per cent copper. Small amounts of millerite, gersdorffite, siegenite, pyrrhotite, and pentlandite are present. Pentlandite veinlets and patches occur in the serpentinite nearby. The ore minerals within the larger zone are mainly fine-grained pyrite, chalcopyrite, chalcocite, bornite, sphalerite, galena, and rare arsenopyrite. A potential orebody of about a million tons of low-grade material has been outlined in which the nickel and zinc minerals are found mainly in irregular stringers in brecciated masses of fine-grained pyrite.

Copper

Native Copper

Numerous occurrences of native copper are known in igneous rocks of various ages in the Appalachian Region, but none has been of sufficient importance to mine. The native copper is found as disseminations, thin sheets, amygdules, and calcite veins. It has been noted in association with cuprite and minor amounts of chalcopyrite, chalcocite, melaconite, tenorite, and malachite in serpentinized Ordovician(?) dunite.

At the *Gregory River* copper lodes in Newfoundland (Smith, 1958) copper minerals occur as: pyrite–chalcopyrite–arsenopyrite–sphalerite–quartz veins; chalcopyrite–nickeliferous pyrrhotite veins and disseminations associated with fracture zones in basic intrusive rocks of the Ordovician Bay of Islands Igneous Complex or their intrusive contacts with basic volcanic rocks; native copper in calcite veins in the basic volcanic rocks of the Ordovician Humber Arm Group. In the Mount St. Gregory area south of Bonne Bay, the occurrences, which were discovered in 1921, were drilled in 1953 and the known limits of the Court A zones extended for 1,000 feet along strike. The sulphide minerals, pyrite, chalcopyrite, and arsenopyrite, occur in quartz veins and disseminated through the metavolcanic host rocks that form the roof of the intrusive gabbro, and along the Court A tear fault.

Pyrrhotite–Chalcopyrite

Pyrrhotite, rather than pyrite, is an important constituent of a few of the chalcopyrite deposits such as the Lake Memphremagog, Huntingdon, and Ives mines in the serpentine belt of the Eastern Townships; and the Sugarloaf Island, Gullbridge, and Hodder deposits in Newfoundland. The predominance of pyrrhotite over pyrite in these deposits suggests an environment of higher temperatures and pressures, either within or near an igneous source. In this regard they are similar to the Ducktown-type copper deposits of the southern Appalachian Mountains. The occurrences of pyrrhotite are commonly nickeliferous, and in places may also carry millerite, as at Orford, Quebec, and Tilt Cove, Newfoundland; or pentlandite, as at St. Stephen, New Brunswick. Many are low in nickel, however, and do not show the customary relationship with a basic igneous source.

The *Huntingdon mine* (52)[1], near Lake Memphremagog, Quebec, has been worked to a depth of 1,000 feet. Chalcopyrite and pyrrhotite occur as four lenticular ore zones flanking serpentine and steatite bands within sheared, silicified, mafic volcanic rocks. The largest ore zone is 100 feet wide, 300 feet long, and 1,300 feet deep. The mine has produced more than a million tons of ore. The deposit at nearby Ives mine is similar.

Gaspé Copper Mines, Quebec (72). Pyrrhotite and chalcopyrite are important constituents of the skarn ore at *Needle Mountain* near Murdochville, in Gaspé Peninsula (Fig. VII-6). Since the mine opened in 1955, 25 million tons of copper ore has been produced, about half the original reserves. Molybdenite, silver, bismuth, selenium, and gold are also recovered. The ore occurs along several faults within a skarn zone in altered calcareous shales and limestones of the Lower Devonian Grand Grève and underlying Cape Bon Ami Formations, and as a stockwork of fracture fillings (Ford, 1959). Ore minerals comprise chalcopyrite, molybdenite, and native bismuth. Scheelite, sphalerite, and pyrite are rare and occur partly as small exsolved crystals in the chalcopyrite. Late quartz veins carrying bornite, chalcopyrite, chalcocite, tennantite, and fluorite cut the normal ore. Wallrocks are well-bedded argillaceous limestone or calcareous shales, and siliceous limestones that have been altered respectively to diopside–andradite, porcellanite, and wollastonite–diopside skarn.

At *Copper Mountain* nearby, chalcopyrite and pyrite, with minor molybdenite, chalcocite, sphalerite, galena, and bornite, and secondary limonite, azurite, and malachite, are present in skarn beneath an oxidized capping and above a biotite granite intrusion within which minor veinlets of quartz locally carry chalcopyrite and molybdenite. This deposit came into production by open pit early in 1968. Quartz–feldspar porphyry dykes are abundant, the feldspars being highly altered to fine-grained

[1] Number is that assigned the deposit on Mineral Deposits of Canada, Map 1252A.

sericite and clay. Potassium–argon dates of 350 m.y. and 395 m.y. have been obtained from the granite and porphyry dykes, respectively, suggesting that they are Devonian, intruded during the Acadian Orogeny.

The pyrite–chalcopyrite–molybdenite deposit and occurrences in granite underlying Copper Mountain show marked similarity to the porphyry copper–molybdenum-type deposits of the western Cordillera as exemplified by Climax, Colorado and Bingham, the Utah deposits, as well as the Copper Mountain, Bethlehem, and Endako deposits in British Columbia. They are, however, of much younger geological age, but there can be little doubt that these large, low-grade skarn and disseminated deposits are all genetically related to intrusive granitic rocks. The Bingham district is also a large producer of lead, zinc, and silver from fissure veins and replacement lodes that lie in a zonal arrangement in metamorphosed Pennsylvanian limestone beds peripheral to the central core of the copper- and molybdenum-bearing Utah Copper stock. An interesting analogy is that quartz–calcite fissure veins carrying sphalerite and galena with minor chalcopyrite have been found in the Cascapedia River area within Devonian limestones and other rocks to the south and north peripheral to the Shickshock and Copper Mountain granite bodies (Fig. VII-7). Silver-bearing galena also occurs in fractures in Devonian rocks at Cross Point on the Bay of Chaleur.

The *Hodder* copper prospect was found half a mile northwest of Fleur de Lys, Newfoundland in 1913. The main ore mineral is chalcopyrite; it occurs in bands or veins parallel to the foliation, and as disseminations in two bands of impure silicated marble (skarn?) each about 5 feet thick. The marble lies within black calcareous gneiss between garnetiferous gneiss on the west and garnet-free gneiss on the east. The upper marble band is more mineralized than the lower and carries as much as 17 per cent copper with traces of gold and silver. Pyrrhotite is slightly more abundant than pyrite; magnetite, bornite, and marcasite are also present. Quartz gangue fills fractures in all of these minerals. According to Fuller (1941): "Mineralizing solutions penetrated the calcareous gneiss and apparently replaced the relatively pure marble lenses. The solutions were probably derived from a subjacent granitic magma and are believed to be related to those which deposited the minerals at the molybdenite and lead prospects". There is some similarity in type of mineralization between this occurrence and the Gaspé Copper Mines deposits which are indeed related to an underlying, probably Devonian, granitic intrusion.

The *Terra Nova* (5) deposit at Baie Verte, Newfoundland, lies within the zone of ultramafic rocks of the serpentine belt. It is a steeply dipping sulphide lens about 30 feet wide lying along the contact of serpentinized peridotite and Ordovician chlorite schist. The deposit is believed to have been worked to a depth of 300 feet and to have yielded ore which averaged 2.41 per cent copper, 37.23 per cent sulphur, and $2.50 per ton each of gold and silver. As at the Huntingdon mine in Quebec, pyrite

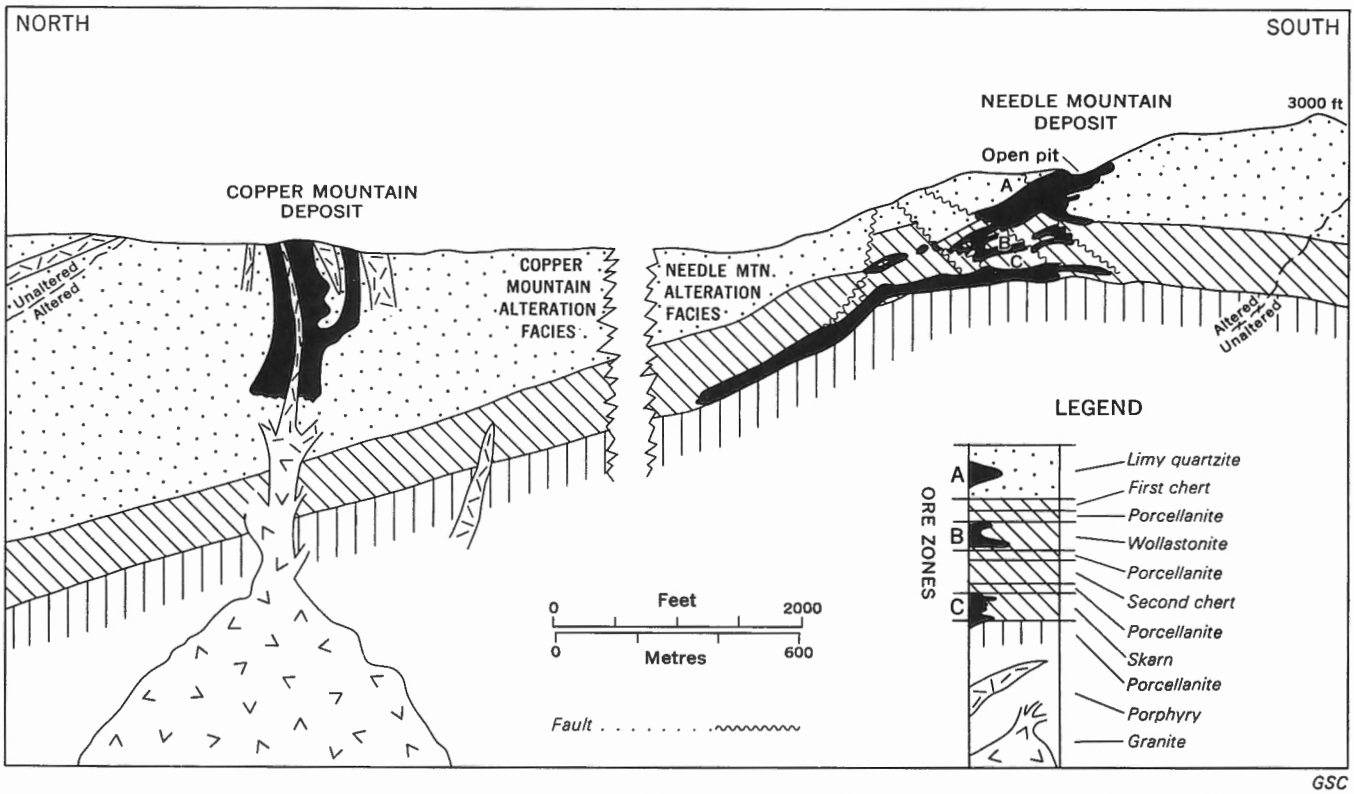


FIGURE VII-6. Generalized cross-section, Needle Mountain and Copper Mountain deposits, Quebec (courtesy of Gaspé Copper Mines Ltd.).

is the predominant sulphide mineral with considerable pyrrhotite and chalcopyrite, minor sphalerite, and rare specks of native copper.

Gullbridge mine (12). An occurrence of sulphide minerals on Gull Pond, near Badger in central Newfoundland, has been known since 1905. A four-compartment shaft has now been sunk to a depth of 1,050 feet, and the mine is being prepared for production. Pyrrhotite, pyrite, chalcopyrite, and magnetite, along with quartz, chlorite, sericite, biotite, cordierite, and tremolite–actinolite, occur in lenticular masses along sheared zones in andesitic or dacitic lavas, cherts, and siliceous tuffs of the Ordovician or Silurian Roberts Arm Group. The possible Silurian age of these rocks suggests that mineralization may not be confined to Ordovician rocks in Newfoundland. The group is flanked by intrusive diorite on the east and by the Silurian Springdale sandstone and Devonian granite on the west. It is intruded by swarms of dykes of acidic and basic composition. An aureole of pyrite surrounds the ore. Reserves of 4 million tons averaging more than one per cent copper are reported.

Copper–Zinc–Lead

Pyrite–Chalcopyrite–Sphalerite–Galena

Pyrite is the predominant sulphide mineral in most producing base-metal mines in the Appalachian Mineral Province. Together with sphalerite, galena, and chalcopyrite, with minor or rare tetrahedrite–tennantite, bornite, molybdenite, magnetite, pyrrhotite, arsenopyrite, and native bismuth, the pyrite forms irregular lenticular, massive and disseminated ore deposits. They characteristically occur in zones of sheared, folded, and brecciated rocks and most commonly in sequences of basic volcanics with intercalated siliceous sediments. Many occurrences are in eugeosynclinal Cambrian, Ordovician, and Silurian rocks that have been intruded and metamorphosed by igneous rock of acidic to basic composition, but few of the deposits show a sufficiently close association to any particular rock type to establish a clear-cut genetic relationship for this sub-type. Logan and Murray placed great emphasis on the association of these deposits with the serpentines and limestones, but most later geologists

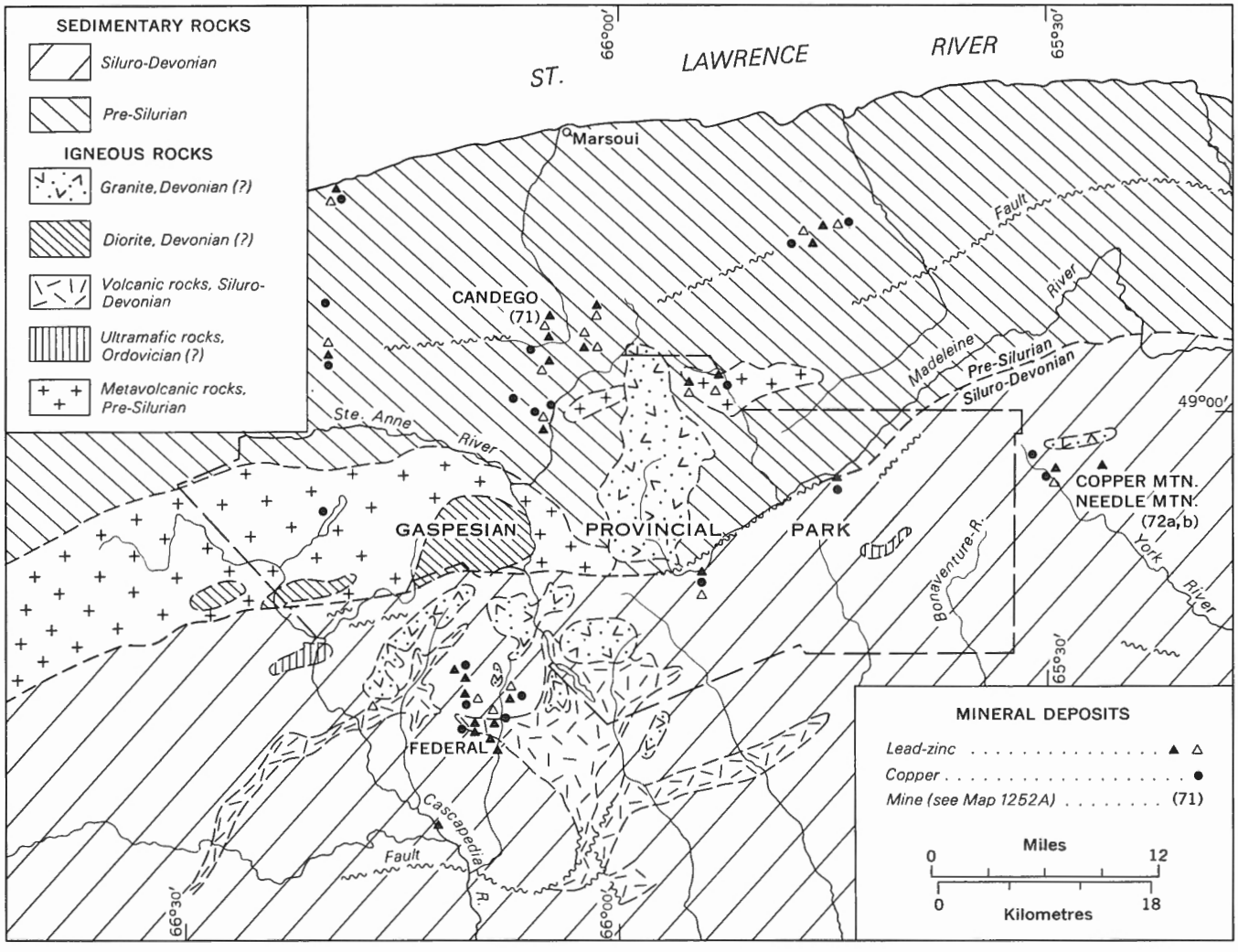


FIGURE VII-7. Relationship of copper and lead–zinc deposits to Devonian granite in Shickshock Mountains, Gaspé, Quebec (by E. R. Rose).

believed them to be replacement deposits probably associated with the emplacement of Devonian granitic intrusions. Many geologists now consider them to be stratiform deposits related to volcanic extrusion, gaseous exhalation, or to attendant sedimentation; others regard them as of metamorphic or multiple origin.

The *Harvey Hill* mine, Quebec, is a pyrite–chalcopyrite–bornite–chalcocite deposit in slaty schists of the Cambrian(?) Sutton Group. The ore occurs in transverse veins and lenticular bands that parallel the schistosity. An intrusion of pyroxenite, now altered to talc or soapstone, passes beneath the main workings and may have been the source rock. The deposit was mined sporadically for about 30 years; for 10 years it was mined along with the richer but smaller deposits of the Acton mine near Acton Vale. The latter orebodies occur in three lenses of brecciated, cherty, dolomitic limestone resting on deformed Ordovician black slate and shale. Intrusive greenstone (diabase) dykes and sills may have been source rocks for the mineralization.

Eustis–Capelton mines (53b). Two similar pyrite-chalcopyrite deposits in greenschists and green carbonate rock occur 7 miles south of Sherbrooke, Quebec. Pyritic lenses, 2 to 25 feet wide, arranged *en echelon* and carrying 0.05 to 1.25 per cent copper, were mined along strike for 250 feet and to a depth of about 1,000 feet. Below, a lens much richer in chalcopyrite was mined to a depth of 4,800 to 5,200 feet. The lens varied in width from 6 to 40 feet and averaged 110 feet along strike. At the southwest end it assayed 1.5 per cent copper and 40 per cent iron, and at the northeast end, 16 per cent copper and 26 per cent iron. About 2.5 million tons of pyritic copper ore was produced, mainly from the Eustis mine.

The *Suffield* or *Ascot* (53c) mine and the nearby *Albert* mine (53a), a few miles northeast of Sherbrooke, Quebec, were operated intermittently from 1865 to 1958. Pyrite, chalcopyrite, and sphalerite, with a little galena and traces of silver and gold, were produced from two flat lenses in green chloritic tuff on the footwall of a band of Ordovician or Cambrian quartzite. The largest lens was 200 feet wide by 1,500 feet long. About a million tons of ore was produced.

The *Moulton Hill–Aldermac mine* (54) northeast of Sherbrooke, Quebec, was developed as a pyrite and copper producer. A new deposit was found in 1942, with the aid of geophysical surveys, along a shear zone a quarter mile south of the old mine at the contact of sheared rhyolite and sedimentary schists. It consisted of pyrite, sphalerite, galena, chalcopyrite, and tetrahedrite with a gangue of barite, quartz, and calcite. The mine was operated intermittently for several years, producing about 350,000 tons of pyritic copper–zinc–lead ore.

Weedon mine (55), a pyrite–chalcopyrite deposit discovered in 1908, 40 miles northeast of Sherbrooke, Quebec, was mined for 20 years to a depth of 1,150 feet. The ore, consisting of granular pyrite with chalcopyrite and minor pyrrhotite, sphalerite and galena, formed a lenticular deposit about 50 feet wide and 570 feet long, mainly along the contact of sericite (hanging-wall) and chlorite (footwall) schists. The ore follows folds in the schists, which are similar to those in the Eustis and Capelton mines, about 50 miles to the southwest. These schists are now believed to represent altered rhyolitic pyroclastic rocks, greywackes, and mafic volcanics. They are intruded by albite or bluish grey porphyries. An intrusion of Devonian granite forms the hanging-wall of the deposit in the lower levels of the mine, but its precise relationship to the ore deposit is not certain. The mine was an important pyrite and copper producer during the periods 1909–21 and 1951–59, and has produced more than a million tons of copper–zinc–lead ore, as well as 200,000 tons of pyrite for the manufacture of sulphuric acid.

At *Cupra* and *Solbec mines* (56a, b) two deposits of pyrite–chalcopyrite–sphalerite ore were found recently along the contact between Ordovician(?) sericite and chlorite schists northeast of Sherbrooke, Quebec. Production began at Solbec copper mines in 1962 and at Cupra mines in 1965. The combined mill feed in 1966 was half a million tons. Combined reserves are about 3 million tons, averaging 3.5 per cent copper, 3.5 per cent zinc, 0.5 per cent lead, and 1.3 ounces of silver and 0.02 ounce of gold per ton. The ore consists mainly of granular pyrite with interstitial chalcopyrite, sphalerite, bornite, galena, and minor tennantite. The Cupra deposit forms a tabular mass about 10 feet thick and more than 250 feet long that extends for more than 5,000 feet down dip. Bornite is abundant in the upper levels but has not been found in the Solbec deposit. A very thin siliceous iron-formation is present in parts of the Solbec ore zone.

Bathurst–Newcastle area, New Brunswick (1 to 5). The *Brunswick No. 6* (3a) and *No. 12* (3b) sulphide deposits were discovered in 1952 by diamond drilling on magnetic anomalies near the Austin Brook magnetite iron-formations and an associated pyritic sulphide body. The association between the iron-formation and sulphide deposits subsequently led to intensive exploration of magnetic anomalies and the use of other geophysical and geochemical methods of prospecting by which twenty-five or more sulphide deposits were discovered in the Bathurst–Newcastle area (Fig. VII-8). Some fifteen of these deposits, listed in Table VII-2, held reserves in excess of 120 million tons of copper–lead–zinc ore (Davies and Smith, 1966).

The sulphide deposits occur mainly in sedimentary rocks, iron-formation, and schists of Ordovician and Silurian age that are highly deformed and metamorphosed, commonly interlayered with volcanic rocks, and intruded by igneous rocks ranging from gabbro to granite. The deposits are usually overlain by thick sooty gossans with zones of secondary enrichment (Boyle and Davies, 1964). The deposits are fine-grained assemblages of: (1) pyrite with sphalerite, galena, minor chalcopyrite and pyrrhotite, rare arsenopyrite and tetrahedrite–tennantite, and a gangue

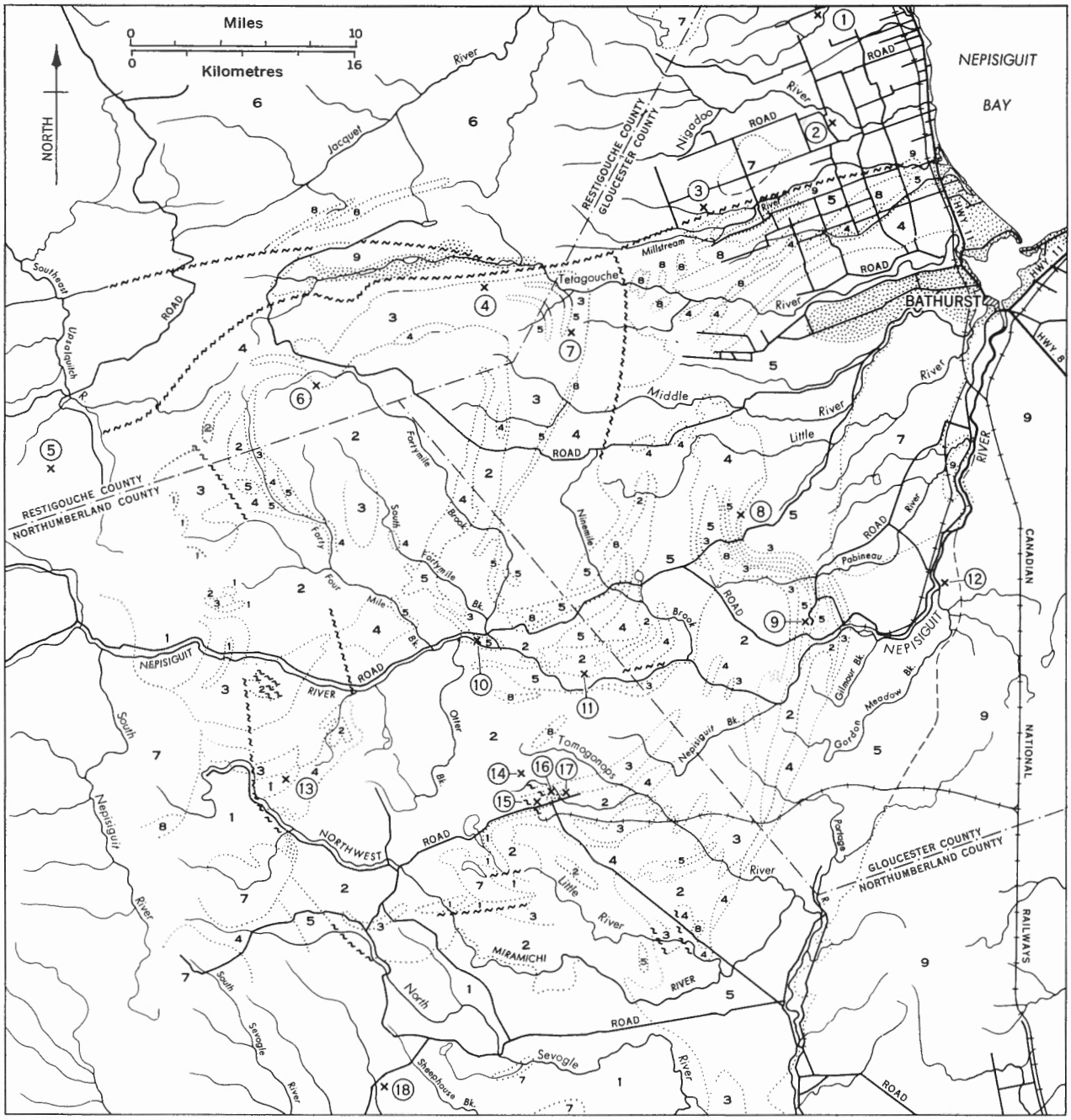
of quartz, sericite, chlorite, magnetite, biotite, calcite, siderite, and in places, barite; (2) pyrrhotite with chalcopyrite, in a gangue of chlorite and quartz. The latter type transects both schistosity in the host rocks and the massive pyrite-type ore. Other metallic minerals in the sulphide deposits are: magnetite, hematite, stannite, cassiterite, boulangerite, enargite, native silver, domeykite, marcasite, acanthite, argentite, gold, cubanite, covellite, chalcocite, bornite, native bismuth, limonite, beudantite, freibergite, anglesite, cerussite, and pyromorphite. Trace elements are silver, arsenic, bismuth, cadmium, germanium, indium, manganese, molybdenum, tin, titanium, and vanadium.

The massive pyritic sulphide deposits form lenticular, pipe-like, or sheet-like bodies, that are partly banded, folded, brecciated, and perhaps recrystallized. Some appear to be localized or thickened in places around the crests of folds; others do not. Some of the deposits are rudely zoned into zinc–lead-rich portions, copper-rich parts, and barren pyrite, but generally the zones are erratic and intermixed. The deposits are regarded by some as epigenetic replacement bodies in places possibly directly related to intrusive rocks, and by others as chemically precipitated sedimentary deposits. Evidence for a sedimentary mode of origin is largely inferred.

TABLE VII-2 *Size, tonnage, and grade of some sulphide deposits of the Bathurst area, New Brunswick (Davies and Smith, 1966; modified after McAllister, 1960)*

Deposit	Location (approx.)	Size (Feet)			Thousand tons	Grade			
		Length	Width	Depth		% Cu	% Pb	% Zn	Ag oz./ton
Anaconda (Caribou)	47° 34′ N	3800	20–130	1200	28,000		2	5	1
Anaconda (Armstrong Brook)	47° 36′ N 66° 17′ W	A—450 B—450	50 (max) 150 (max)	800 800					
Brunswick No. 6	47° 28′ N 65° 53′ W	1000	200	1000	22,278 4,350 1,684	0.4 0.5 0.8	1.9 0.4 0.5	5.0 1.0 0.5	1.6 0.7 0.3
Brunswick No. 12	47° 28′ N 65° 53′W	1200	200	1400	15,787 13,500	0.3 0.7	3.6 1.0	9.5 3.2	2.6 1.6
Captain mines	47° 17′ N 62° 52′W	500	80	800	802	1.15			
Canoe Landing Lake	47° 25′ N 66° 12′ W	3200	30	700	3,500	0.5	0.5	1.5	1.5
Chesterville mines (massive zone only)	47° 07′ N 66° 12′ W	650	650	30	650	0.92	1.62	3.4	
C.M. and S. (Nepisiquit A)	47° 22′ N 66° 03′ W	700 700	75 50	350 500					
C.M. and S. (Nepisiquit B)	47° 22′ N 66° 03′ W	900	75	300					
C.M. and S. (Wedge mine)	47° 24′ N 66° 08′ W	850	50	500					
Heath Steele mines	47° 17′ N 66° 05′ W	400–600	50	400–800	4,700	1.1 1.4	3.0 1.3	7.2 3.6	3.3 2.0
Kennco (Murray Brook)	47° 32′ N 66° 27′ W	1000	300	800	23,000	0.44	0.86	1.95	0.91
Key Anacon mines	47° 36′ N 65° 42′ W	1200[1]		1500	3,000		2.35	6.4	2.16
Middle River Mining Co. (Texas Gulf Sulphur)	47° 17′ N 66° 19′ W	1000	75	1000	5,000	1.0	2	5	
New Calumet (Orvan Brook)	47° 38′ N 66° 08′ W	6000 800[3]	14[2] 14[2]	500 500	200		3.25	6.3	1.0
New Jersey Zinc (Teck Corp.)	47° 38′ N 66° 08′ W	1500	100	500	3,000		5	9	3.0
Stratmat (C.M. and S.)	47° 19′ N 66° 06′ W	600	100	500					

1 Several lenses along zone 1,200 feet long
2 Maximum width
3 Ore zone



PENNSYLVANIAN

9 *Undivided*

DEVONIAN

8 *Mafic intrusive rocks*

7 *Felsic intrusive rocks*

SILURIAN

MIDDLE SILURIAN

CHALEURS BAY GROUP

6 *Undivided*

ORDOVICIAN

MIDDLE ORDOVICIAN

TETAGOUCHE GROUP (1–5)

5 *Slate division*

4 *Greenstone division*

3 *Porphyry division*

2 *Felsic volcanic division*

1 *Siliceous metasedimentary rocks*

MINING PROPERTIES

(1) Keymet
(2) Nigadoo Mines Ltd.
(3) Sturgeon River Mines
(4) New Calumet
(5) New Jersey Zinc
(6) Anaconda Co. (Caribou)
(7) Anaconda Co.
(8) Brunswick No. 12
(9) Brunswick No. 6
(10) Consolidated Mining and Smelting Co.
(11) Consolidated Mining and Smelting Co.
(12) Anacon Lead Mines (New Larder "U")
(13) Middle River Mining Co. Ltd.
(14) Stratmat
(15) Heath Steele
(16) Heath Steele Mines Ltd.
(17) Heath Steele Mines Ltd.
(18) Kennco Explorations (Canada) Ltd. (Clearwater)

Drift-covered area . . . . . *Sulphide deposit* . . . . . × *Fault (assumed)* . . . . . ~~~~

GSC

FIGURE VII-8. Principal base-metal deposit, Bathurst–Newcastle district, New Brunswick (Roy, 1961).

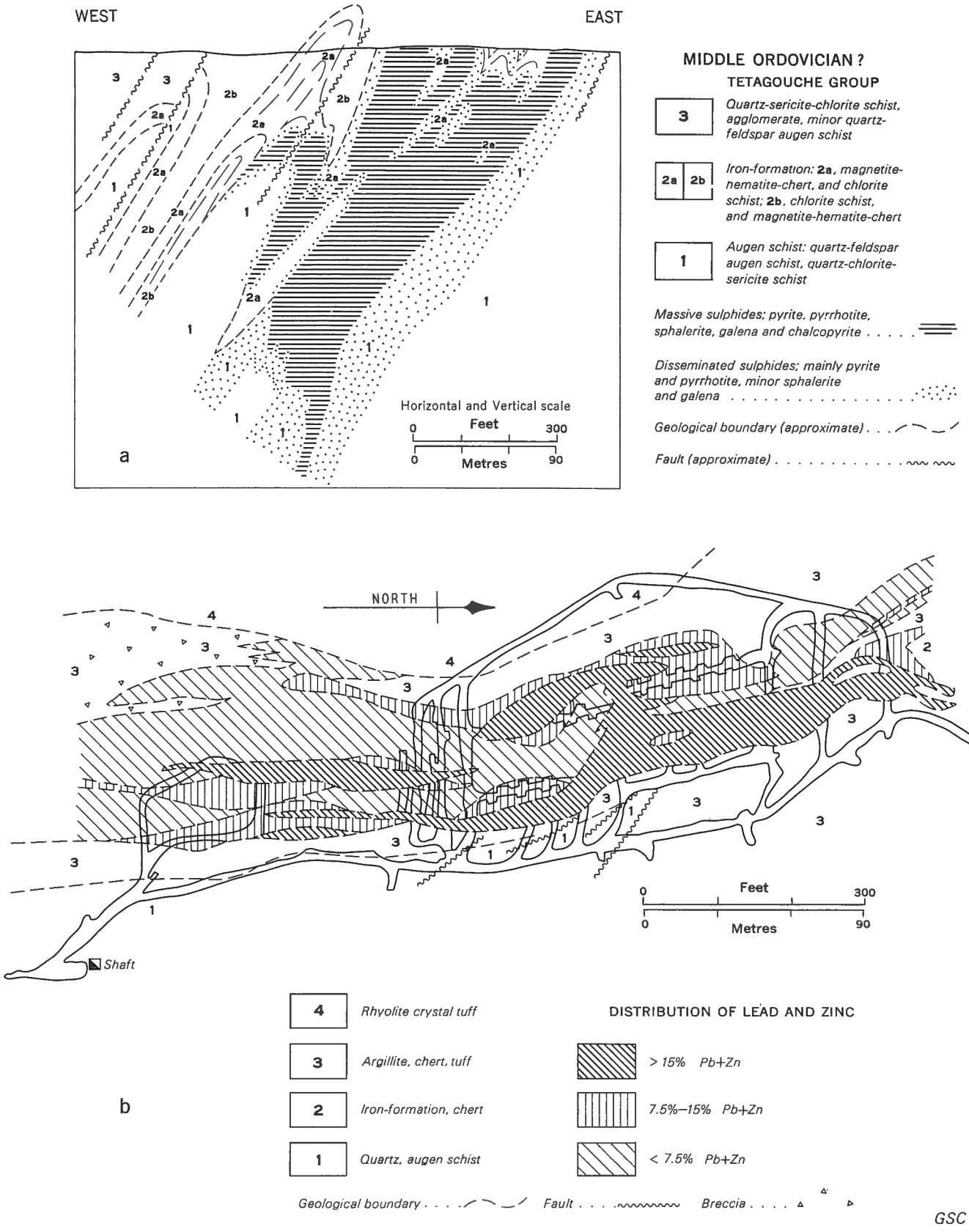


FIGURE VII-9. Bathurst sulphide deposits, New Brunswick (Davies and Smith, 1966). (a) Cross-section through the Brunswick No. 6 deposit. (b) Geology of the 650 level, Brunswick No. 12 deposit showing distribution of lead and zinc.

The Brunswick No. 6 ore deposit occurs in complexly folded, regionally metamorphosed acid volcanic and sedimentary rocks of the Tetagouche Group that have been intruded by sills of diabase and gabbro (Fig. VII-9a). Near surface the central part of the large massive sulphide lens, which is being developed by open-cut, contains a well-mineralized and replaced core of iron-formation of which only the more siliceous parts have resisted replacement. The lens is 400 feet thick and 1,000 feet long. A smaller, saddle-shaped, massive pyrite body, at most 100 feet thick, merges with the main lens. A third lens about 35 feet thick, formed of pyrite, pyrrhotite, and chalcopyrite, follows a sheared zone in the footwall schist. The deposits hold more than 28 million tons averaging 4.1 per cent zinc, 1.6 per cent lead, 0.4 per cent copper, and 1.4 ounces per ton of silver.

Brunswick No. 12 deposit (Fig. VII-9b) also consists of three parts: a large, irregular, massive pyrite lens with sphalerite, galena, and chalcopyrite; a large irregular body of pyrite; and a chalcopyrite–pyrrhotite-rich zone. The deposit held more than 28 million tons above the 2,500-foot level averaging about 9 per cent zinc, 3 per cent lead, 0.3 per cent copper, and 2.3 ounces of silver per ton. The chalcopyrite–pyrrhotite zone contains about 2,700,000 tons averaging 1.33 per cent copper. The high-grade zinc–lead part of the deposit shows an intimate association with a reverse S-shaped fold in iron-formation at the northern end of the body.

The *Wedge mine* (4) sulphide orebody lies concordantly between a porphyritic and tuffaceous rhyolite and overlying argillaceous fragmental rock on the south limb of a complex fold in the Tetagouche Group. The ore consists of 95–100 per cent sulphides, averaging 3 per cent copper and 1.75 per cent zinc. It forms a fishhook-shaped lens, 15 to 150 feet wide, 1,200 feet long, and 500 feet deep, controlled by an east-plunging drag-fold. Pyrite is the main mineral, with varying amounts of chalcopyrite, sphalerite, galena with associated silver, and minor tennantite. The relatively high copper content is unusual for deposits in this area. Chalcopyrite, the copper-carrying mineral, is concentrated within coarser grained pyrite along the hanging-wall in the wider eastern part of the deposit, along the northern rhyolite contact, and near the footwall in the eastern part. Local bands of sphalerite and galena within fine-grained barren pyrite also occur near the footwall and in the eastern part of the deposit.

At the *Heath Steele mine* (5) twelve irregular and pipe-like sulphide bodies occur in the minor folds in siliceous and basic volcanic and sedimentary rocks of the Tetagouche Group, at or near the contact between quartz–feldspar augen schist and a mixed assemblage of chlorite schist, quartz–sericite schist, iron-formation, and feldspathic sedimentary rocks (Fig. VII-10). Pyrite, sphalerite, and galena are the main sulphide minerals in most of the deposits, but chalcopyrite and pyrrhotite predominate in one and are important in three others. One of the orebodies was partly covered by a gossan 45 feet thick, the basal 2 to 6 inches of which was black mud rich in the silver minerals acanthite and argentite. The upper 80 feet of the orebody contained disseminated secondary chalcocite.

The *Keymet* (1) deposit, 15 miles northwest of Bathurst, occurs in a fault zone in sedimentary rocks of the Chaleurs Bay Group of probable Silurian age, north

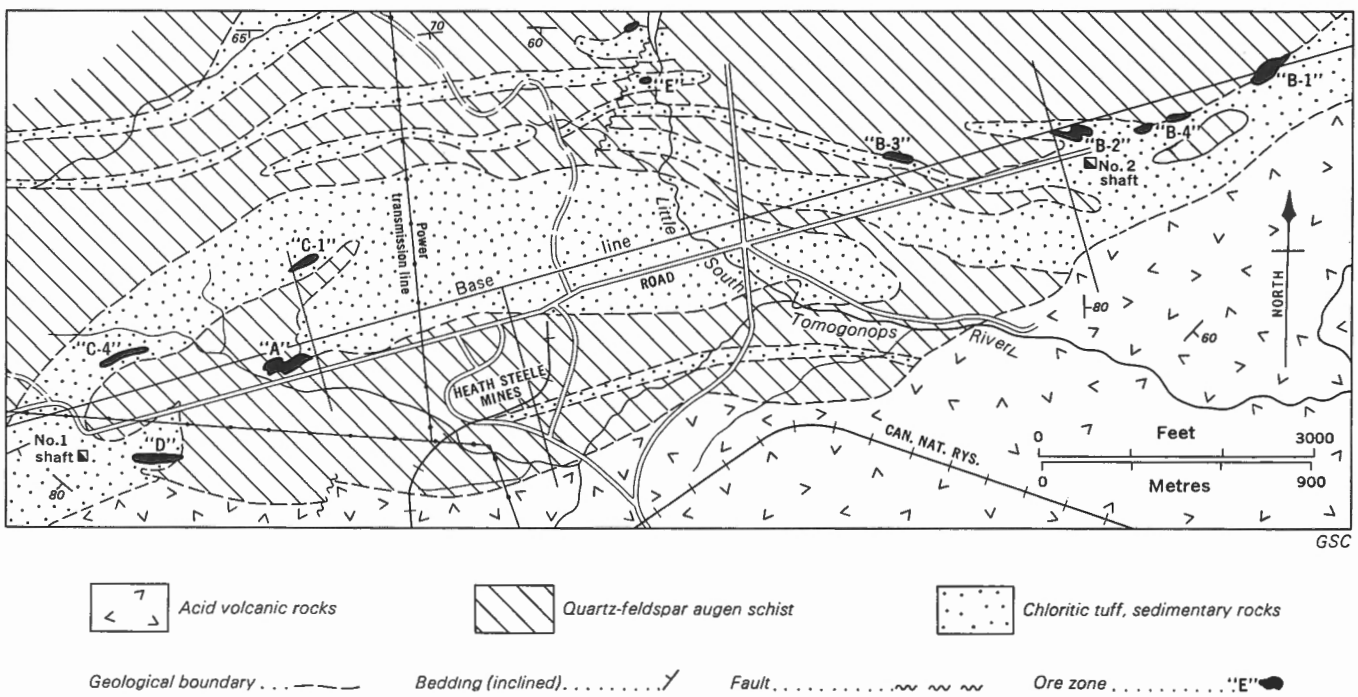


FIGURE VII-10. Geology and ore zones, Heath Steele mine, New Brunswick (McFarlane and Gates, 1966).

of the Rocky Brook – Millstream fault (Fig. VII-8). Lenses and veins of sulphides penetrate shale, conglomerate, and argillite along a crosscutting fault and fracture zone. The ore consists mainly of pyrite, arsenopyrite, pyrrhotite, sphalerite, galena, and chalcopyrite, with minor tetrahedrite and marcasite, in a gangue of calcite, quartz, and a little fluorite. Reserves at the mine were estimated to be 112,000 tons averaging 4.41 per cent lead, 4.31 per cent zinc, 0.55 per cent copper, and 1.42 ounces of silver per ton.

The *Nigadoo* (2) deposits, 2½ miles north of the Rocky Brook – Millstream fault, occur as lenses and veins along a fault zone within green and graphitic schists, shales, and argillites of the Elmtree Group in the contact zone of a differentiated intrusive porphyry stock of probable Devonian age and also within the intrusive itself. The best ore is mainly within the porphyry or in the argillite a short distance from the contact. The ore minerals are similar to those at the Keymet deposit, and they extend to a depth of at least 900 feet.

Mindamar (Stirling) mine (30). The Stirling zinc mine on Cape Breton Island was first opened as a copper prospect in 1904, and developed as a zinc mine in the period 1927–36. It was reopened in 1950 as a lead–zinc–copper mine, the Mindamar mine, and produced about a million tons of ore before closing again in 1956. Pyrite, sphalerite, galena, chalcopyrite, and minor tennantite form lenses of massive, fine-grained, banded sulphides in a zone of talc–carbonate–sericite schist within rhyolitic flows, tuffs, and interbanded minor sediments similar to those of the Middle Cambrian Bourinot Group. Banding in the ore is parallel to the strike of the deposits and to the schistosity of the enclosing volcanic rocks and schists. A variety of igneous rocks intrude the group in the vicinity, and a major fault zone intersects these rocks near Stirling.

Central mineral belt, Newfoundland. The mines producing base metals in Newfoundland in 1966—the Buchans, Tilt Cove, Gullbridge, Little Bay, Whalesback, and Rambler mines—are all in sulphide bodies within metavolcanic rocks that are invaded by a variety of intrusive rocks ranging from ultrabasic to acidic. The volcanic rocks, for the most part now chloritic schists of basic composition, are intercalated with graptolitic slates in places, and are considered to be mainly Ordovician, although some may be Silurian. Many past producers, such as the Terra Nova, Betts Cove, Little Bay, and Pilley's Island mines in the same geological environment and region, have made Newfoundland an important copper, zinc, and lead producer. In 1966 the combined output of copper, zinc, and lead from these mines was valued at more than $30 million, mainly from the Buchans mines.

Buchans mines, Newfoundland (13). The first Buchans River deposit, "Old Buchans", was found in 1905 (Fig. VII-11). A body of 100,000 tons of very fine grained zinc–lead–copper sulphides was indicated. The mine was closed in 1911 because of metallurgical difficulties in processing the ore, but these difficulties were subsequently overcome in 1925. The Lucky Strike and the Oriental deposits, discovered in 1926, are now largely mined out. After mining and milling began in 1928 they produced more than 9 million tons, averaging 16.5 per cent zinc, 8.1 per cent lead, 1.54 per cent copper, 3.37 ounces silver, and 0.54 ounce gold per ton.

The Rothermere and Maclean orebodies outlined in 1947 and 1950 are presently producing. They form irregular, massive, gently pitching replacement lenses, mainly within tuff and breccia in the basic volcanic rocks of the Buchans Group (George, 1937). These rocks are intruded by diabase, granite, andesite, dacite, and quartz porphyry dykes. The deposits lie in the loop of a sigmoid curve in the Buchans Group in a re-entrant into a Devonian granite batholith to the west, and between two northeast trending diorite plutons along strike. The orebodies are both about 2,000 feet long, as much as 750 feet wide, and from 20 to 200 feet thick. They are intersected by zones of soft schist. Smaller zones of low-grade, disseminated sulphides occur below and marginal to the high-grade orebodies (Fig. VII-11b). The ore is of three main types: breccia ore containing numerous rock fragments, baritic ore containing as much as 35 per cent barite, and normal ore consisting of massive sulphides with a few rock fragments and minor barite. The normal ore is a fine-grained mixture of sphalerite, galena, pyrite, chalcopyrite, and a little tetrahedrite. The gangue consists mainly of barite with some quartz and calcite. Minor amounts of bornite, native silver, gold and silver tellurides, hematite, and sulphates of zinc, copper, iron, and lead are present.

Tilt Cove mines (7). The old Tilt Cove (Union) mine (Pl. VII-2), discovered in 1857, produced about 1½ million tons of high-grade copper ore from 1864 to 1917. It was reopened in 1957 and yielded an additional 6 million tons of lower grade ore. Between 1869 and 1876 small shipments of selected ore contained 24 per cent nickel and some gold. Two separate deposits, the East and West, occurred in chloritic schist and altered pillow lava of the Ordovician Headland Group at or near the contact of dykes and sills of intrusive, serpentinized, and steatitized pyroxenite. Dykes of quartz porphyry or granite cut the pyroxenite. They formed the footwall of the ore in the West mine and separated the ore masses in the East mine, the massive sulphides following both contacts of the dyke. The dykes are believed to be related to the intrusive Burlington granite that has been dated as Devonian.

The East mine deposits were irregular, massive, fine-grained pyritic lenses with abundant chalcopyrite and some sphalerite. Large masses of magnetite that were partly brecciated and cemented by sulphides were encountered locally. Some specular hematite occurred in vugs in the magnetite. The West mine deposit was

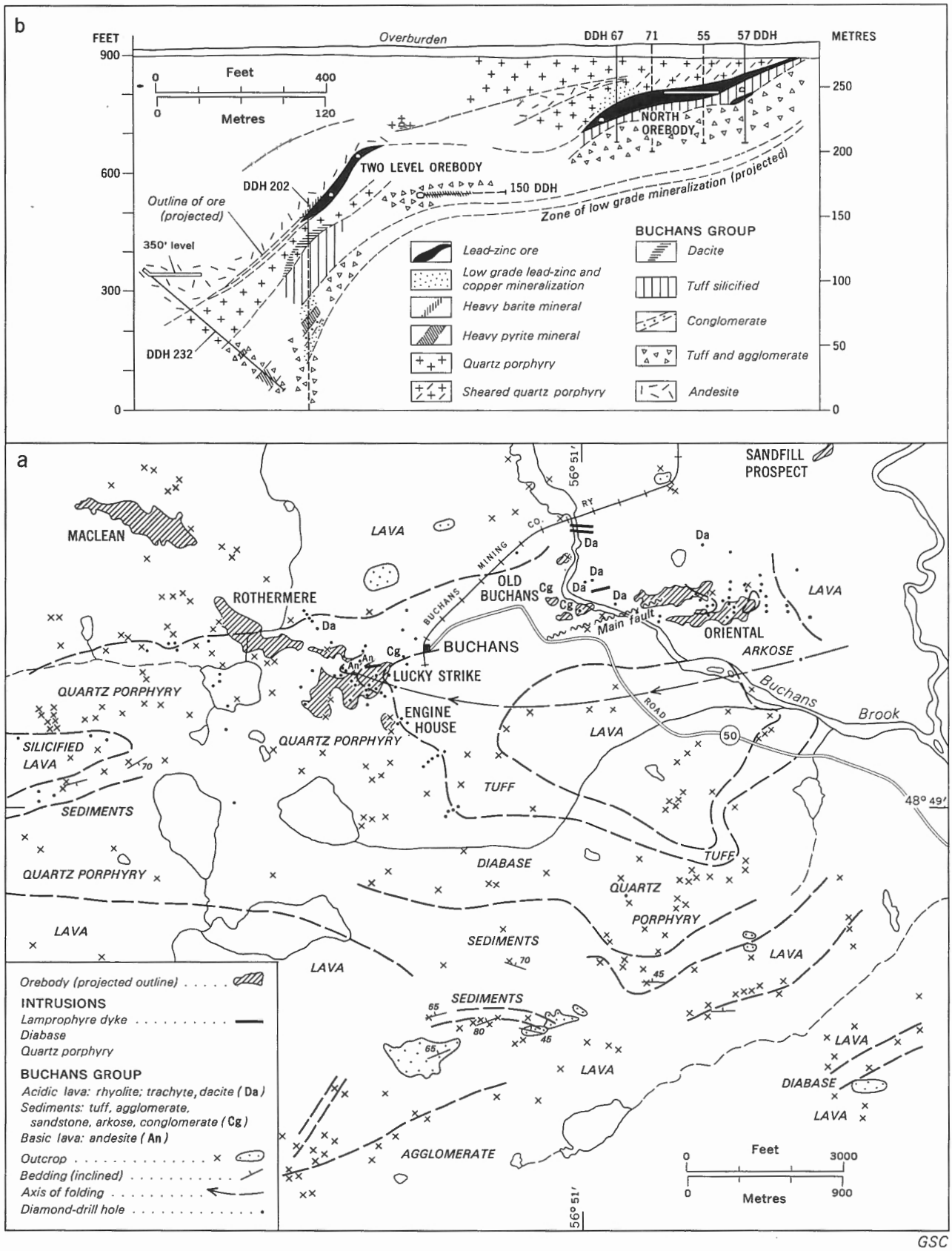


FIGURE VII-11. Buchans copper-zinc-lead deposits, Newfoundland (George, 1937). (a) Geology in vicinity of Buchans orebodies. (b) Cross-section through Lucky Strike North deposit.



PLATE VII-2. First Maritime Mining Corporation Limited copper–zinc–lead mine at Tilt Cove, Newfoundland.

formed of veins and disseminations of chalcopyrite, pyrite, pyrrhotite, quartz, and ankerite within altered chloritic andesite. Nickel- and cobalt-bearing minerals, including maucherite, niccolite, chloanthite, gersdorffite, rammelsbergite, arsenopyrite, violarite, and millerite, were found in the West mine as lenses of breccia or veins of dense grey dolomite in a talc–carbonate zone that was formed where serpentinite is intruded by a quartz porphyry dyke. Rare calcite veins in the quartz porphyry carry niccolite, maucherite, gersdorffite, and rammelsbergite.

Betts Cove (8) mine. In 1875 lenticular deposits of massive sulphides were opened in chloritic schist within pillow lava of the Ordovician Headland Group on Burlington Peninsula, Newfoundland. The group is intruded by a serpentinite sill or dyke and by bodies of diorite and quartz diorite that almost engulf the older rocks west of the mine. All Betts Cove copper deposits are less than 500 feet from exposed diorite, which is itself slightly mineralized in places. Olivine diabase dykes cut one of the ore zones, and small basalt dykes are common in both the lavas and diorite. The ore consists mainly of pyrite and chalcopyrite with bands of sphalerite and jasper. The mine produced 130,682 tons of high-grade copper ore carrying traces of gold, as well as several thousand tons of material that was rejected because of its zinc content. Several attempts to reactivate the mine have failed.

The *Little Bay* (9) deposit on Notre Dame Bay was discovered in 1877 in a fault zone in pillow lavas of probable Ordovician age. It produced about 200,000 tons of high-grade copper ore prior to 1900, and more than a million tons of gold-bearing copper ore since it was reopened in 1961. The ore occurs as veins and disseminations of pyrite, chalcopyrite, sphalerite, and magnetite with considerable quartz gangue in lenticular zones of chloritic schist within altered basic lava. The main ore lens, 30 feet wide and 900 feet long, strikes easterly, dips steeply south, and extends more than 1,000 feet down dip. Diorite intrudes the lavas about half a mile west of the mine, and numerous barren diabase dykes, as thick as 25 feet, occur within the ore zone. At *Whalesback Pond mine* (10), in similar chloritic schists about 3 miles to the west, the main lenticular pyrite–chalcopyrite ore zone is 40 feet wide and 1,200 feet long; it strikes easterly and dips steeply south. It extends below the 850-foot level and averages 1.75 per cent copper. A second ore zone, known as the Little Deer Pond deposit, has been found near by.

Consolidated Rambler mines (6). The Rambler prospect was found in 1905 on Burlington Peninsula and developed for production in 1964. There are two main ore zones, the "Rambler" and the "East". Both occur in chlorite–sericite schists north of a salient of the Devonian Burlington granite that intrudes the Ordovician Baie Verte volcanic rocks. A host of minor intrusions of basic to acidic composition intersects the volcanic rocks. The Rambler orebody is 5 to 25 feet thick and 450 feet long, dipping 40 degrees to the north. It consists of pyrite, chalcopyrite, and sphalerite impregnations in silicified, sericitized, chloritic schist, and has a high gold and silver content in places. The Rambler deposit averages 1.4 per cent copper, 2 per cent zinc, 0.12 ounce gold, and 1 ounce silver per ton. The East orebody about 5,000 feet to the east comprises heavily pyritized chlorite schist with some chalcopyrite. The deposit does not outcrop. It is reported to hold 2 million tons averaging 1.65 per cent copper.

Pilley's Island (11) mines. A pyritic deposit on Pilley's Island in Notre Dame Bay, carrying chalcopyrite, minor sphalerite, rare galena, and abundant quartz, was mined sporadically from 1889 to 1908 for the pyrite, from which sulphur was recovered; about 600,000 tons of ore were produced. In 1920 diamond drilling revealed that part of the deposit had been offset by faulting and was estimated to hold a similar tonnage of pyrite–chalcopyrite ore. The occurrence is unusual in that the ore is found within green rhyolite flows of the Ordovician and Silurian Roberts Arm volcanics, whereas all other copper occurrences in this part of Newfoundland are in basic volcanic rocks.

Lead–Zinc–Copper–Silver

Galena–Sphalerite–Chalcopyrite–Silver

The *Magnet Cove* (8) barite quarry and underlying lead–zinc–copper–silver mine near Walton, Nova Scotia (Fig. VII-12), lie in an area of faulted Carboniferous sedimentary rocks—red sandstone, shale, anhydrite, and

gypsum (Boyle and Jambor, 1966). In 1956 diamond drilling at the site of the barite quarry disclosed a lead–zinc–copper–silver orebody below, at a depth of 250 feet, lying mainly within a breccia zone along the footwall fault contact of the barite deposit (Fig. VII-13). Since its opening in 1961 the sulphide deposit has yielded more than 10 million pounds of lead, 5 million pounds of zinc, 2 million pounds of copper, and 2 million ounces of silver. The ore is reported to average 10 per cent lead, 4 per cent zinc, 1 per cent copper, 15.7 per cent iron, 30 ounces per ton silver, and traces of gold, cadmium, nickel, cobalt, arsenic, antimony, and bismuth. The principal minerals in the deposit are siderite, barite, hematite, nickeliferous and cobaltiferous pyrite, marcasite, galena, sphalerite, chalcopyrite, tennantite, acanthite, proustite, and pearcite. Although similar, this assemblage differs from that in Bathurst, New Brunswick, mainly in its higher barium, lead, and silver content and its framboidal texture.

Boyle and Jambor (1966) suggested that the source of the elements in the barite and sulphide deposits at Walton is the enclosing sedimentary rocks. They look to the Cheverie sandstones which have higher-than-average contents of barium over large areas, and to the Horton Bluff Formation of black pyritiferous shales that are moderately enriched in sulphur, copper, and silver, and in which the pyrite is particularly high in nickel, cobalt, and arsenic. The method of concentration is uncertain. Circulating brines may have been the medium of deposition. On the other hand, an igneous source of Triassic or younger age may underlie the area.

Lead–Zinc

In Gaspé Peninsula, Quebec, attempts were made as early as 1665 to obtain lead and silver from occurrences along the shore of Gaspé Bay. The occurrences are essentially small fissure veins, mineralized with calcite, quartz, galena, sphalerite, and a little chalcopyrite. They are found in fractured Lower Devonian Gaspé Limestone at Percé, in the overlying Middle Devonian Gaspé Sandstone at Indian Cove, and with a little fluorite in the Ordovician Trenton limestone at Baie St. Paul. In central Gaspé on the flanks of the Devonian intrusives there are numerous occurrences of sphalerite and galena. At the Candego mine, on the north flank, galena and sphalerite with minor pyrite occur in quartz veins in faults and fractures in silicified shale. During 1948–51 about 14,000 tons of ore averaging 3–6 per cent lead, 1–6 per cent zinc, and 3–5 ounces per ton of silver was

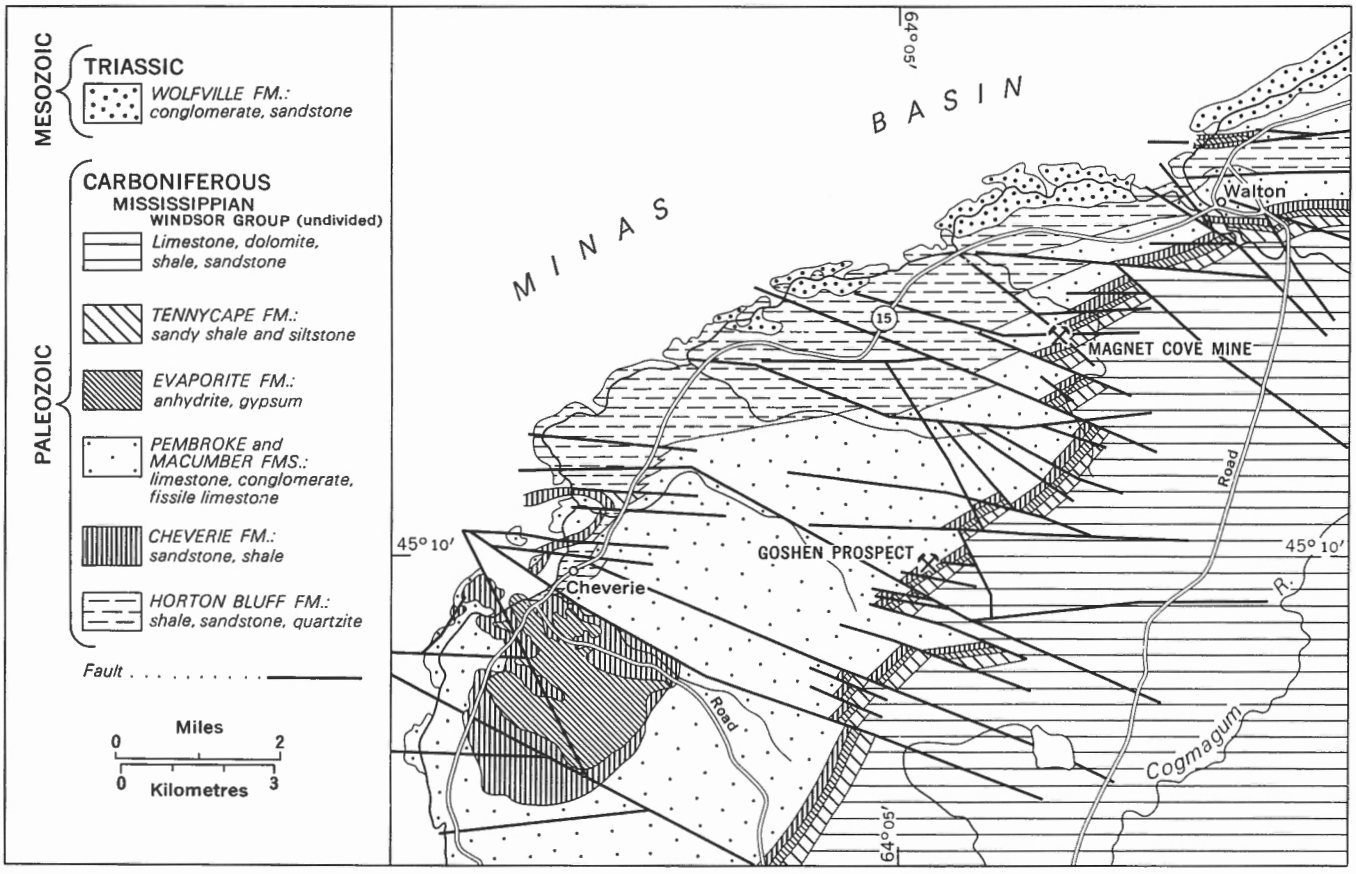


FIGURE VII-12. Geology of the Walton–Cheverie area, Nova Scotia (Boyle and Jambor, 1966).

milled. At the Federal mine, on the south flank, lead–zinc minerals occur in the form of veins and breccia zones in Lower Devonian limestone, argillite, and tuff. The veins there also cut syenite that is intrusive into the Lower Devonian rocks, but not the Gaspé Sandstone.

In Nova Scotia, numerous lead–zinc occurrences have been noted in fissures and disseminations in Carboniferous limestones and dolomites, notably at Jubilee and Smithfield, and associated with barite at Magnet Cove. At Leadville, galena, sphalerite, and some barite occur as replacements in grey limestone breccia at or near its contact with Horton shale and sandstone. The breccia, containing blocks of galena–sphalerite, is exposed over a width of 25 feet for a length of 220 feet.

On the west coast of Newfoundland galena and sphalerite occur in small vertical fissure veins with barite and celestite in places, in Ordovician Table Head limestone, and in grabens of brecciated Mississippian rocks. Extensive occurrences of disseminated sphalerite have recently been found within a zone 200 feet thick of brecciated, gently dipping beds of dolomitic limestone of the Ordovician St. George Formation north of Daniels Harbour. Reserves of 3.7 million tons averaging 7.3 per cent zinc are indicated. Lead–zinc occurrences are also reported on the north coast at Bear Cove and Fleur de Lys and in metasedimentary rocks that are invaded by granite batholiths on the south coast of Newfoundland, particularly in the La Poile Bay and Bay d'Espoir areas.

At the *Talisman (Silver) mine*, Cape Breton Island, pyrite and argentiferous galena are disseminated through Mississippian Windsor limestone and Devonian rhyolite, and fine-grained galena is disseminated through a sandstone bed—a basal member of the Pennsylvanian Canso Group. Diamond drilling has shown this bed to be 37 feet thick with an average grade of 3 per cent lead. It has been traced along strike for 6,900 feet and down dip for 1,000 feet.

The *La Manche* vein on Placentia Bay in the Avalon

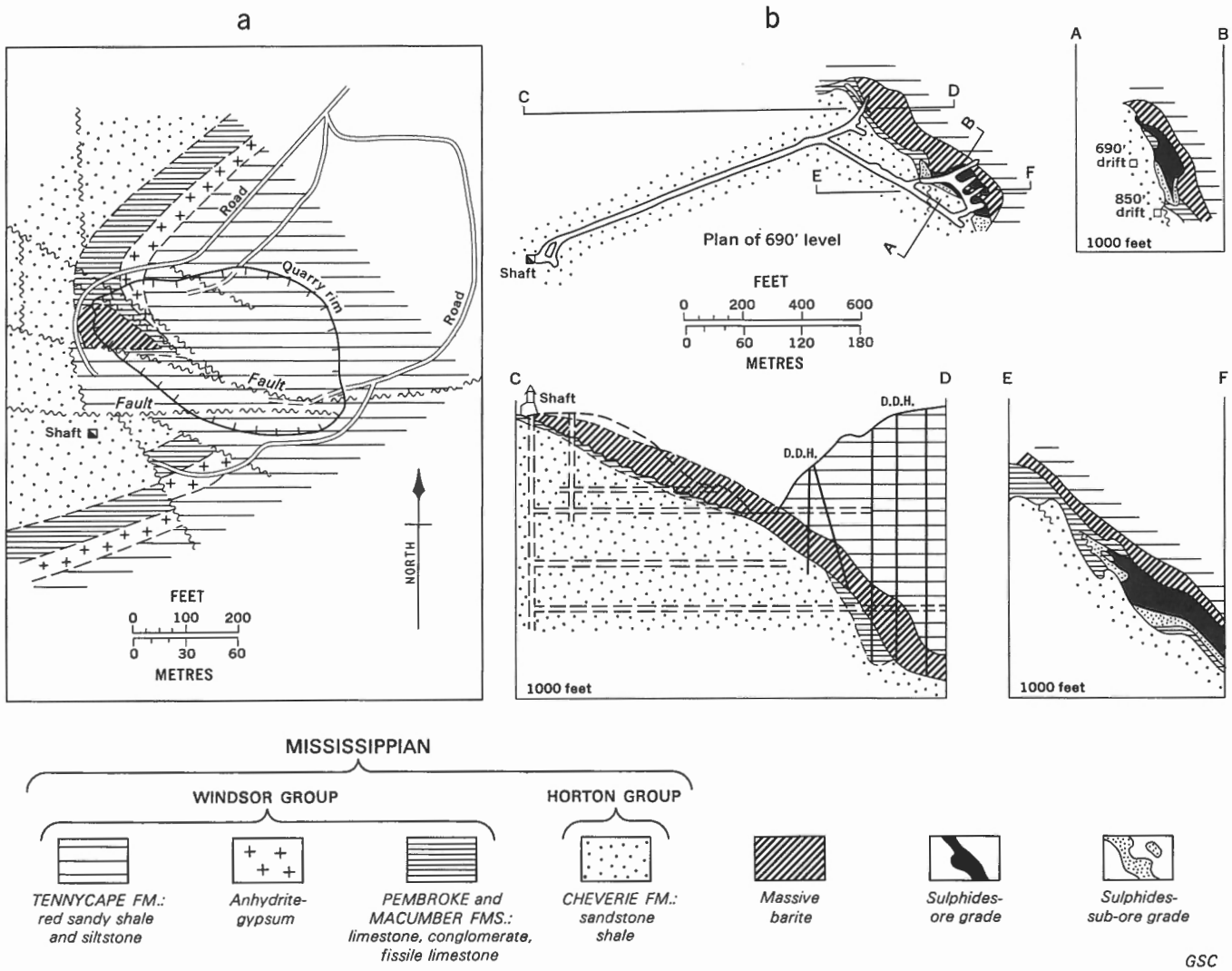


FIGURE VII-13. Magnet Cove barite–sulphide mine, Walton, Nova Scotia (Boyle and Jambor, 1966). (a) Geological map of quarry and mine. (b) Plan of 690 level and cross-sections A–B, C–D, and E–F.

Isthmus, Newfoundland, was opened in 1857 and mined until 1873 (later sporadically) for a length of 1,685 feet and to a depth of 400 feet. Although the mine is now flooded the ore is not exhausted. The vein extends inland for 10,000 feet and beneath the bay for an unknown distance. It is formed of coarse calcite with aggregates of galena and minor amounts of sphalerite, chalcopyrite, quartz, barite, and pyrite. It follows steeply inclined, northeast-trending, parallel fissures within and along the contact of a trap dyke that intersects thinly bedded, westerly dipping green siltstones of the Hadrynian Connecting Point Group. The vein shows no sign of pinching out, but varies in width from a few inches to 5 feet. It averages 5 per cent lead over an average width of 2½ feet for at least 700 feet and carries traces of gold and silver. The vein was formed by repeated opening and filling of the fissure and is characterized by banding and abundant vugs. The presence of so much calcite as well as galena is difficult to account for in these rocks. The nearest limestones are those of Lower Cambrian age some 10 miles away. They once overlapped the La Manche area, and conceivably may have contributed calcite as well as barite to the vein.

Lead–zinc–silver occurrences with associated quartz calcite, and barite gangue were once mined at the *Silver Cliff* mine near Argentia. There small veins and breccia cements in the Hadrynian Bull Arm Formation carry auriferous pyrite, argentiferous galena, sphalerite, chalcopyrite, quartz, carbonates, and barite. Similar occurrences have been noted near Placentia, in Southeast Arm (Strouter adit), in Northeast Arm southeast of Seven Islands in Placentia Bay, and at Jerseyside.

Gold–Silver

The most recent production of gold and silver has come as a byproduct of smelting of base-metal ores in Newfoundland, New Brunswick, and Quebec. Silver appears to be associated with the galena-rich deposits, and the gold with the chalcopyrite–pyrite occurrences. In 1966 more than $8 million in silver and $1.8 million in gold were recovered from base-metal ores.

Gold nuggets found in 1823 in gravel in the Chaudière River valley near the mouth of the Gilbert River led to placer mining in the Eastern Townships of Quebec in the period 1847–85 when about $3 million in gold was recovered. Many rough angular nuggets weighing from 1 ounce to 30 ounces were found, some with bits of quartz attached as if they were not far removed from their quartz vein source. Most gold was obtained from the older, preglacial gravels. Prospecting for the bedrock source was rewarded by discoveries of several precious metal-bearing copper deposits and many quartz veins, but no high-grade lode gold deposit has yet been found in the region.

In Nova Scotia lode gold mining began in 1862, and production reached a high of more than a million dollars in 1939. Scores of small mines produced gold from quartz veins that intersect slate and quartzite of the Ordovician Meguma Group. The veins follow or cut across the bedding planes and become more numerous and wider on the anticlines and domes where they form 'saddle reefs'. Free gold is commonly concentrated in shoots within the quartz veins along with pyrite, arsenopyrite, calcite, and galena. The deposits show a spatial relationship to intrusive Devonian granites and appear to have been deposited from solutions generated by them. On Middle River in Cape Breton Island, gold has also been found in quartz veins in altered sedimentary rocks of Precambrian or early Paleozoic age and in the granitic rocks that intrude them. Some has been recovered from the gravels derived from them.

In Newfoundland small amounts of gold have been produced at Ming's Bight and Sop's Arm from metamorphosed Ordovician rocks.

Iron

The Appalachian Mineral Province has a great number and variety of iron occurrences including titaniferous magnetite lodes of magmatic origin in anorthositic rocks, magnetite replacement deposits of hydrothermal origin in basic volcanic rocks and sediments, magnetite–hematite–chert iron-formations of volcanic–sedimentary aspect, replacement hematite–siderite–ankerite masses, primary sedimentary oölitic hematite–chamosite–siderite beds, and manganiferous bog iron. Early mining was from relatively small deposits of the various types of iron ore in Nova Scotia and New Brunswick, but after World War I production was from the oölitic deposits at Wabana, Newfoundland, which closed in 1966 because of increasing competition.

Oölitic Hematite

Nictaux–Torbrook deposits (9). Iron deposits were discovered in 1820 and 1825 near Annapolis, Nova Scotia, and were mined from 1825 to 1916. The hematite–greenalite–magnetite ore is locally oölitic, and in places carries brachiopod and cephalopod shells. It forms thin beds in zones as thick as 18 feet that are intercalated with early Devonian slates and quartzites. The ore lies in two zones each more than 4 miles long and about a mile apart on opposite limbs of a syncline. Most of the ore has come from three bands that are 4 to 9 feet thick. The lean sedimentary iron-formation in which the ore occurs has been partly metamorphosed by intrusions of granite and gabbro, and magnetite has formed at the expense of some of the other iron minerals. The deposit is credited with a production of 350,000 tons of iron ore.

Similar but smaller deposits of this type have been noted in Middle Cambrian beds on Gillis Brook at Grand Mira South on Cape Breton Island. Near Arisaig, beds of oölitic hematite occur in the Lower Ordovician Browns Mountain Group, and can be correlated with the Wabana beds.

Wabana Iron mines (16). The Wabana iron deposits were discovered on Bell Island in Conception Bay, Newfoundland, in 1892 and have produced oölitic hematite ore from surface cuts and submarine mines. Five main bands and many thin seams of oölitic hematite are intercalated with fossiliferous Lower Ordovician sandstones and dark shales. They dip gently beneath Conception Bay where they lie conformably above the Cambrian exposed along the south shore of Conception Bay, and form part of an outlying basin, truncated by faults, underlying the bay. Three of the oölitic hematite bands, separated by 240 feet and 58 feet of strata, have been mined—the Lower (Dominion), Middle (Scotia), and Upper (Little Upper) beds. The possible extent and form of these deposits as they may exist beneath Conception Bay is shown on Figure VII-14. It is evident that large tonnages of potential ore, some of which may never be recovered, underlie the bay. Most production has come from the Lower bed that attains a thickness of 40 feet in places and has been mined down dip for a distance of almost 2 miles (Fig. VII-15). The Lower and Middle beds average 61 and 55 per cent hematite, respectively, and yield iron ore that ranges from 45 to 57 per cent iron; the average is about 52 per cent iron. Silica at 12 per cent, and phosphorus at 0.85 per cent are the most objectionable impurities in the ore. Sulphur is less than 0.03 per cent, and calcium less than 2 per cent. The silica occurs mainly as chamosite and detrital quartz, the phosphorus and calcium in the shell fragments, and the sulphur as pyrite.

The ore beds are composed essentially of sand-like grains of concentrically banded hematite, siderite, chamosite, quartz, and an occasional shell fragment set in a matrix of siderite and a paste of comminuted quartz, chamosite, hematite, shell fragments, manganese oxides, and carbonates—a mixture of chemically precipitated and subordinate clastic material (Pl. VII-3). The oölites are formed by direct chemical precipitation and accretion of hematite about nuclear clastic grains. Penecontemporaneous erosion, transportation, and deposition of oölites are shown by crossbedding, interformational conglomerate beds carrying oölitic hematite, and the off-shore bar arrangement within the ore zones. The ore zones are remarkable for their continuity. They include lenses of dark shale and sandstone. Variations in thickness suggest deposition as near-shore sand bars (Fig. VII-16). The beds show many features such as ripple-marks, rain drop impressions, cut and fill structures, algal structures, worm burrows, and abraded clastic materials—also suggestive of shallow-water deposition. The top of the Lower bed is eroded and overlain by a persistent conglomerate carrying pieces of hematite and shale. The overlying bed of

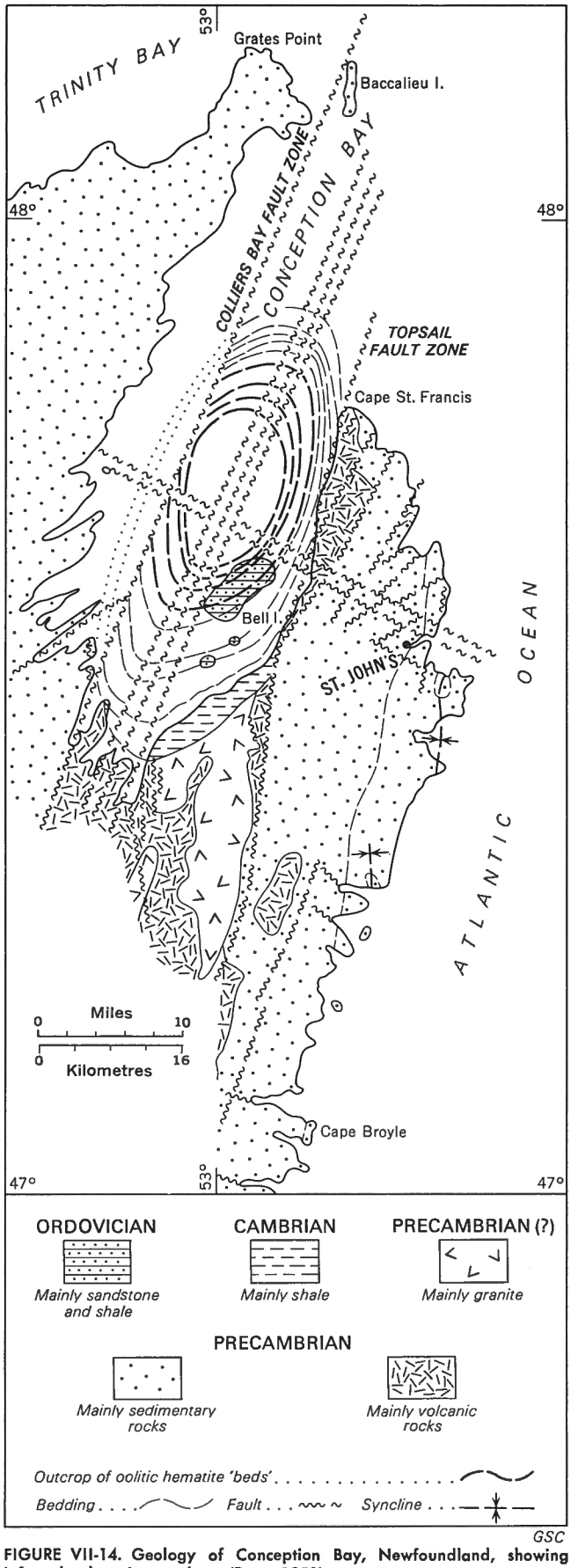


FIGURE VII-14. Geology of Conception Bay, Newfoundland, showing inferred submarine geology (Rose, 1952).

granular oölitic pyrite, as much as 1½ feet thick, carries graptolites and is succeeded by dark shale with dolomite nodules. The texture of the oölitic pyrite is remarkably similar to that of the ore beds and suggests deposition in a reducing environment. Thin nodular phosphatic beds composed of shell fragments, nodules, and oölites of calcium phosphate, siderite, and pyrite occur above the Middle bed.

Siderite–Hematite

Londonderry iron deposits (3). From 1849 to 1908 about 2 million tons of siderite and hematite ore were produced from one of the earliest known iron deposits in Canada, that of the Londonderry area in the Cobequid Mountains of Nova Scotia. The deposits are hematite–limonite-rich zones that resulted from surficial enrichment of irregular lenticular masses of ankerite–siderite that

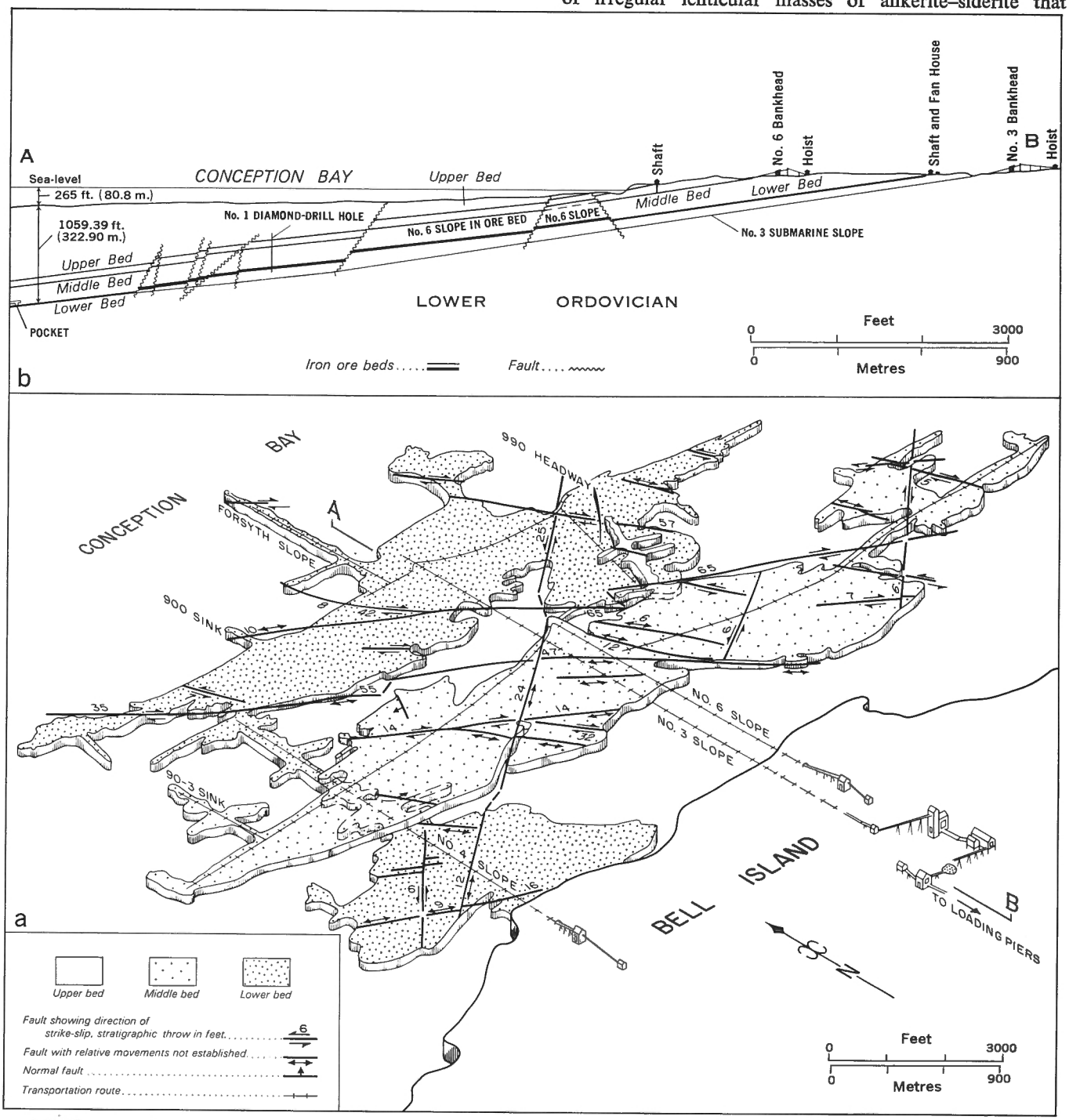


FIGURE VII-15. Wabana iron mines, Newfoundland. (a) Schematic drawing of workings in Lower, Middle and Upper beds (Norris, 1956). (b) Section along line of No. 3 slope (by J. B. Gilliatt).

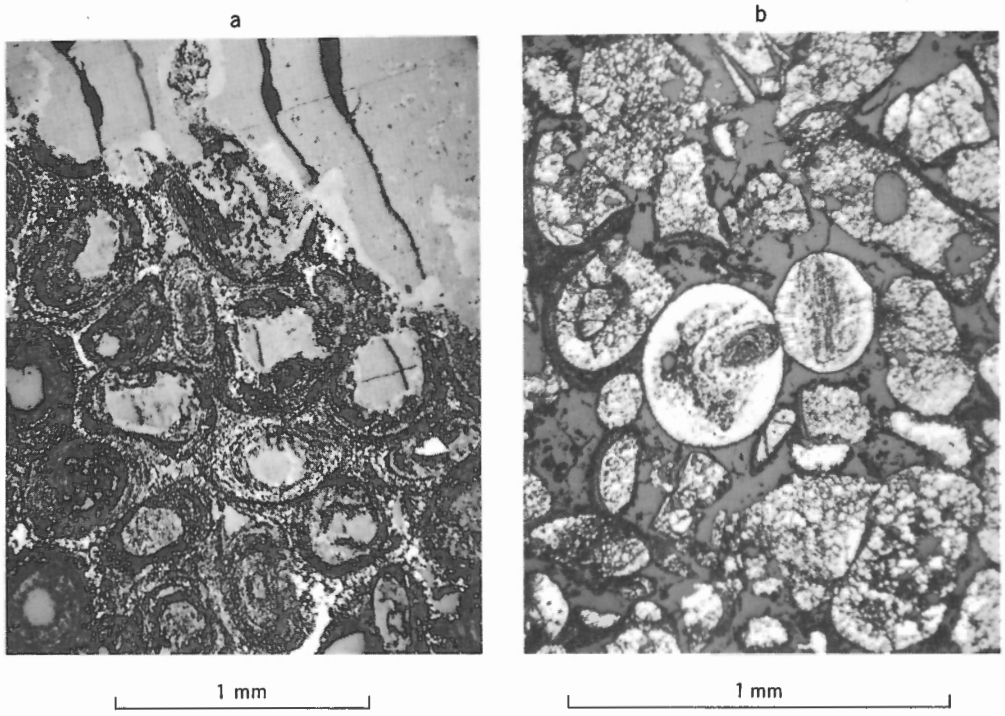


PLATE VII-3
Photomicrographs of Ordovician iron deposits at Wabana, Newfoundland; (a) Hematite oöliths with cores of detrital quartz and shell fragments; (b) Oölitic pyrite.

occur *en echelon* in a belt about 12 miles long parallel to a fault zone in pre-Carboniferous altered volcanic and sedimentary rocks. The carbonate bodies are as much as 100 feet wide and several thousand feet long; they contain irregular pockets of enriched oxide ore 100 feet wide and 300 feet deep underlain by unaltered carbonate. The enriched ore is reported to have carried from 47 to 57 per cent iron, and the unaltered carbonate material about 4 to 10 per cent iron. The origin of the carbonate masses is believed to have been connected with the emplacement of granitic intrusions into rocks of the pre-Carboniferous Cobequid complex.

Magnetite

Bathurst iron mine (6). Magnetite concentrations in iron-formation at the Drummond pit near Bathurst, New Brunswick, produced 200,000 tons of ore between 1910 and 1915. The magnetite results from the intrusion of quartz porphyry into Ordovician banded iron-formation. As noted in the section on Copper–Lead–Zinc, many of the large base-metal sulphide deposits of the Bathurst area are spatially related to magnetite bodies and iron-formations within the Tetagouche Group. Samples from the deposit have 42 to 49 per cent iron, 1 to 16 per cent manganese, 20 to 26 per cent silica, 0.6 to 0.7 per cent phosphorus, and 0.05 to 0.12 per cent sulphur.

Titaniferous Magnetite

Titaniferous magnetite and associated minor hematite and ilmenite occurrences have been explored sporadically in western Newfoundland. Disseminations and massive segregations occur within a gneissic, banded gabbro phase of the Indian Head anorthosite intrusion near Stephenville on St. George's Bay. Pegmatite dykes cutting these deposits have been dated by K–Ar methods at 910 m.y. The mass is therefore considered to be an inlier of anorthosite, part of the underlying Grenville basement. Segregated bands and disseminations of massive, hard, dark hematite intergrown with magnetite and ilmenite, appear in granitic rocks and gneiss to the north. At Steel Mountain, in the interior of the Flat Bay Brook anorthosite intrusion to the east, several dyke-like deposits of titaniferous magnetite cut the anorthosite. All these titaniferous magnetite occurrences carry strong traces of vanadium.

Manganese

Manganese is common in the Appalachian Mineral Province where it forms deposits of several types, and although some have been mined on a small scale from time to time, none is productive at present (Johnston and McCartney, 1965). The most numerous and important deposits are of replacement and secondary origin occurring within Carboniferous beds, particularly the limestones of the Windsor Group. They include deposits at Markhamville, Turtle Creek, and Shepody Mountain in New Brunswick; Tennycape in Nova Scotia; and the Magdalen Islands in the Gulf of St. Lawrence. A second, related type occurs within fissures in pre-Carboniferous rocks where solution and concentration has been effected by circulating groundwaters. Examples of this type have been found at Gowland Mountain, Jordan Mountain, and

Tetagouche Falls in New Brunswick; and at New Ross, Nova Scotia. A third type, potentially the most important source of manganese in the region, is of a primary sedimentary nature, found interbedded with sedimentary rocks of Cambrian age in Newfoundland, and Ordovician age in New Brunswick. A fourth type, bog manganese, is common throughout the region, but most occurrences are of limited extent and of insufficient purity to be worked at present. Bog manganese occurs near Moncton and on Renour River, in New Brunswick; and on Terra Nova River in Newfoundland.

Low-grade manganiferous hematite iron-formation near *Jacksonville* (7), northwest of Woodstock, New Brunswick, was mined for iron from 1848 to 1884; about 70,000 tons of ore was smelted. The iron-formation is interbedded with folded and metamorphosed red and green slates of Silurian age. The deposits carry from 9 to 22 per cent manganese in places.

Manganese-bearing beds occur in Cambrian limestone, breccia, and shale mainly within a zone at the contact of Lower and Middle Cambrian beds, at numerous localities in the Avalon Peninsula, and around *Conception Bay*, Newfoundland. Manganese carbonates and oxides occur as nodules, disseminated lenticles, and bands in the dark shales. The manganese rocks are essentially calcareous or dolomitic argillites with carbonates and oxides or carbonate–oxides of manganese, hematite, barite, and tri-calcium phosphate as the chief accessories. The primary manganese and barium-bearing carbonates in these beds are more or less surficially altered to pyrolusite, and in some places secondary enrichment in manganese oxides has occurred; for the most part, however, the manganiferous zone seldom exceeds 12 feet in thickness or 12 per cent manganese. At Manuels, samples across widths of 5.8, 7.8, 13.4, and 5.9 feet of the gently dipping beds contained 11.5, 10.73, 8.29, and 10.35 per cent manganese, respectively. Similar thicknesses and grades are indicated at Kelligrews Brook. At Brigus South Head on the west side of Conception Bay, a bulk sample from the central 1.5 feet of a manganiferous zone, 15 feet thick in steeply dipping Cambrian beds, carried 33.35 per cent manganese, 3.80 per cent iron, and 18.69 per cent silica. According to Johnston and McCartney (1965) this untreated material yielded silico-manganese in a single-stage electric smelting test, and in other tests was concentrated

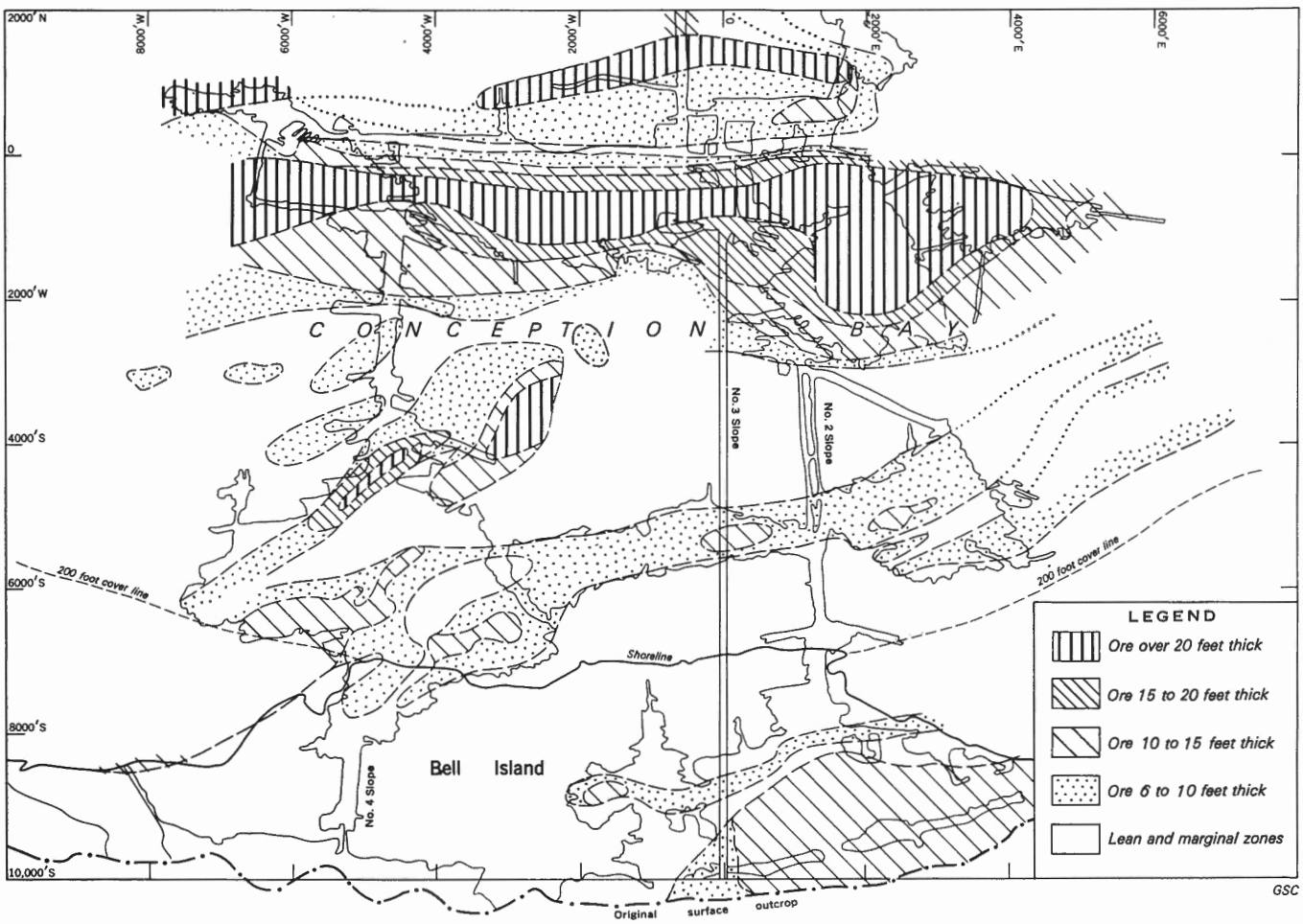


FIGURE VII-16. Ore trends and thickness of the Lower Bed, Wabana iron mines, Newfoundland (Lyons, 1957).

by a sink–float process to yield a roasted sink product containing 57.06 per cent manganese.

Molybdenum

As mentioned in the section on Copper, molybdenite is a byproduct of the Needle Mountain and Copper Mountain deposits at Gaspé, Quebec. In 1966, 531,598 pounds of molybdenum valued at $832,976 was recovered from these ores.

Numerous occurrences of molybdenite in quartz veins, aplite, and pegmatite, associated with granitic intrusions, are known at Gabarus Bay on Cape Breton Island, and near Jordan Falls, Nova Scotia. Small, high-grade molybdenite occurrences are reported in serpentine and gneiss intruded by granite pegmatite near Fleur de Lys, northern Newfoundland. Disseminations of molybdenite occur within a granitic intrusion and along its contact with the Belle Bay volcanics at Rencontre Lake near Fortune Bay, Newfoundland. The disseminated pyrite and molybdenite, and molybdenite in quartz veinlets, occupy a large, almost circular area in a fluorite-bearing alaskite granite, the Ackley granite, of probable Devonian age. Small occurrences are known at Lawn on the Burin Peninsula.

Niobium (Columbium)

The first and only producer of niobium in Canada is the deposit at *Oka* (49), Quebec. Pyrochlore, a dark brown mineral of complex composition carrying niobium, titanium, thorium, uranium, and rare earths, along with perovskite, magnetite, and some other minerals with radioactive and rare elements, are disseminated through a complex, multiple, circular stock of alkaline syenite that penetrates a Precambrian inlier of anorthosite, granite, gneiss, and crystalline limestone as well as the marginal remnants of brecciated Paleozoic limestone around it (Rowe, 1958). The niobium-bearing syenite is believed to be Triassic and is probably related to the Monteregian Intrusions that penetrate Paleozoic sediments in the St. Lawrence and Appalachian Provinces near by. As suggested in Chapter V it may be distantly related to a suite of niobium-bearing Precambrian alkaline intrusions that occur in the Canadian Shield. In 1966, 2,637,997 pounds of niobium oxide valued at $3,182,170 were contained in pyrochlore concentrates recovered from the deposit.

Tin

Tin is a minor constituent of the base-metal ores of the Bathurst area in New Brunswick and of the pegmatites and greisen zones in a muscovite-bearing leuco-granite near New Ross, Nova Scotia. At the Reeves mine in Lunenburg county, Nova Scotia, a pegmatite dyke in granite carries cassiterite, amblygonite, lepidolite, topaz, tourmaline, fluorite, monazite, and columbite–tantalite, rare beryl, and quartz crystals as much as 3 feet long.

At *Pleasant Mountain* north of St. George, New Brunswick, a low-grade tin deposit occurs in fractures in a complex of altered volcanic, pyroclastic, and metasedimentary rocks that are mainly Mississippian. The volcanic rocks at the mine are silicified and chloritized, in places replaced by fluorite, topaz, sericite, and kaolin, and impregnated with varying amounts of cassiterite, sphalerite, arsenopyrite, pyrite, chalcopyrite, galena, stannite, molybdenite, wolframite, and other minerals (Mulligan, 1966). Two zones of potential ore 4,000 feet apart, controlled by a complex fracture system, lie in the general line of several small granite stocks that outcrop in a belt north of the main batholith of Devonian(?) granite. The ore is reported to carry 0.6 per cent tin, 2.3 per cent zinc, 0.30 per cent copper, and 0.36 per cent lead.

Titanium

Titaniferous magnetite in anorthositic rocks of the Indian Head and Steel Mountain areas of Newfoundland, and elsewhere in the Appalachian Mineral Province, form disseminations and dyke-like bodies of iron and titanium oxide minerals, mainly vanadiferous titanomagnetite and ilmenite. Most occurrences have not been developed. Low-grade deposits of titaniferous magnetite may be concentrated magnetically and pelletized. Pelletized iron ore now commands premium prices in the steel industry, and recovery of the byproducts of titanium and vanadium, as well as iron, may also be feasible.

Tungsten

At *Burnt Hill,* 35 miles northwest of Fredericton, New Brunswick, Devonian granite intrudes thick Silurian beds of grey argillite and quartzite that are altered to hornfels over wide areas. The metasediments are also cut by basic dykes, aplite dykes, and greisenized granites that grade into quartz veins. Some greisens and quartz veins contain wolframite, molybdenite, and cassiterite. Numerous vertical quartz fissure veins, as much as 6 feet wide and 100 feet long, cut across the beds and carry pyrrhotite, chalcopyrite, pyrite, fluorite, chalcocite, locally abundant clusters of wolframite, sparse cassiterite and molybdenite, and abundant topaz. The deposit was opened in 1916 and about 200 tons of ore was milled (Little, 1959). The deposit was reopened in 1939 and 1953, but only limited production has ensued. The Square Lake occurrence, 20 miles northwest of Saint John, also consists of zones of wolframite–molybdenite–topaz-bearing greisen zones and quartz veins in a Devonian biotite granite. The quartz veins carry bismuthinite and native bismuth as

well, and crystals of wolframite as much as a half inch or more long.

In Nova Scotia, numerous occurrences of scheelite, cassiterite, and wolframite have been found in quartz veins, gold–quartz–antimony veins, and greisenized zones associated with the emplacement of the central Devonian granitic batholith in Lunenburg and Halifax counties. Wolframite and chalcopyrite occur in quartz veins in the Grey River area of the south coast of Newfoundland. Scheelite and tungstite were discovered in a large quartz vein near Charles Cove on the west side of Gander Bay, Newfoundland. The quartz vein, from 2 to 15 feet wide, intersects the contact of granodiorite and Silurian(?) slates and has been traced for a length of three-quarters of a mile.

Chromium

Chromite deposits are confined to the belts of serpentinized ultrabasic rocks that contain the asbestos and nickel occurrences in the Eastern Townships of Quebec, the Mount Albert intrusion in the Gaspé Peninsula, the Bay of Islands and Hare Bay Complex in western Newfoundland, and two belts in northern and central Newfoundland. There has been sporadic production from a number of these deposits, mainly during times of war, but none have produced since 1947. Most of the higher grade occurrences are irregular pockets, but a few are tabular. The orebodies consist of disseminated chromite and segregations or veins of massive chromite, mainly in dunite and peridotite that form lower layers of partly differentiated ultrabasic rocks (Stockwell, 1944). Some deposits, such as the *Reed-Belanger (Chromeraine)* deposit in Coleraine township, are as much as 500 feet long, 8 to 60 feet wide, and more than 300 feet deep. Workable ore consists of a series of irregular lenses in a zone 1,500 feet long and 100 feet wide. Some of the ore mined was as low as 10 per cent Cr_2O_3. Shipments averaged about 35 per cent Cr_2O_3, with a chromium to iron ratio of 2.6:1. Small tonnages of chromite with a 3:1 ratio have been produced from the deformed serpentinites of the Eastern Townships, but the known deposits with this ratio are too small to compete with foreign sources.

Chromite occurrences in the deformed serpentinite belts of central Newfoundland show a ratio slightly higher than 3:1, but these deposits are rather small. Thin layers and bands of chromite are found in the relatively undeformed, stratiform Bay of Islands Complex. The chromite is concentrated in the upper horizons of the ultrabasic rocks immediately below the overlying gabbro (Smith, 1958). The Stowbridge deposit is the largest known concentration in the North Arm Mountain pluton. The chromite is reported to be a magnesium-rich variety, carrying about 40 per cent Cr_2O_3 and having a maximum chromium to iron ratio of 1.70:1.

Vanadium

Vanadium is a common trace element in deposits of titaniferous magnetite, oölitic hematite, chromite, copper, and uranium-bearing sandstones of the red bed type. It has been found in the titaniferous magnetite of St. George's Bay in Newfoundland in amounts ranging from 0.02 to 1.1 per cent; in Wabana hematite from 0.007 to 0.093 per cent; in Coleraine chromite at 0.2 per cent; and in red Carboniferous sandstone beds north of Fredericton at 0.04 per cent (Rose, 1967). Vanadium is also commonly associated with dark shales, lignite, coal, and hydrocarbons, and has been detected in oil shale from Newfoundland and New Brunswick, from soft coal at New Glasgow, Nova Scotia, and from the albertite at Albert mines, New Brunswick.

INDUSTRIAL MINERAL DEPOSITS

The industrial mineral output of the Appalachian Mineral Province includes asbestos, salt, gypsum, fluorite, barite, soapstone, talc, and pyrophyllite. The value of asbestos alone exceeds that of the metals produced in 1966. Exclusive of asbestos, however, the value of the industrial mineral production of the region is but a third of that of the metals. This is true also for Newfoundland, New Brunswick, and Quebec, but not for Nova Scotia where the industrial mineral output exceeds that of the metals. In the St. Lawrence Mineral Province bedded sedimentary deposits of salt, gypsum–anhydrite, limestone, dolomite, quartz sandstone, and shale are produced, as well as structural materials such as sand, gravel, crushed stone, and building stone (Table VII-1).

Asbestos

Chrysotile asbestos was first mined near Thetford, Quebec, in 1876. Since that time many large deposits of asbestos have been found within the elongated, sill-like, differentiated intrusions of serpentinized ultrabasic rock that spread through the Eastern Townships and Gaspé Peninsula into northern Newfoundland. The most productive part of the belt is 55 miles long, extending from Asbestos to Broughton township. Nine mines are currently active, and thirteen deposits have been mined out. Production from Quebec in 1966 was 1,336,566 tons valued at $141,559,725. Important deposits were developed in 1963 near Baie Verte in Newfoundland. In

1966, this mine produced 64,850 tons of fibre valued at $10,300,000.

Eastern Townships, Quebec. Serpentinized peridotite and pyroxenite occur as sheet-like stocks, dykes, and sills in what is known as the serpentine belt in Quebec. The belt actually extends along almost the entire length of the Appalachian fold belt in North America. The ultrabasic rocks invade, for the most part concordantly, stratified metasedimentary and metavolcanic host rocks that range from Cambrian to Ordovician. The emplacement of the ultrabasic intrusions at Mount Albert has been dated at 495 m.y. (Lower Ordovician), but they are generally believed to have been emplaced during either the Taconian or Acadian Orogeny. They are intruded in many places by granitic bodies of probable Devonian age. The formation of asbestos has been ascribed by Hess to autometamorphism within the cooling ultrabasic magma. The asbestos in the Eastern Townships is regarded as a product of later orogeny and invasion by granite (Cooke, 1937; Riordon, *in* Convey, *et al.*, 1957).

Two types of asbestos are common—cross-fibre and slip-fibre chrysotile. The former occurs in veins as wide as 4 inches, but most production is from veins less than ⅜ inch wide. The ore is crushed and screened and the asbestos fibre is lifted off by suction. With the split-fibre veins a separation may occur in the central part of the vein where grains of serpentinized peridotite, magnetite, and chromite may be found. Ribbon veins fill parallel fractures in serpentinized peridotite and in places these form parts of larger, composite veins. In a few places asbestos veins grade into picrolite (harsh fibre) veins.

Extensive zones of soapstone and talc–carbonate rocks occur near or along the contacts of the serpentinized ultrabasic rocks. Magnetite, in places titaniferous, and chromite, in both disseminated and massive, segregated form are also commonly associated with these rocks. Nickeliferous pyrrhotite and nickeliferous pyrite also appear to be related to the ultrabasic rocks.

During the past decade many of the producing mines have been consolidated as pits converged, and much of the production during these years has come from the following mines: *Carey-Canadian Mines Limited* (64): a company formed in 1955 to mine a series of four chrysotile deposits located about 20 miles northeast of Thetford Mines. Open-pit mining began in 1958. About 4,000 tons per day are milled. *King–Bell–Johnson–Beaver mine* (61d): a series of deposits extending over 7,000 feet in length, 2,500 feet in width, and 1,300 feet in depth, worked from the earliest days of asbestos mining in the area, is being mined by block-caving methods underground and by open pit (Figs. VII-17, 18). The mill processes up to 8,000 tons of ore daily. *The British Canadian mine* (61a): the British Canadian mine (Fig. VII-19) located in Black Lake, opened in 1885, has now been merged with the Megantic mine of Johnson's Company Limited. From the pit, which is over a mile long and half a mile wide, about 40,000 tons is mined daily. *Normandie mine* (60b): this open-cut mine, located 4 miles west of Black Lake, was opened in 1955; its annual production is about 85,000 tons of asbestos. *Nicolet Asbestos mines* (59): an open-cut asbestos mine opened recently at Norbestos, in Tingwick township, about 10 miles northeast of Asbestos. The Nicolet cross-fibre chrysotile deposit is 3,300 feet long and 500 feet wide. It has been developed to a depth of 100 feet. *National Asbestos mines* (62a): an open-pit asbestos mine 4 miles east of Thetford Mines was brought into production in 1958. It is producing at the rate of 3,600 tons daily. *Flintkote Mines Limited* (62b): this company began an open-pit asbestos operation 3 miles east of Thetford Mines in 1946. About 2,000 tons is mined daily. The deposit is within a band of serpentinized peridotite known as the Pennington dyke. *Lake Asbestos mine* (61e): an asbestos deposit beneath the waters of Black Lake was prepared for mining during the period

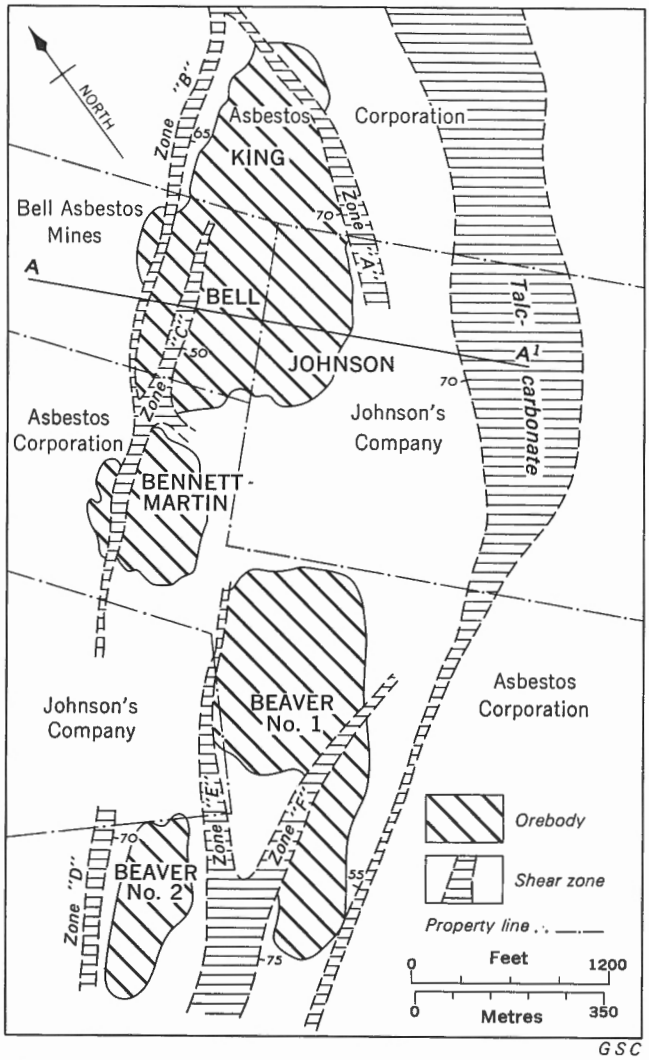


FIGURE VII-17. Asbestos deposits at Thetford Mines, Quebec, showing relationships with shear zones and talc–carbonate, or Thetford fault zone (Riordon, 1957).

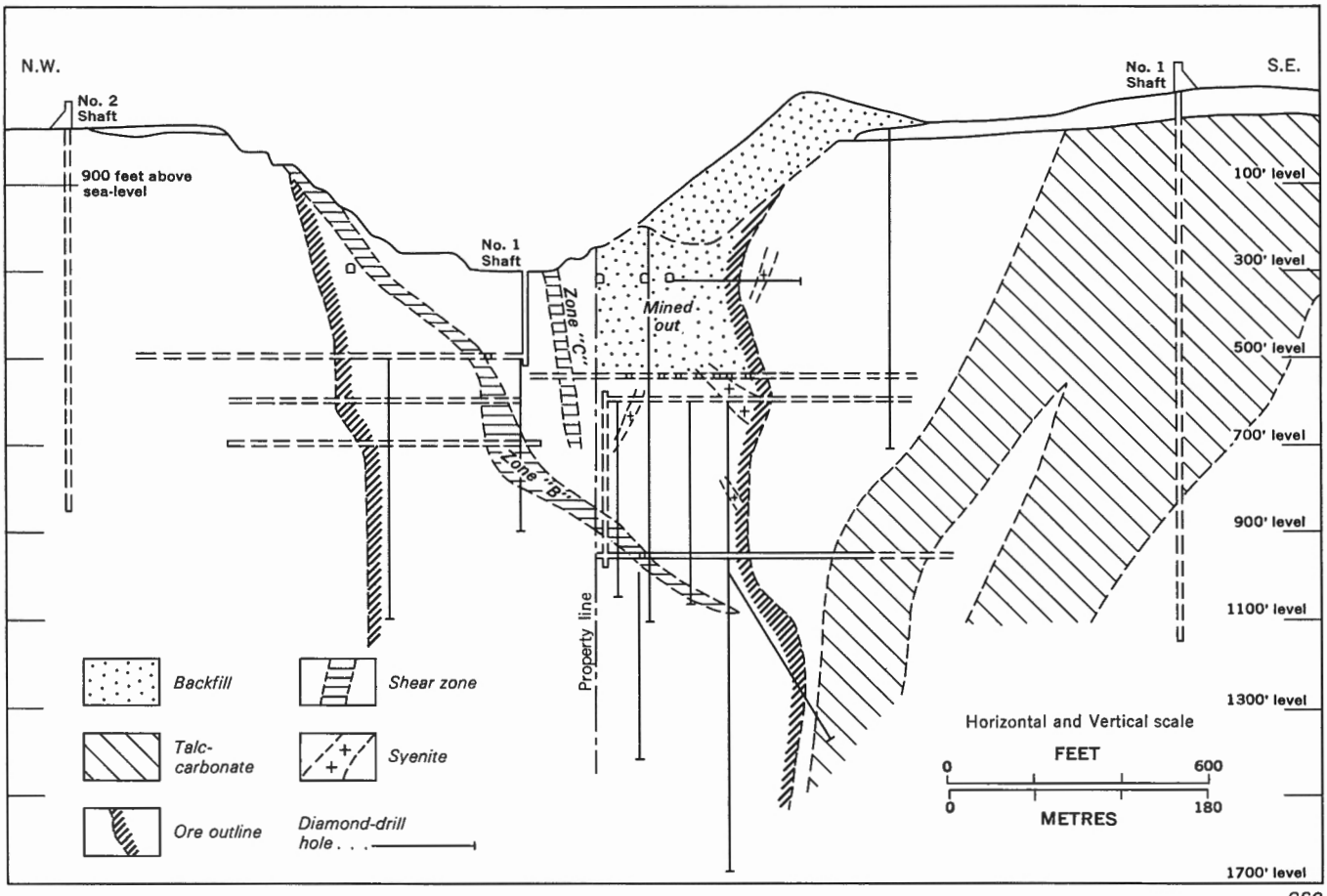


FIGURE VII-18. Cross-section through the Bell–Johnson deposit, Thetford Mines, Quebec; along line A–A[1] of Figure VII-17 (Riordon, 1957).

1955–58, by draining most of the lake and diverting the Bécancour River. About 31 million cubic yards of alluvium was dredged from a 483-acre area of the lake bottom. Production capacity of the mill is rated at 125,000 tons of asbestos per year.

The *Jeffrey mine* (58) at Asbestos, Quebec, is one of the largest open-cut mines in Canada. In 1966 this mine produced 610,305 tons of fibre, 45 per cent of the country's total. The open pit measures 3,500 feet by 3,800 feet, and is 590 feet deep. About 32,000 tons of "ore" and 16,000 tons of waste are mined per day. The pit is being extended to increase annual production by 1971 to more than 10 million tons. Black slates, interbedded with impure quartzite and quartz–actinolite schist, form the north or footwall side of a sill 4,800 feet thick consisting of serpentinized peridotite, dunite, and pyroxenite. The beds, believed to be Cambrian or younger, strike N60°E and dip 65° southeast. Several feet beyond the contact the slate is crumpled and altered to phyllite. Andesitic lava and volcanic (explosion) breccias form the south or hanging-wall side of the sill. Gabbro intrudes pyroxenite along the hanging-wall contact. Small granitic intrusions occur within the deposits. The most favourable host rock for asbestos is the massive peridotite. The deposit forms an elliptical cylinder raking 60° south.

Newfoundland. Ultrabasic intrusions of the same type as those in Quebec occur in two belts in central Newfoundland. They have long been known to carry serpentine, chromite, and talc–carbonate, but it was not until 1963 that commercial deposits of asbestos were located. At *Advocate mine* (4) near Baie Verte, asbestos-bearing rock forms a mass about 1 mile long and ¼ mile wide within a sill-like body of highly serpentinized peridotite that intrudes chloritic schists of the Ordovician Baie Verte Formation. In 1966, 65,201 tons of fibre was produced from 2,006,735 tons of ore. The stratiform, serpentinized, ultrabasic intrusions of the Bay of Islands Igneous Complex of western Newfoundland also carry small deposits of asbestos, and similar intrusions at Hare Bay on the Northern Peninsula have narrow veins of asbestos. Six occurrences are known in the eastern (Gander River) belt.

Barite

Walton, Nova Scotia (6). This great pipe-like barite deposit, one of the largest in the world, yields most of Canada's barite. The Magnet Cove deposit has been described previously in the section on lead–zinc–copper; it is an argentiferous massive sulphide orebody that was

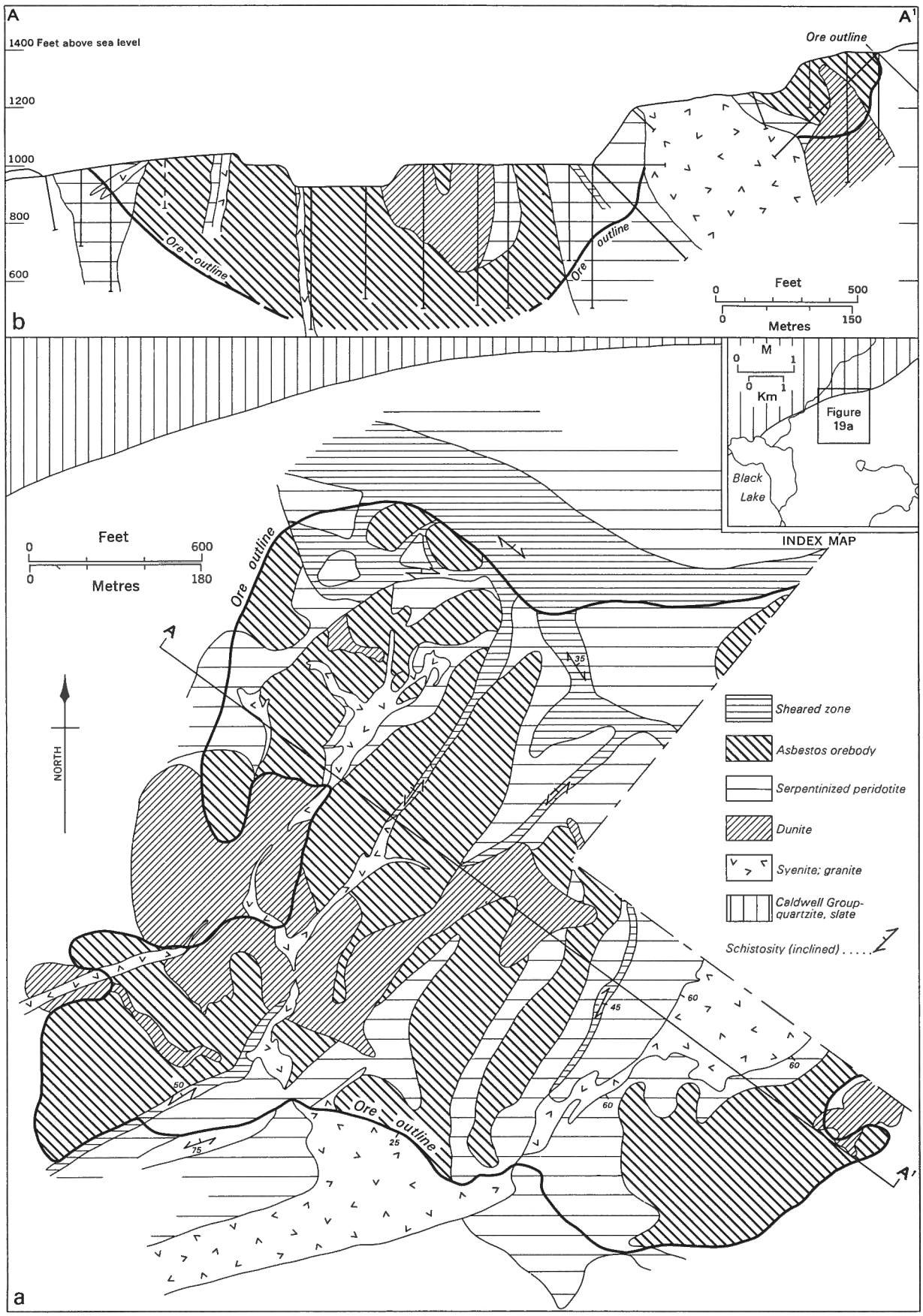


FIGURE VII-19. British Canadian orebody, Black Lake, Quebec (Riordon, 1957). (a) Geology. (b) Cross-section along line A–A[1].

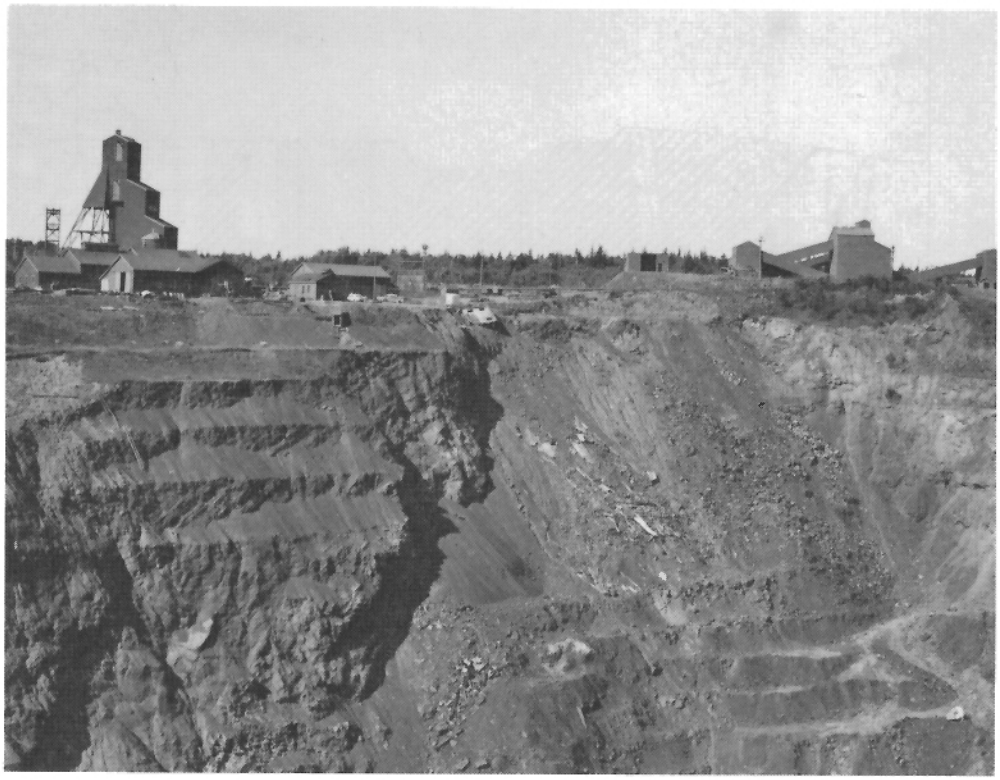


PLATE VII-4

Magnet Cove Barium Corporation, Walton, Nova Scotia. Open pit is in the barite deposit and headframe covers shaft leading to underlying lead–zinc–silver orebody.

found beneath the barite (Pl. VII-4). More than 4 million tons of barite valued at about $50 million has been produced to the end of 1966. The orebody, 90 per cent barium sulphate ($BaSO_4$), appears to be a replacement body localized by a fold and intersecting faults in the gypsum and limestone-bearing sediments of the Windsor Group. Triassic trap and diabase intrusions occur within 15 miles of the deposit. Barite deposits in the United States from Missouri to New York are believed to be associated with deep faulting and folding accompanying late Triassic igneous activity, and this may also be true at Walton. Boyle and Jambor (1966) suggested that the enclosing sedimentary rocks are the source of the elements in the deposits at Walton.

Fluorite

The *Director mine* (15b) at St. Lawrence, Newfoundland, is the only fluorspar mine currently in operation in the district (Fig. VII-20). More than forty fluorite-bearing veins are known, but only a few have been mined. The fluorite occurs as fracture fillings within alaskite granite and rhyolite porphyry of probable Devonian age that have intruded Ordovician and older volcanic and sedimentary rocks. The Director vein has been traced for more than 6,000 feet, mainly in the granite and at the contact between granite and wall-rock. Three ore lenses occur along the vein; the main lens is 90 feet wide in one place, but is generally less than 50 feet wide and more than 1,000 feet deep. Fluorite forms most of the vein, along with some quartz, calcite, barite, pyrite, chalcopyrite, galena, and sphalerite. It also forms stringers or cements breccia adjacent to the veins. The veins are banded parallel to their walls (Pl. VII-5), and locally the fluorite forms concentrically banded nodules around granite fragments. Locally, comminuted fluorite is cemented by silica and is called "blastonite", a material undesirably high in silica.

Steatite, Talc, and Pyrophyllite

Soapstone (steatite) and talc, the hydrous silicates of magnesium, occur in serpentinized peridotite and associated zones of talc–carbonate rock in the serpentine belt of the Appalachian Mineral Province. Some of these occurrences in the Eastern Townships of Quebec were worked as early as 1871. Products of the *Baker Talc* mines (50, 51) and *Broughton Soapstone and Quarry* (63a) include soapstone blocks, talc crayons, and pulverized talc or soapstone for use as dry carriers in insecticides, fillers in paint, paper, rubber, linoleum, tile, soap, plastics, and cosmetics.

Pyrophyllite, the hydrous aluminum silicate resembling talc, has been found at *Manuels* (17), Newfoundland. It occurs along with quartz in irregular silicified zones in the Harbour Main volcanic rocks near the intrusive Holyrood granite contact. The deposits were opened in 1904, sporadically worked, and reopened in 1956 to supply markets in the ceramics industry.

Salt

Salt deposits occur in the Upper Silurian of southwestern Ontario, and are known in Mississippian strata in various parts of Nova Scotia, New Brunswick, and Prince Edward Island. Salt springs have also been noted at the margin of Mississippian basins in western Newfoundland, Nova Scotia, and New Brunswick. Recovery operations currently conducted in Ontario and Nova Scotia are by artificial brine and conventional mining methods. The former method has been in use for more than 100 years in Ontario. Wells are drilled into the salt beds, cased, then tubing of smaller diameter placed inside the casing. Fresh water forced into the well through the outer tube dissolves rock salt to form a brine which is pumped to surface through the inner tube and then the salt is removed by evaporation.

The largest consumers of salt are the chemical industries in the manufacture of soda ash and chlorine. Other major uses are the food processing industries such as meat packing, fish processing, baking, canning, brewing, and the manufacture of confectionary and dairy products. Large quantities are also used for tanning and in the metallurgical and oil refining industries; it is also being used during the winter months in ever increasing quantities for de-icing streets and highways.

Ontario. Salt was discovered in Ontario in 1866 when a company was formed at Goderich for the purpose of drilling for oil. The first well was drilled on the north side of the Maitland River at Goderich, but instead of striking oil, a bed of rock salt was encountered at a depth of 964 feet, and thus a new industry was initiated. Drilling for salt quickly spread to neighbouring centres, and a

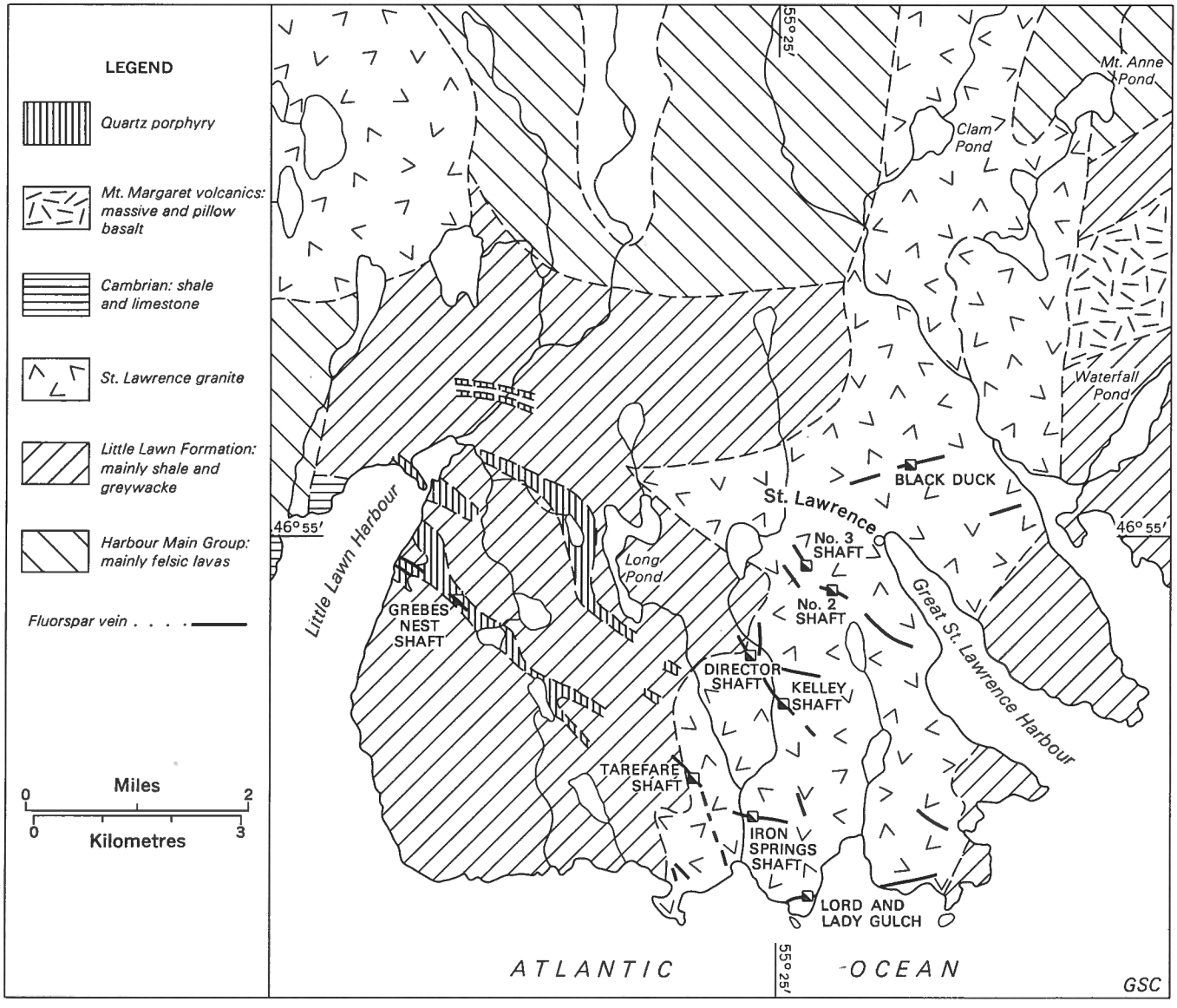


FIGURE VII-20. Geology of the St. Lawrence fluorspar area, Newfoundland (van Alstine, 1948).



PLATE VII-5. Banded fluorite vein, Director mine, Newfoundland Fluorspar Limited, St. Lawrence, Newfoundland. Vein is a fissure filling in Devonian alaskite granite.

number of artificial brine fields came into production throughout the western part of the Ontario Peninsula. Brine fields in operation in 1966 were at *Goderich* (116), *Sarnia* (117), *Windsor* (118), and *Amherstburg* (119a). The brine is taken from the B, D, and F salt zones of the Upper Silurian Salina Formation (Fig. VII-21). In recent years, salt mines at *Ojibway* (119b) and *Goderich* have come into production in Ontario. The Ojibway (Pl. VII-6) produces from one of the F unit salts at a depth of 948 feet. The mine at Goderich is producing from an 80-foot salt zone in the A-2 unit at a depth of 1,675 feet. These operations have greatly increased the annual salt production in Ontario, and have added considerably to the versatility of the industry. Although salt has been produced continually since its discovery, production figures are only reliable for the years 1895 to 1966 inclusive when 48,560,299 tons valued at $241,578,079 was recovered. Production during 1966 was 3,782,191 tons valued at $15,243,791.

Nova Scotia. For many years, the *Malagash* rock salt mine on Northumberland Strait was the chief source of salt in Nova Scotia, but it has gradually become exhausted and a new property has been developed at *Pugwash* (2). In addition, production is obtained from an artificial brine field at *Nappan* (1). Drillholes in the Malagash, Pugwash, and Nappan areas have penetrated salts as thick as 1,500 feet at depths of from 85 to 3,500 feet. These deposits are Mississippian and occur in the Windsor Group. Salt production in Nova Scotia during 1966 was 474,981 tons valued at $4,724,993.

Gypsum

Large deposits of gypsum occur in Ontario, Newfoundland, New Brunswick, and Nova Scotia. During the early development of the industry gypsum was utilized solely for fertilizer and the market was local. It now has many other uses, principally in the construction industry for the manufacture of plaster of paris, gypsum lath, and wall board, and is also used as a control in the setting of portland cement or as filler in paint and paper.

Ontario. Gypsum occurrences have been known in Ontario for 145 years; the first mine was opened about 1822 near the site of the present city of Paris. Since then, about 15 deposits, all located along the Grand River valley, have been worked intermittently. The most important occurrences are in Seneca, Oneida, and North Cayuga townships of Haldimand county. The deposits are lenticular, as much as 11 feet thick, and occur sporadically at various horizons interdigitating with dolomites and shales within the Upper Silurian Salina Formation.

In 1966 two mines were operating in Ontario, one at *Caledonia* (114) the other near *Hagersville* (115), both in Haldimand county. Their combined output during that year totalled 565,185 tons valued at $1,581,010.

Nova Scotia and New Brunswick. Gypsum occurs in central and eastern Nova Scotia, southern New Brunswick, and on the Magdalen Islands. All the deposits are in Mississippian beds of the Windsor Group. The main gypsum deposits of Nova Scotia are near *Windsor* (7) in Cape Breton Island. In New Brunswick production is confined to the *Hillsborough* (9) area of Albert county. In 1966, Nova Scotia produced 4,502,836 tons valued at $8,140,651, and New Brunswick produced 108,207 tons ($413,074).

Newfoundland. In Newfoundland, strata of the Mississippian Codroy Group underlie the lowland bordering St. George's Bay, and gypsum is found at a number of places, the largest producer being *Flat Bay* (14). Production in 1966 was 459,685 tons valued at $1,173,401.

Limestone and Dolomite

The relatively undisturbed Paleozoic strata of Southeastern Canada provide an adequate supply of limestone and dolomite for the manufacture of a wide variety of structural products. These include crushed stone for road construction, concrete aggregate, railway ballast, fluxstone, building stone, terrazzo, stucco dash, riprap (armourstone), and pulverized stone as an asphalt filler, cement, and chemical lime. Although more commonly used in the

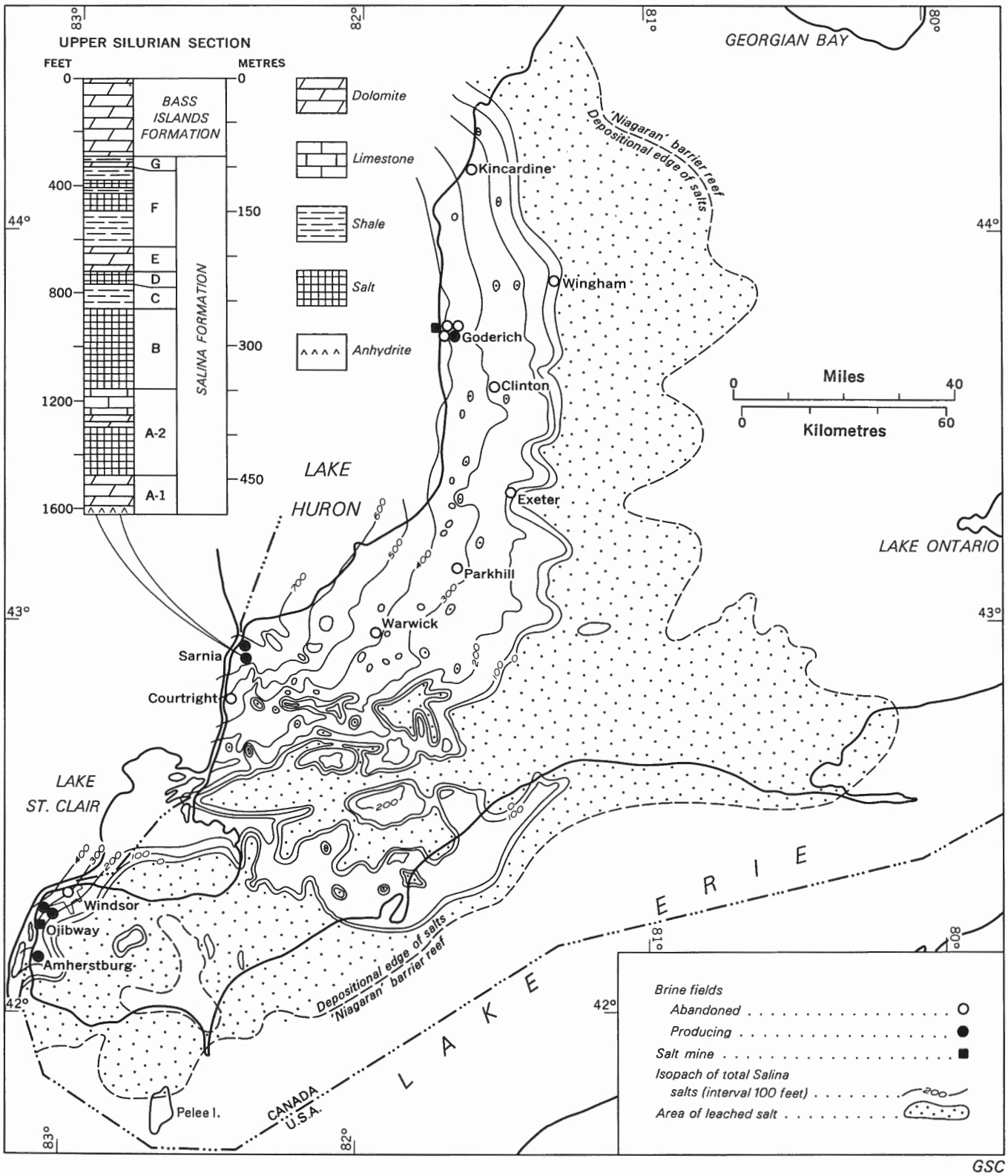


FIGURE VII-21. Salt mines and brine fields in the Silurian Salina Formation of southwestern Ontario (Sanford, 1965).

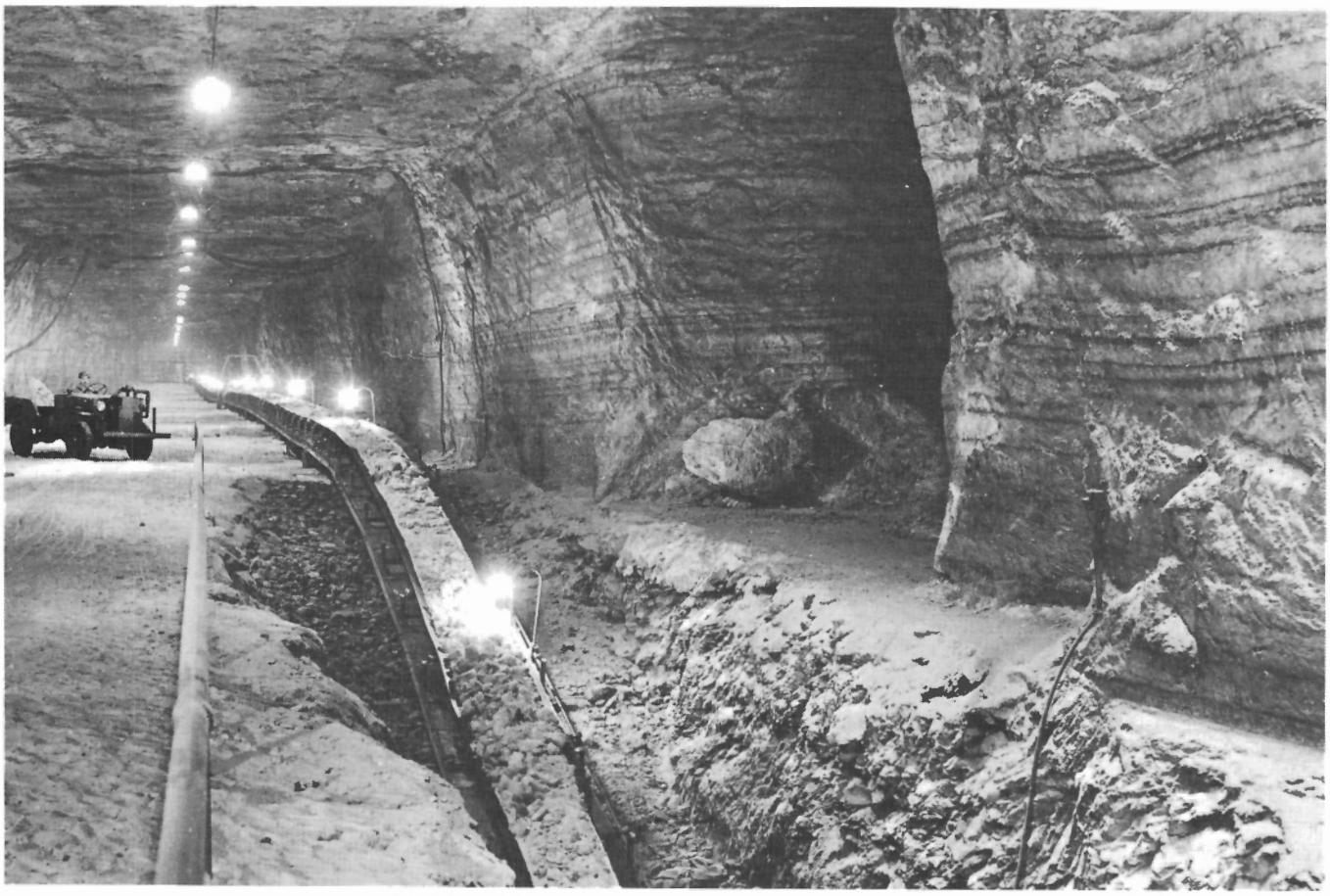


PLATE VII-6. Salt mine of Canadian Rock Salt Company Limited at Ojibway, Ontario, in F unit of the Upper Silurian, Salina Formation.

building trades, chemical lime is an important ingredient in the manufacture of iron and steel, uranium processing, pulp and paper, glass making, gold milling, refining of sugar, smelting, tanning, and as a soil conditioner.

Approximately 68.5 million tons valued at nearly $68 million was quarried in various parts of Ontario, Quebec, and western Newfoundland, and utilized for one or more of the foregoing products. More than half of this production is from the Ottawa–Montreal–Quebec region, and is derived from the Beekmantown, Chazy, Black River, and Trenton Groups, all of Ordovician age; most of the remaining production is quarried in southwestern Ontario. Many of the carbonate formations are utilized, particularly the Ordovician Black River and Trenton Groups; the Silurian Guelph, Lockport, and Amabel Groups; and the Devonian Dundee, Lucas, and Amherstburg Formations. Approximately 250,000 tons of limestone was quarried in western Newfoundland during 1966; production is from the Middle Ordovician Table Head Formation. Much of the limestone in Nova Scotia and New Brunswick is derived from the Mississippian Windsor Group. In 1966 the two provinces had a combined limestone production of about 1.5 million tons valued at $2,235,000.

Sandstone and Shale

The sandstones of Ontario and Quebec have long been utilized as building stone. In southwestern Ontario the Lower Silurian Whirlpool and the Lower Devonian Oriskany beds have been used most. The Upper Cambrian Nepean sandstone in the region of Ottawa and the Quebec Lowland is famous for its durability and pleasing qualities as a building and ornamental stone. The Parliament Buildings in Ottawa are constructed of this sandstone obtained from a local quarry. Because of their exceptional purity, the sandstones are also used in glass making, the manufacture of silicon carbide, the processing of alloys, for foundry purposes, as a filler in asbestos cement, and in various cleaners.

Brick and tile are the chief products manufactured from the shales of Ontario and Quebec, particularly the Upper Ordovician Meaford, Dundas, and Queenston strata below the Niagara Escarpment in Ontario, and equivalent beds in the Ottawa and St. Lawrence Valleys. In addition, the late Middle Devonian shales of the Hamilton Group in southwestern Ontario are excellent for the manufacture of brick and tile.

Petroleum and Natural Gas

Petroleum and natural gas are produced from the Cambrian, Ordovician, Silurian, and Devonian rocks of southwestern Ontario and from Mississippian strata in New Brunswick (Fig. VII-1). The fields in southwestern Ontario occur along Algonquin and Findlay Arches and the flanks of the adjoining Michigan and Allegheny Basins. Other sedimentary basins of St. Lawrence Platform—the Quebec Basin, Ottawa Embayment, and Anticosti Basin—contain similar, little deformed Paleozoic strata that rest on a basement of Precambrian crystalline rocks and are generally considered to be potentially productive of oil and gas. The Silurian and Devonian strata of Gaspé Trough have been explored from time to time without success. The strata were folded in the Devonian Acadian Orogeny and rest on Ordovician and Cambrian low-grade metamorphic rocks deformed in the Taconian Orogeny. The Maritime sedimentary basins of Carboniferous, Permian, and Triassic strata that rest on a basement of older Paleozoic and Late Proterozoic rocks presumably also underlie parts of the Gulf of St. Lawrence and Atlantic Continental Shelf. Currently, there is moderate exploration activity, mainly offshore.

In Ontario the oil industry began with the utilization of oil from seepages occurring in glacial deposits along Black Creek in Lambton county. The first wells that were drilled into bedrock formations at Oil Springs were highly successful and many flowing wells were soon completed. After the discoveries at Petrolia in 1861 and Bothwell in 1862, the output of oil steadily increased until a peak of 829,000 barrels was recorded in 1894. The annual yield then declined to 113,000 barrels in 1945, but recent discoveries have increased production to 1,323,781 barrels during 1966 valued at $4.5 million (Table VII-1).

In 1889, the natural gas industry was initiated in Ontario with discovery of the Kingsville field in Essex county. Since then numerous gas fields have been discovered and these now occur over a wide area extending from southern Lake Huron to Niagara River. Several fields also extend for considerable distance beneath the waters of Lake Erie. In 1966 a total 15,537,157 Mcf of gas was produced, having a wholesale value in excess of $5,896,662.

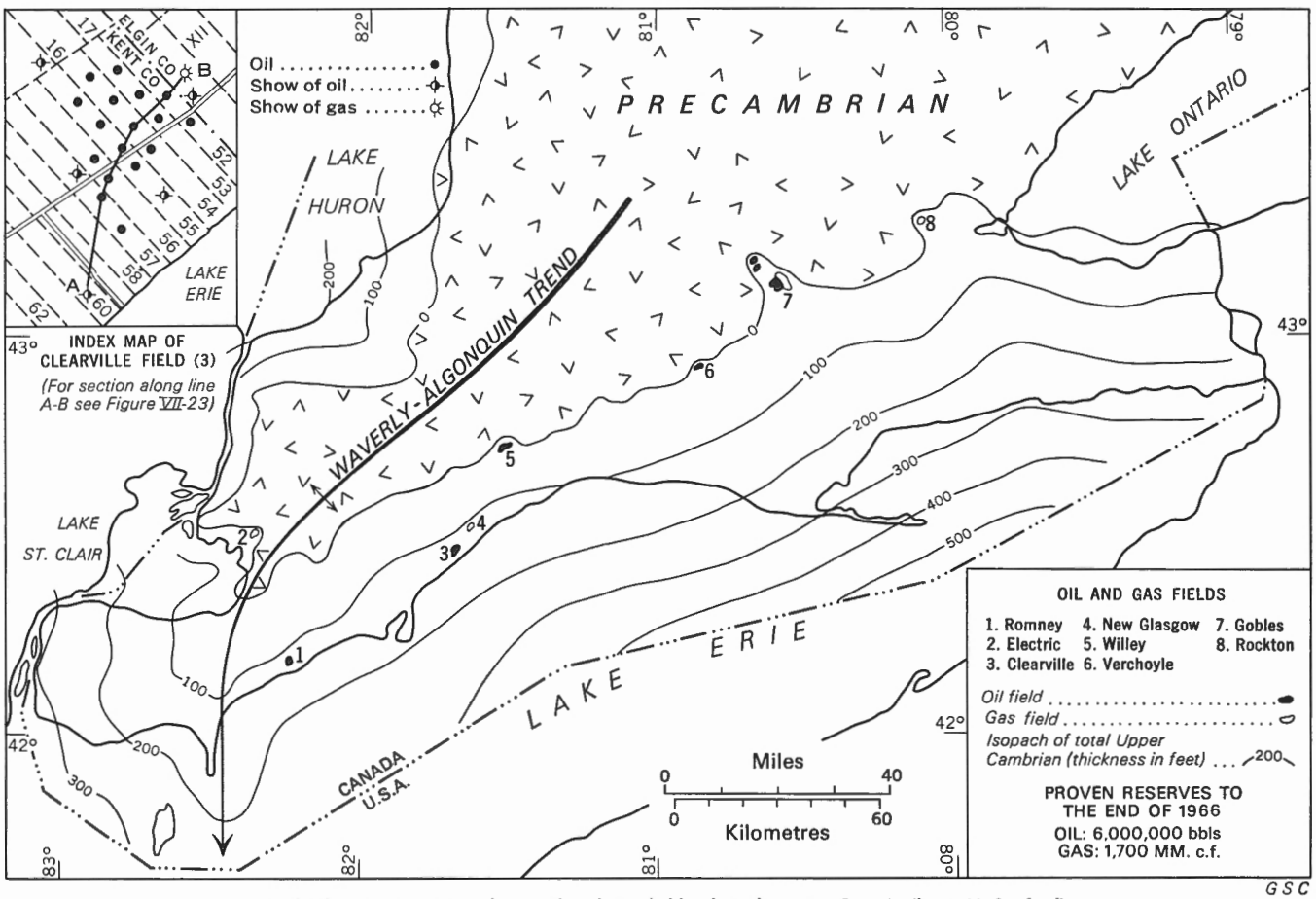


FIGURE VII-22. Upper Cambrian oil and gas fields of southwestern Ontario (by B. V. Sanford).

In New Brunswick drilling for oil commenced about the year 1858, but it was not until 1909 that the Stony Creek field was discovered near Moncton. During the peak years of 1913 and 1941 the field produced 856,156 Mcf of gas, and 33,359 barrels of oil, respectively. It has gradually declined to 97,403 Mcf of gas and 6,836 barrels of oil in 1966.

Southwestern Ontario Oil and Gas Fields

As of the end of 1966, a total of 58,400,000 barrels of oil and 858,150,000 Mcf of gas have been discovered. The first oil production was obtained from the Devonian and these rocks have continued to be the most important. Of lesser importance, although accounting for an appreciable percentage of the ultimate reserves, are rocks of Silurian and Cambrian age. Most gas is from the Silurian but some is produced from Cambrian and Ordovician rocks as well.

Cambrian Oil and Gas Fields

The Cambrian rocks of southwestern Ontario are potential oil and gas reservoirs throughout their entire area of distribution, but discoveries to date are confined to the southern flank of the Algonquin Arch bordering Allegheny Basin (Fig. VII-22). Deep drilling has delineated eight oil and gas fields with estimated ultimate reserves of 6 million barrels of 38° API gravity oil and 1.7 billion cubic feet of gas. The most favoured exploration target is the erosional edge of the sandstones and dolomites of the Potsdam and Theresa Formations, respectively, where the fields lie in embayments on the irregular surface of the Precambrian basement and are overlapped by Middle Ordovician limestones. Five accumulations have been discovered so far in such traps: the Electric, Willey, Verchoyle, Gobles, and Rockton fields. In addition, three fields, the Romney, Clearville, and New Glasgow, have been discovered downdip from the Cambrian edge in the interbedded dolomites and sandstones of the Theresa Formation. Marginal accumulations such as at Romney and New Glasgow occur in sandstone remnants on topographic highs on the Upper Cambrian erosion surface. Accumulation in the Clearville field is controlled largely by normal faults (Fig. VII-23).

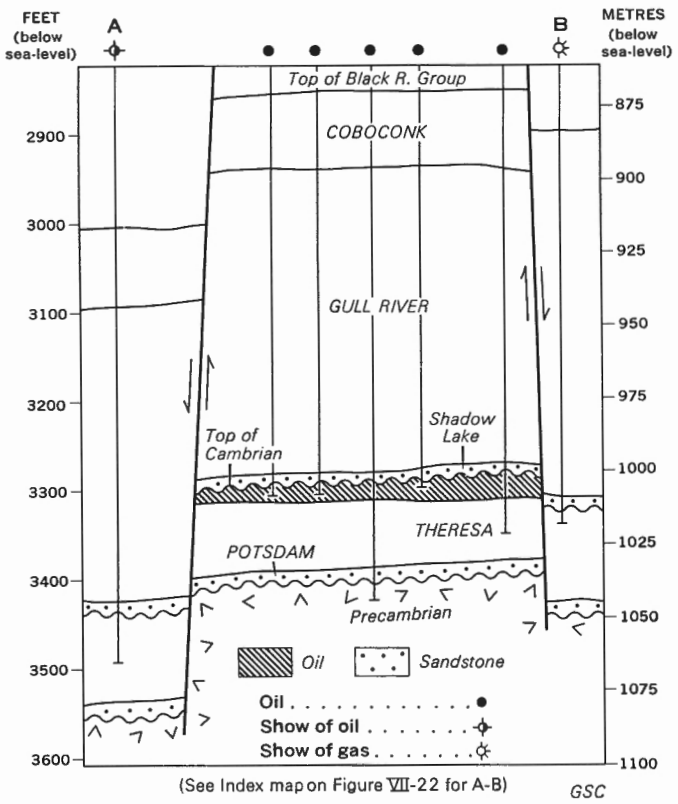


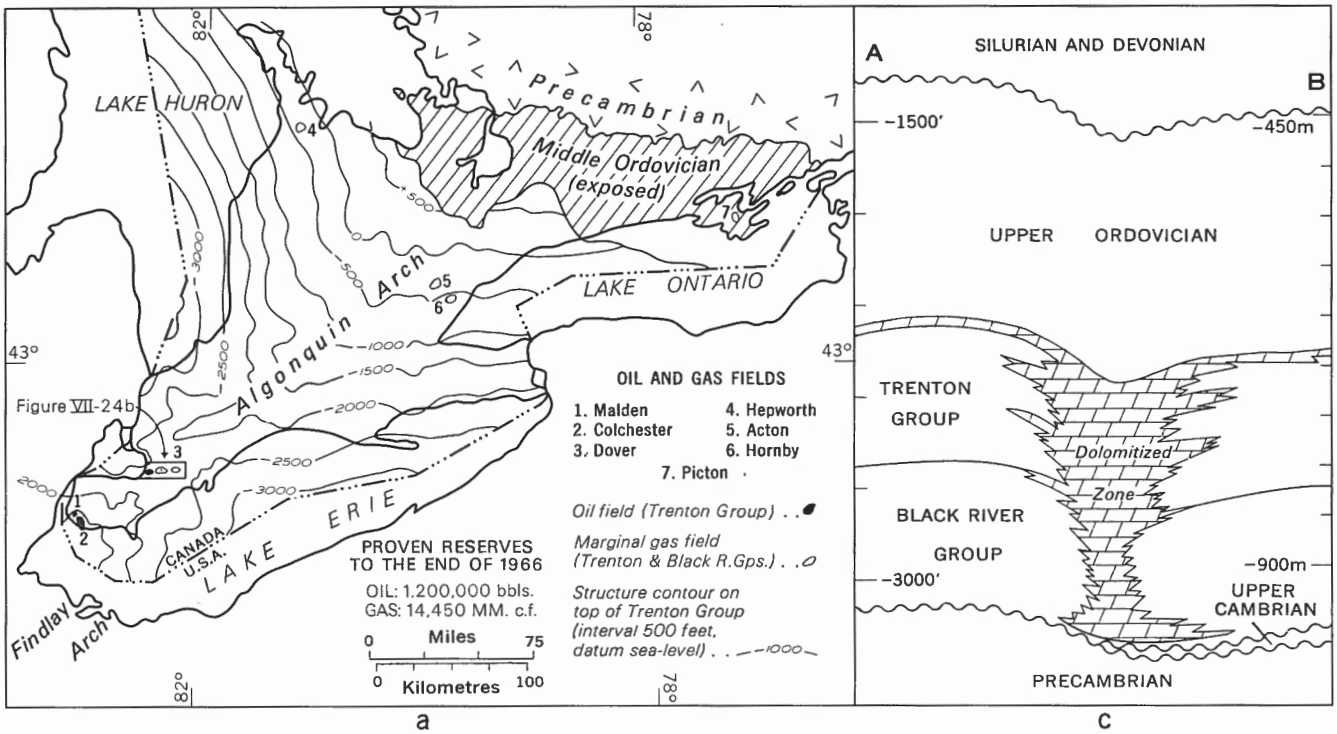
FIGURE VII-23. Upper Cambrian, Clearville oil field, Ontario (after W. E. Koepke).

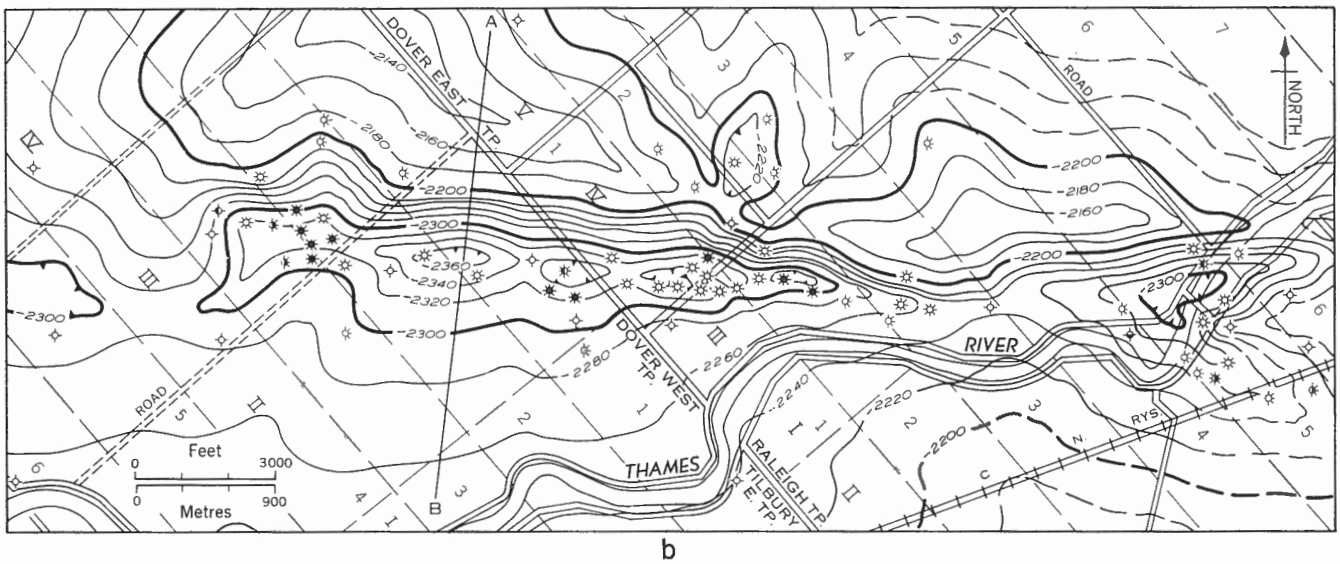
Oil was initially discovered in Cambrian rocks in 1923 at the Romney field, but it was the Gobles field, discovered in 1960, that provided the stimulus for the current deep-drilling program in Ontario. The Gobles field and subsequent discoveries such as Clearville had a combined output of 329,970 barrels of oil in 1966. This constituted 25 per cent of the total Ontario production during that year. In addition, 82,816 Mcf of gas was produced from the Gobles field.

Ordovician Oil and Gas Fields

Middle Ordovician carbonates of the Trenton and Black River Groups have long been regarded as potential oil and gas reservoirs in Ontario, but to date only three fields have been developed (Fig. VII-24a). These have an ultimate yield of 1.2 million barrels of 39° API gravity oil, and 14,450,000 Mcf of gas. Small quantities of oil and gas have been encountered in Middle Ordovician limestones on Manitoulin Island, and in Bruce, Halton, and Picton counties.

The Malden, Colchester, and Dover oil fields are unique structurally in that oil accumulation has taken place along the axis of an elongated syncline. The Dover oil and gas field discovered in 1917 is typical (Fig. VII-24b, c). The structure contours show the configuration of the top of the Trenton Group, and the typical elongated depression that coincides with the dolomitized sequence that extends the length of the field. In the structural depression the Black River and Trenton limestones are intensively dolomitized, a feature that has resulted from mass migration of magnesium-bearing solutions upward through fracture zones in the limestones and thence laterally along the base of the overlying Collingwood shale. The synclinal structure, although partly attributable to faulting, is due primarily to subsidence following the leaching effects of the interstitial moving water. The accumulation of the oil and gas, which coincides with the synclinal structure, is controlled by the porosity development in the dolomitized zones; these are considerably more developed in the bioclastic units of the Trenton Group, particularly the Sherman Fall Formation. To date all the Trenton production in Ontario is confined to dolomitized fracture systems that were induced by uplift





Well, oil and gas producer

Well, gas producer

Well, gas producer with show of oil

Well, with show of gas

Well, with show of oil

Well, with show of gas and show of oil

Well, non-productive

Structure contour on Cobourg Formation (interval 20 feet, datum sea-level)

Lot and concession number. . . . . 4, III

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FIGURE VII-24. Middle Ordovician oil and gas fields of southwestern Ontario (by B. V. Sanford). (a) Distribution. (b) Structure contours on Middle Ordovician Trenton limestones in the Dover field. (c) Cross-section along line A–B through Dover field.

of the Findlay Arch. Consequently those areas of western Lake Erie, Essex, and Kent counties that are in close proximity to this arch are most favoured for future exploration.

During 1966, 42,008 barrels of oil and 34,263 Mcf of gas were recovered from Middle Ordovician strata, representing 3 per cent and 0.3 per cent, respectively, of the total annual yield.

Silurian Oil and Gas Fields

More than sixty Silurian oil and gas fields have so far been discovered in Ontario (Figs. VII-25, VII-29), and these will ultimately yield 842 billion cubic feet of gas and 8.7 million barrels of 36° API gravity oil. The most important reservoirs are the Guelph bioherm reefs that have a wide distribution along the eastern margin of Michigan Basin (Fig. VII-26). The largest and most productive of these are patch reefs such as those at the Kingsville, Mersea-Leamington, Tilbury, and D'Clute fields. They occur along the inner margins of the Guelph barrier reef complex. The largest of this group, the Tilbury reef (Fig. VII-27), was discovered in 1906, and accounts for nearly half of all the gas produced in Ontario from all sources. Although structural closure on the field is only about 160 feet, the reef is productive throughout an area of 38,000 acres, nearly half of which underlies Lake Erie.

In the lower part of Michigan Basin, pinnacle reefs account for a substantial percentage of the rest of the Ontario oil and gas reserves. These reservoirs have a gross pay section of as much as 300 feet, and are productive over areas varying from 300 to 500 acres. Of the forty-three pinnacle reefs so far discovered, some twenty-five are productive, most of which are located in Lambton county. Of these, twenty contain commercial volumes of gas, three contain oil, and two contain both oil and gas. The non-productive reefs are either filled with salt water or have no effective porosity, because they are plugged by salt and anhydrite. A typical reef is the Dawn 156 field (Fig. VII-28); discovered in 1956, it will ultimately yield a total of 26 billion cubic feet of gas. Also within the pinnacle reef belt are a number of low relief biohermal structures that generally have less than 100 feet of structural closure. Several of these contain commercial volumes of oil and gas, but most have proven to be non-productive.

Closely associated with the Guelph bioherm reefs are the oil and gas fields in the carbonates of the A-1 and A-2 units of the Upper Silurian Salina Formation. These beds are locally productive where they formed with synchronous drape over the patch reefs. They are considerably more productive (the A-1 unit in particular) in the lower part of the Michigan Basin where they flank the pinnacle reefs. The Salina A unit carbonates are also locally pro-

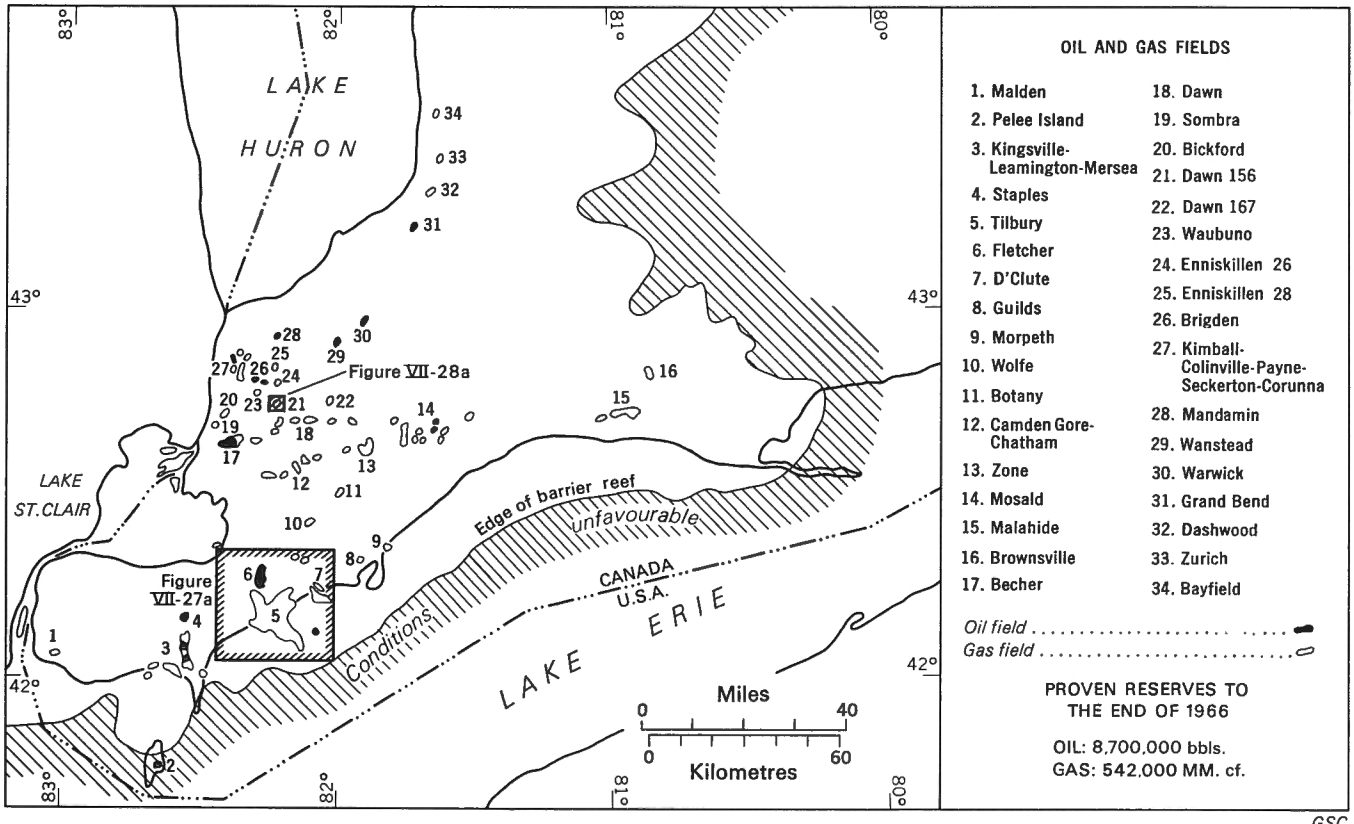

FIGURE VII-25. Principal oil and gas fields in the Silurian, Salina, and Guelph Formations of southwestern Ontario (by B. V. Sanford).

ductive in anticlines and domes that have resulted from normal faulting. Salina and Guelph production in 1966 was 11,441,937 Mcf of gas and 353,909 barrels of oil. This represents 74 per cent and 27 per cent, respectively, of the total gas and oil production in the province during that year.

The Silurian reefs of southwestern Ontario have a twofold importance, for once depleted of their primary production they are utilized as gas storage reservoirs. The storage reservoirs are filled during the summer, and are drawn upon in the winter months when requirements in Ontario and western Quebec exceed the combined local

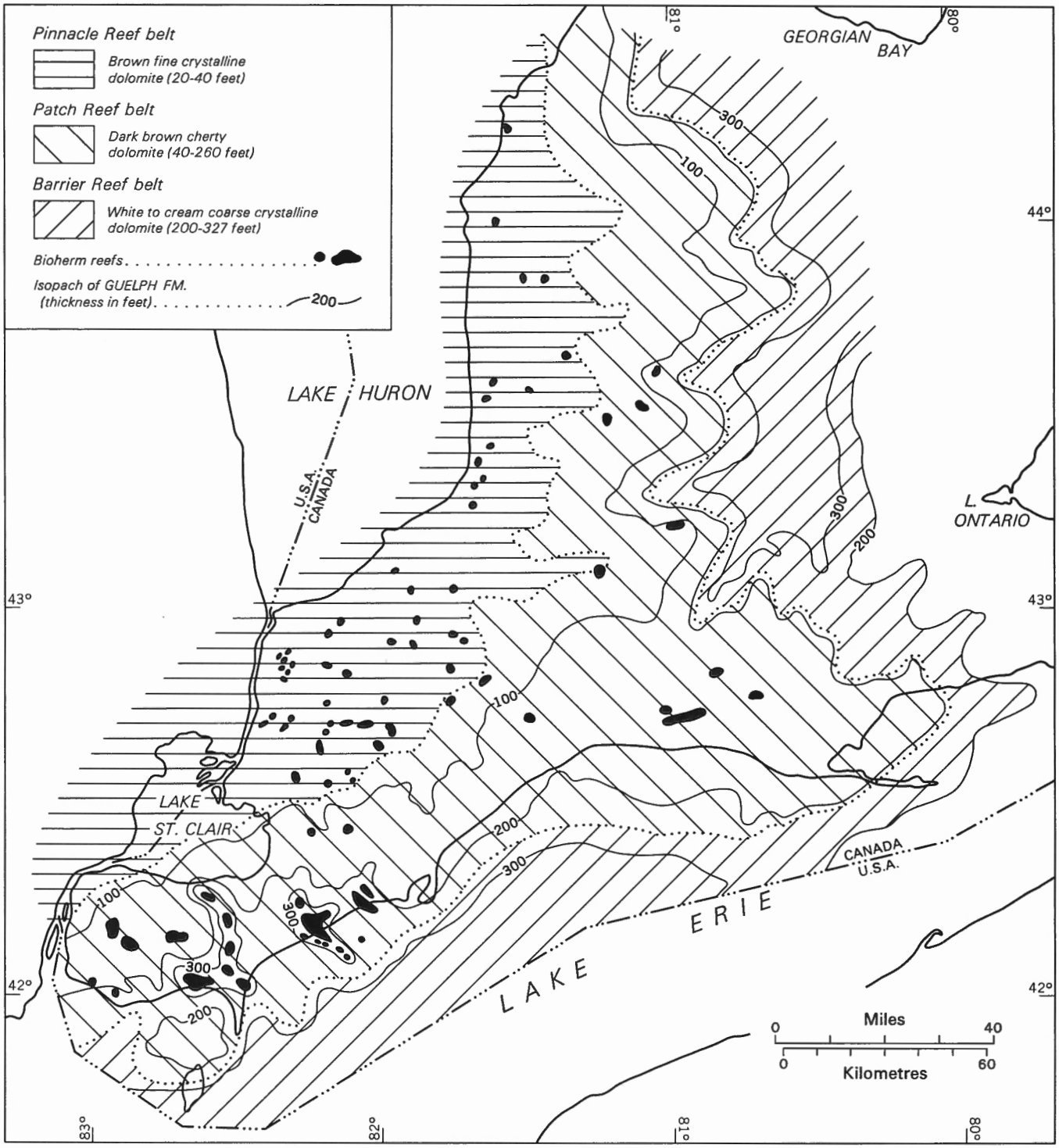

FIGURE VII-26. Facies of the Silurian Guelph Formation, southwestern Ontario (by B. V. Sanford).

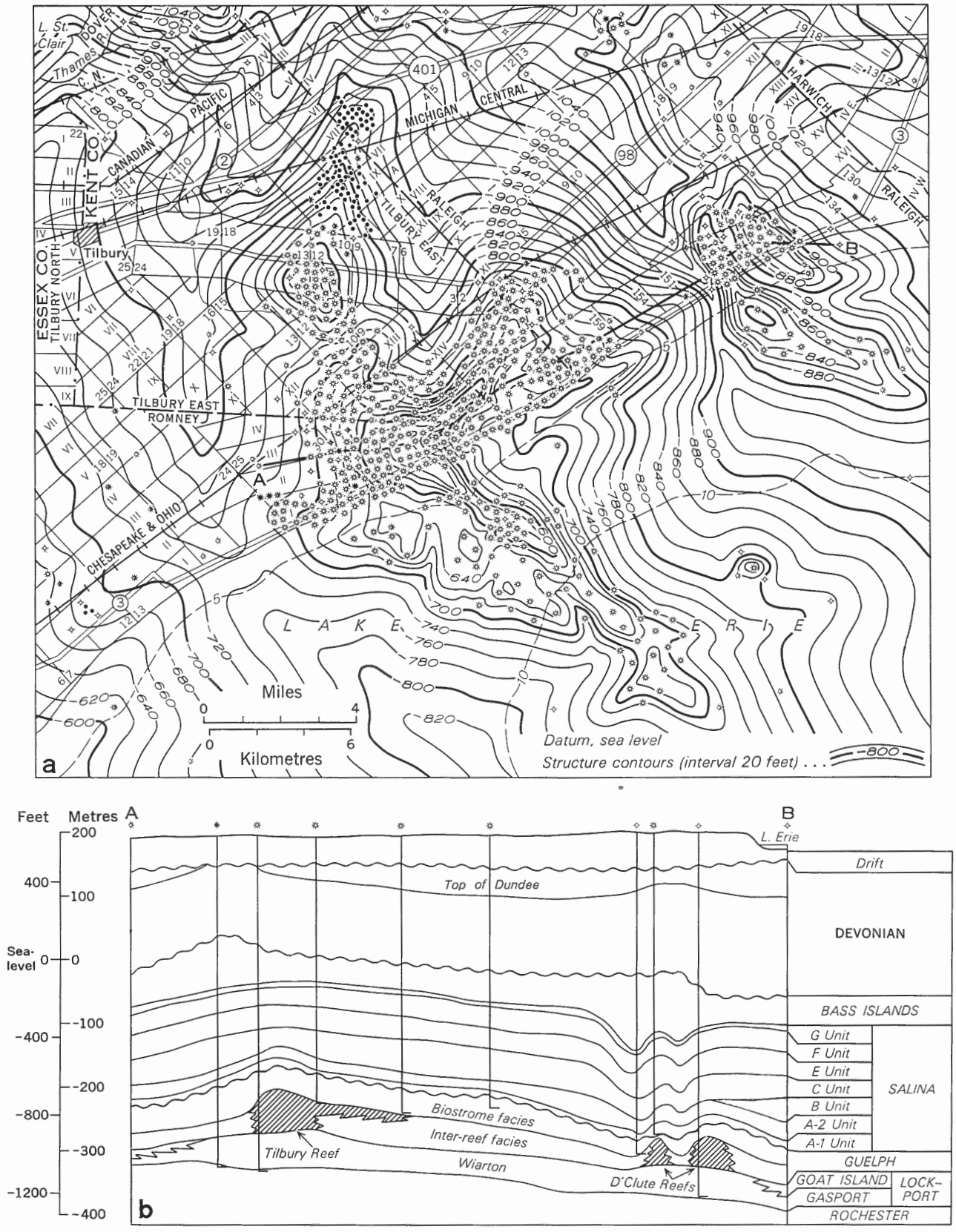

FIGURE VII-27. The Tilbury and D'Clute reefs in the Silurian Guelph Formation, Ontario (by B. V. Sanford). (a) Structure contours on top of the Guelph Formation. (b) Cross-section along line A–B.

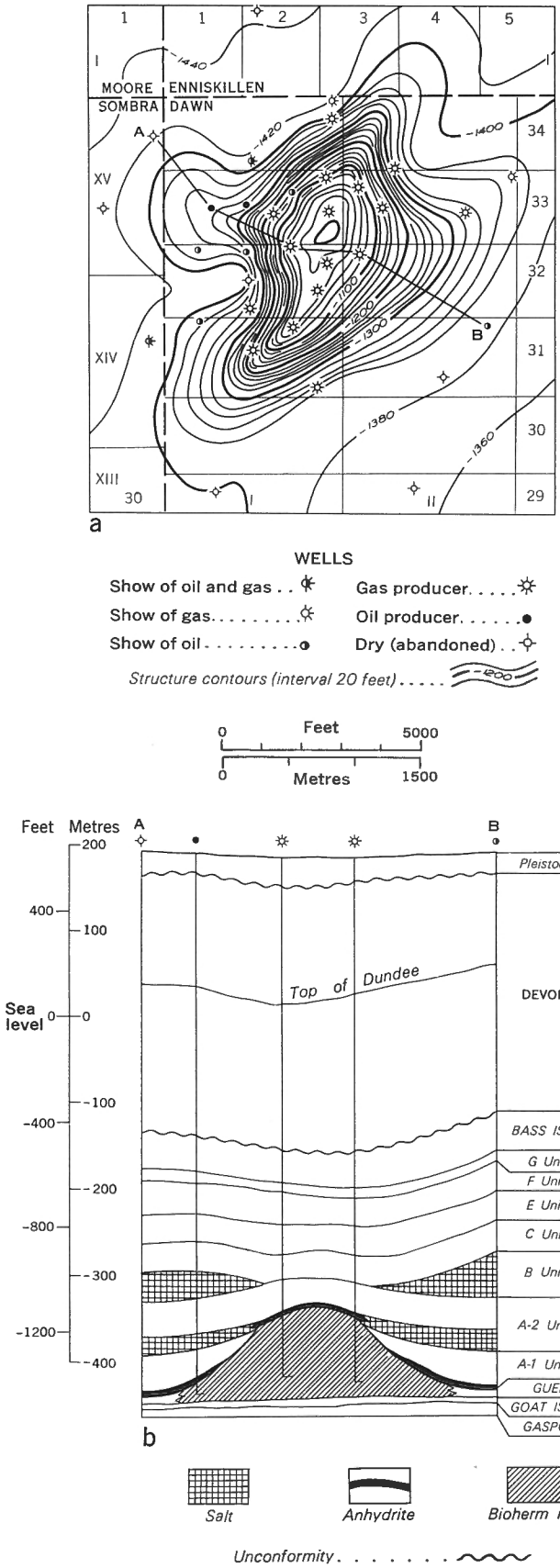

FIGURE VII-28. The Silurian Dawn 156 reef, southwestern Ontario (by B. V. Sanford). (a) Structure contours on top of the Guelph Formation. (b) Cross-section along line A–B.

production and throughput of the pipeline from western Canada. So far, nine pinnacle reefs and two Salina pools have been designated as storage reservoirs, providing a capacity of about 100 million cubic feet of natural gas.

Accounting for an appreciable percentage of current gas production are the sandstone reservoirs of the Whirlpool, Grimsby, and Thorold Formations in the Niagara Peninsula (Fig. VII-29). Four large fields have been developed gradually since the first discovery in this region in 1889: the Bayham, Norfolk, Haldimand, and Welland fields. Their combined area totals 356,460 acres on land and an additional 32,560 acres are now proven productive offshore in Lake Erie adjacent to the Haldimand and Norfolk fields (Pl. VII-7). The fields lie on the south-dipping, northern flank of the Allegheny Basin. Accumulation of the gas is mainly the result of permeability pinchout and is in no way related to local variations in regional dip. The producing formations consist of orthoquartzitic sandstones, and for the most part are rather poor reservoirs. They are extremely fine grained, and due to the advanced state of secondary quartz overgrowths their primary porosities and permeabilities have been considerably reduced. These fields yielded 3,977,142 Mcf of gas in 1966 which represents 26 per cent of the total Ontario production for that year.

Devonian Oil Fields

The distribution of Middle Devonian oil fields in Ontario is confined to an area of about 1,000 square miles extending from the southern extremity of Lake Huron to Lake Erie (Fig. VII-30). The eighteen principal fields discovered to date have an ultimate yield of 42,500,000 barrels of 33° API gravity oil. Accumulation is confined to domes in which the reservoir rocks of the Dundee and Lucas Formations have been locally dolomitized and deformed because of the solution and removal of the underlying Upper Silurian Salina salts. The series of dome structures that underlie the Petrolia region of Lambton county provide good examples of post-Devonian salt leaching. In this area, due to solution collapse, the northwestward regional dip of Middle Devonian carbonate rocks has been reversed along the southern and eastern margins of the field, resulting in structural closure. The Petrolia and Oil Springs fields have produced continuously since their discovery and account for more than half the oil produced to date in Ontario. These fields combined with the more recently discovered Rodney field produced 606,242 barrels of oil in 1966; this represents 45 per cent of the total annual yield during that year.

Quebec Sedimentary Basins

Quebec Basin and Ottawa Embayment

Approximately one hundred and eighty-five exploratory wells have been drilled at widely separated localities

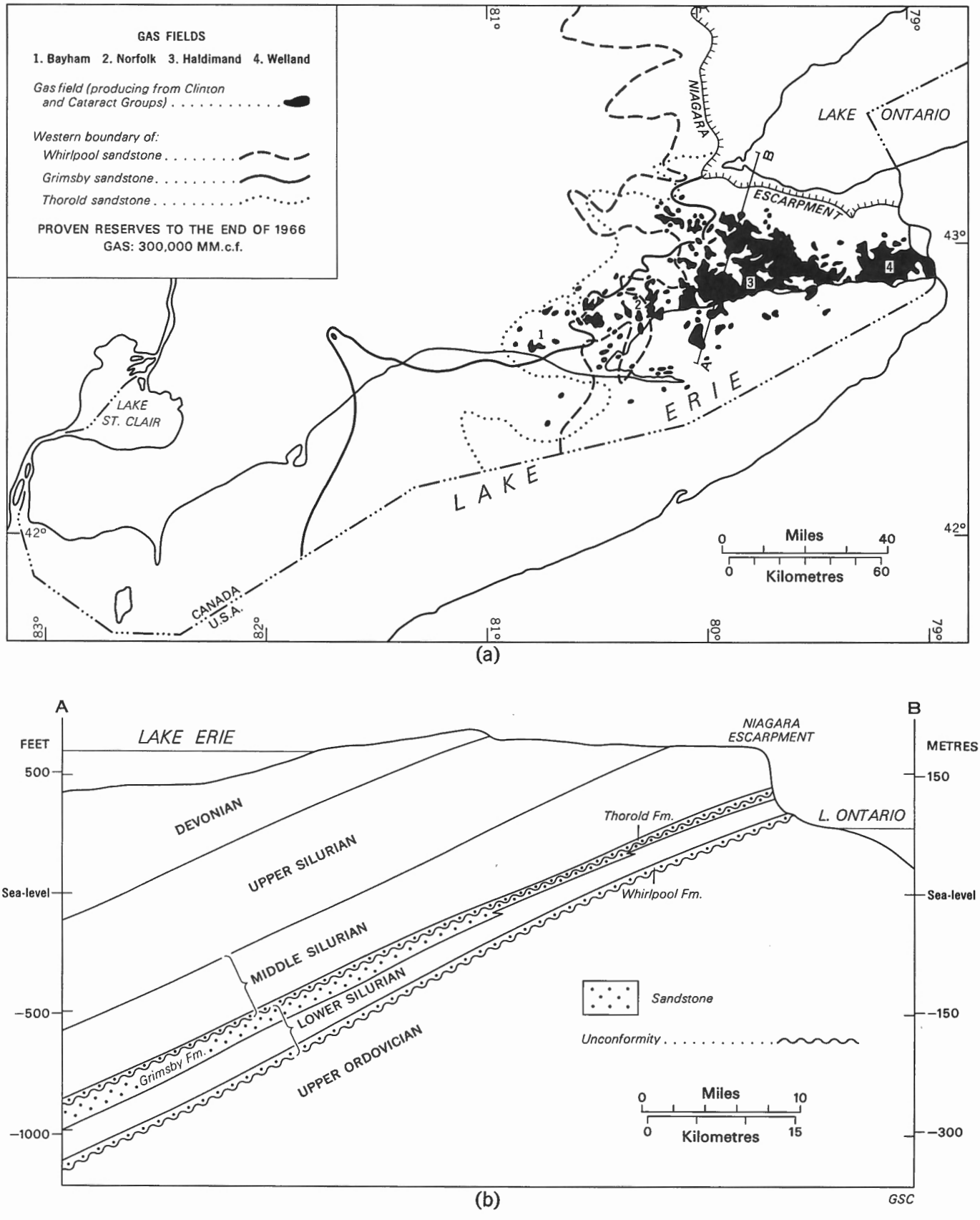

FIGURE VII-29. Gas fields in the Silurian Clinton and Cataract Groups (by B. V. Sanford). (a) Distribution of fields and limits of producing sandstones. (b) Cross-section along line A–B.

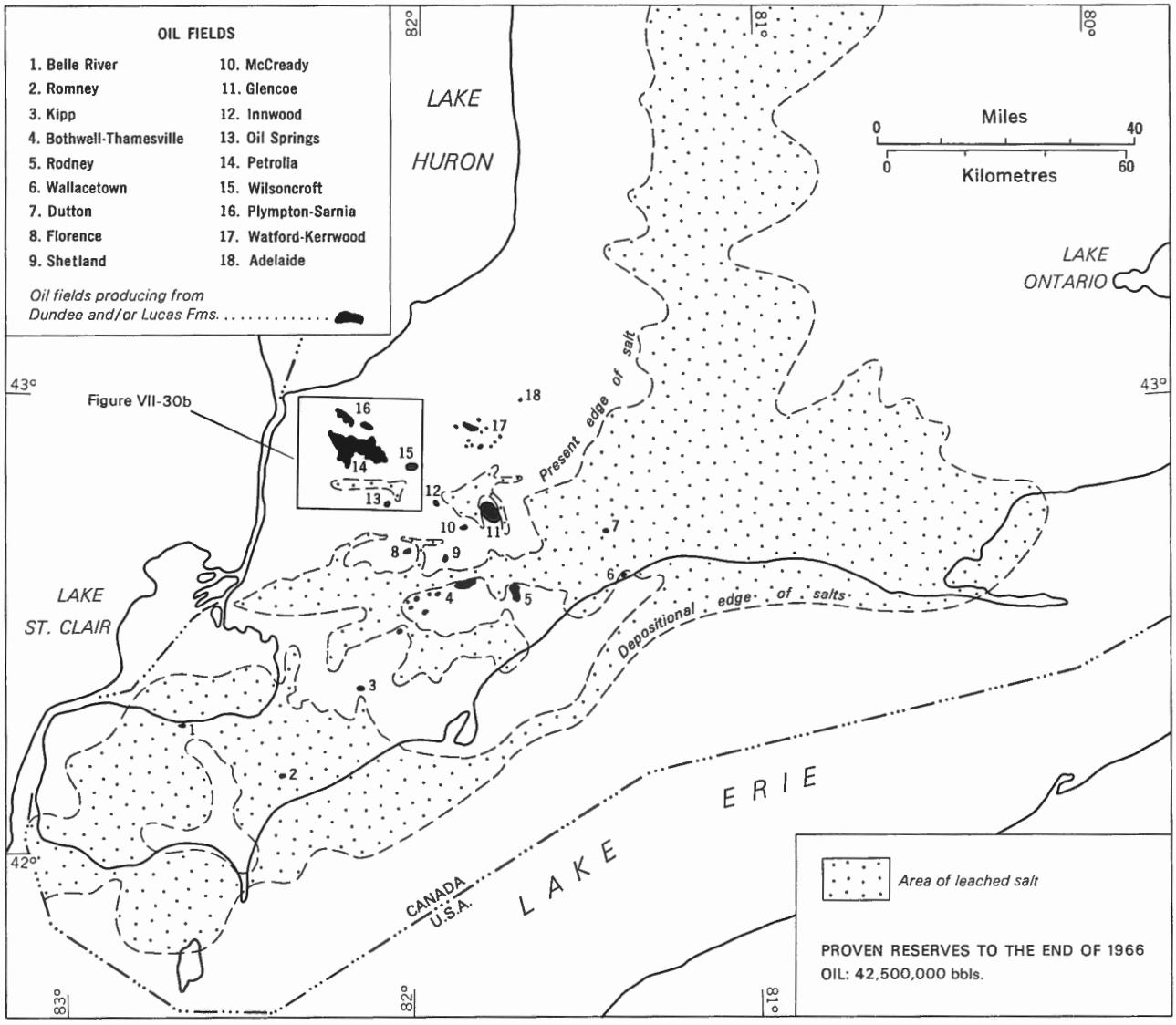

FIGURE VII-30. Devonian oil fields of southwestern Ontario (by B. V. Sanford). (a) Distribution of fields and relationship to areas of leached Silurian salt. (b) Structure contours on top of the Middle Devonian Dundee Formation, Oil Springs and Petrolia fields. (c) Cross-section along line A–B.

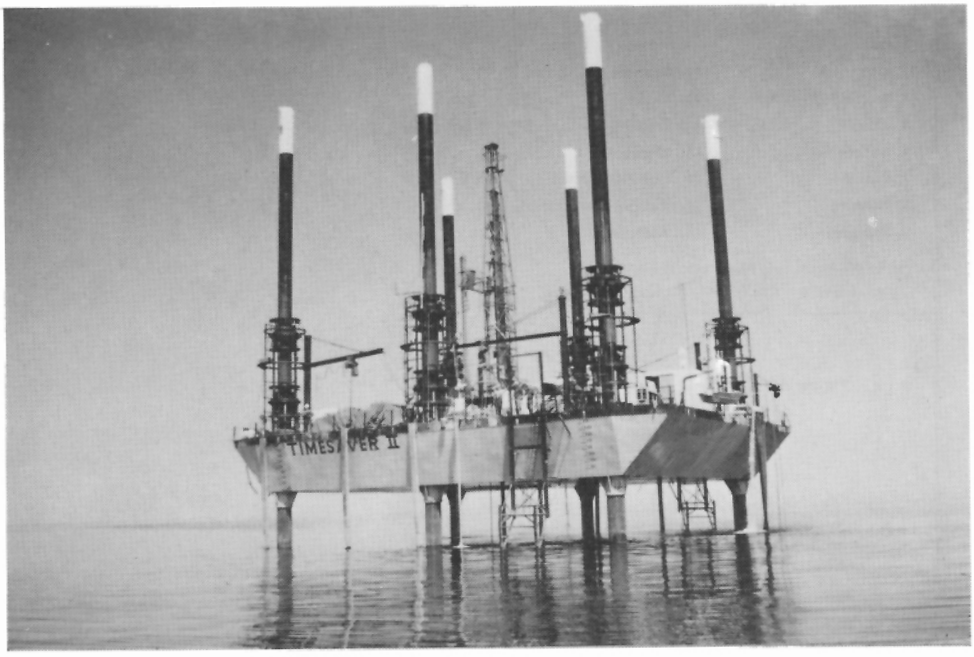

PLATE VII-7

Drilling barge 'Timesaver II' on location in eastern Lake Ontario (offshore extension of Silurian Haldimand gas field). The six hydraulic jacks lift barge above wave level to serve as a drilling platform.

in the Ottawa Embayment and Quebec Basin, none of which was successful in finding commercial quantities of petroleum or natural gas. Of the twenty wells drilled in the Ottawa Embayment, four were completed in basement rocks and ten tested only the Trenton and Black River Groups. Showings of gas have been reported in various parts of the stratigraphic column, mainly the Trenton Group, and in the overlying marine shales of the Billings and Carlsbad Formations. Showings of gas were also locally encountered in surficial deposits in an area underlain by bituminous shales. In the Quebec Basin, more than one hundred wells have been completed to depths in excess of 750 feet; only twelve of these were completed to basement. Much of the deep drilling is concentrated along the St. Lawrence River, and as yet no tests to the basement have been drilled in the deeper parts of the basin. Oil and gas shows have been recorded from the Upper Cambrian Potsdam sandstone; the Ordovician, Beekmantown, Black River, and Trenton carbonates; and from sandstones in the Upper Ordovician marine shales. Many shallow holes drilled for water have encountered substantial gas flows. The gas is commonly found at the contact between the surficial deposits and bedrock surface, generally shales of the Utica or Nicolet River Formations. It also occurs in porous layers of sand within marine clay.

Anticosti Basin

Petroleum seepages were known as early as 1812 on the west coast of Newfoundland in the vicinity of Parsons Pond, St. Paul's Inlet, and on Port au Port Peninsula. Approximately forty wells have been drilled since 1867, ranging from 300 to 4,000 feet in depth. Production was reported at 1 to 6 barrels of 33° API gravity. The reservoirs in the Port au Port wells were porous limestone intercalated with the intricately folded shales of the Lower Ordovician Humber Arm Group. The oil-bearing strata at Parsons Pond and St. Paul's Inlet were the steeply dipping dolomites of the Lower Ordovician St. George Group.

Since 1963, five stratigraphic test holes have been completed on Anticosti Island to depths of 3,239 to 6,146 feet. All are in the northwestern part of the island, and with one exception terminated in crystalline rocks of the Precambrian basement. Shows of oil and gas were encountered in dolomites of the Lower Ordovician Romaine Formation in one well.

Gaspé Trough

The Gaspé Peninsula has long attracted attention as a potential oil-producing province. Seepages of oil were recorded as early as 1836 and about eighty-five wells were drilled between 1860 and 1966. Many are reported to have encountered small quantities of oil varying in specific gravity from 20° to 47° API, but no production of commercial significance has been obtained. Potential reservoir rocks are Silurian bioherm reefs and Lower and Middle Devonian limestones and sandstones. These strata were folded in the Acadian Orogeny and rest on a folded and metamorphosed sedimentary basement complex of Ordovician and Cambrian age deformed mainly by the Taconian Orogeny.

New Brunswick Oil and Gas Fields

The Stony Creek oil and gas field near Moncton is the only commercially productive field in the Appalachian Region of Canada. The field was discovered in 1909, and to the end of 1966 had yielded 697,251 barrels of 42° to 57° API gravity oil and 26,038,162 Mcf of gas. Production in 1966 was 6,836 barrels of oil and 97,403

Mcf of gas. Production is obtained from the early Mississippian Albert Formation (Fig. VII-31). The reservoir rocks consist of sandstone interbedded with grey, grey-green, and black shale; bituminous or calcareous shale; and some siltstone and limestone. The sandstone occurs in zones that vary in number from well to well and reach a maximum of about thirty. The zones are present in groups separated by 50 to 350 feet of predominantly non-bituminous shale. Six groups occur, but the main oil production is from the lower four. The field lies on a gently dipping monocline, the north limb of a syncline, and has a proven area of about 1,600 acres. Of approximately one hundred and forty-five wells drilled, one hundred and fifteen were successful. Although initial production of gas was as much as 5 million cubic feet a day, initial production of oil was small, averaging less than 3 barrels a day.

The Albert Formation contains oil shales at Rosevale and Albert mines in Albert county, and at Taylor village in Westmorland county. Diamond drilling was carried out in 1942 and more than three thousand analyses have been made of the oil content of the shale. The results obtained at Albert mines were the most encouraging. The shales assayed an average of 10.6 gallons to the ton to a depth of 400 feet. Locally the Albert Formation also contains the solid hydrocarbon albertite. The largest deposit at Albert mines was discovered in 1851 and mining was carried on for some 15 years. The material was originally thought to be coal, but subsequently it was found to occur as a vein injected into a fracture crossing the bedding of enclosing shales. The vein had an irregular strike, a steep dip, and was as wide as 28 feet, averaging 8 feet. It was followed underground to a depth of 1,400 feet and for a length of 3,000 feet. Over 200,000 tons of albertite was shipped to the United States, where it was used for the enrichment of coal gas, the yield being 14,500 cubic feet a ton.

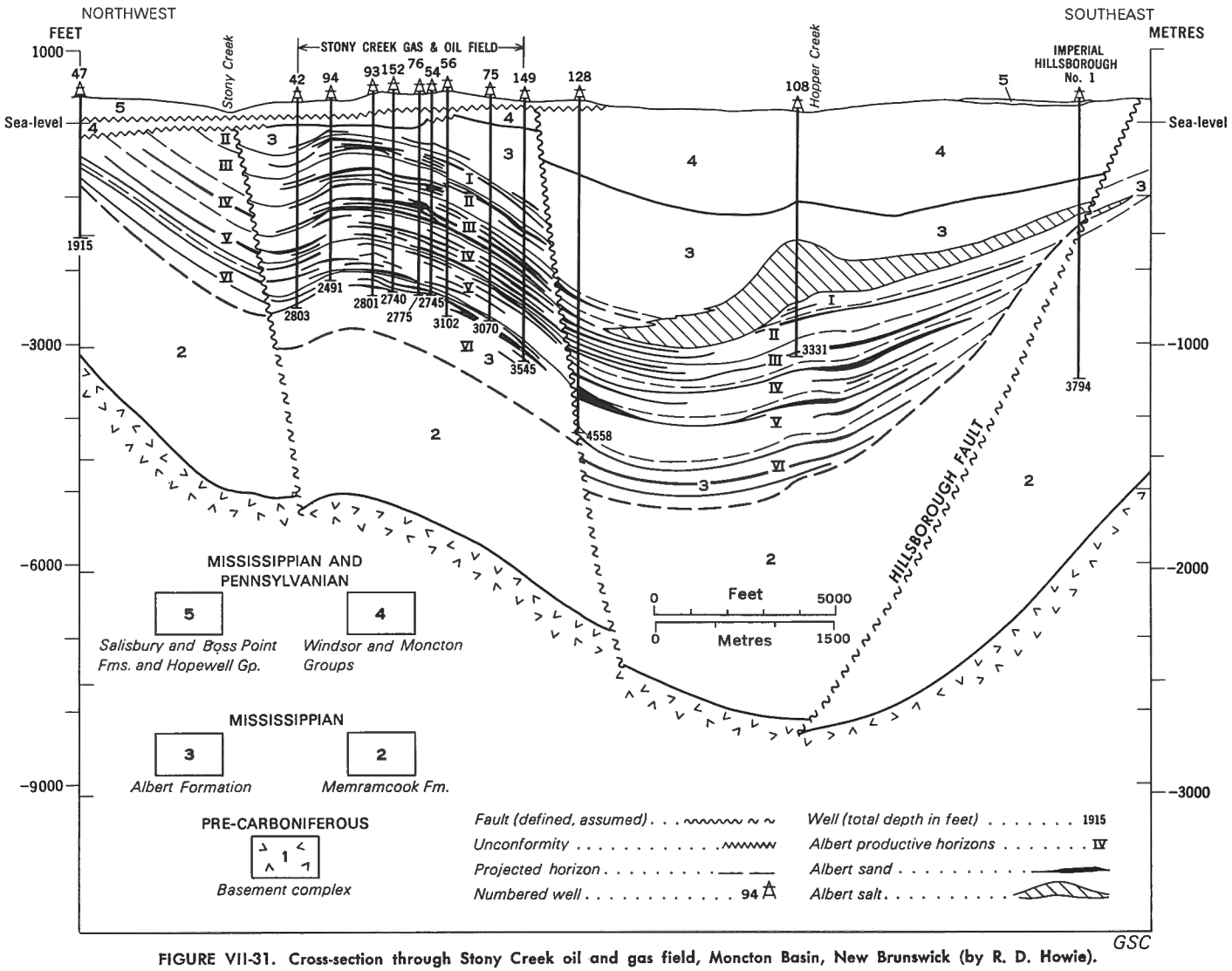

FIGURE VII-31. Cross-section through Stony Creek oil and gas field, Moncton Basin, New Brunswick (by R. D. Howie).

Coal

Coal has long been one of the most important mineral products of the Appalachian Region, and in 180 years of mining a grand total of 482 million tons has been extracted (Fig. VII-32). Although production has declined considerably during the past 25 years, in 1966 it amounted to 4,752,849 tons with a value of $59,411,101 (Table VII-1). The production from Nova Scotia declined from a near record output of 7,735,783 tons in 1940 to 4,391,832 tons in 1959, and has remained at about this figure until now. During the top production years coal was extensively mined in seven separate coalfields of Nova Scotia, with the largest field at Sydney contributing 76 per cent. In 1966, however, this field produced 88 per cent of the total output, largely because of the greatly reduced scale of mining operations in nearly all the other six fields. On the other hand the output from the only producing field in New Brunswick has been almost doubled since 1940, from 530,468 tons to 1,003,362 in 1964, but it decreased to 898,315 tons in 1966. During the past 10 years Appalachian coal production has generally accounted for about 45 per cent of the total annual output in Canada. The coal is Pennsylvanian and of high volatile A or high volatile C bituminous rank. Only in the Pictou coalfield are some medium volatile bituminous coals also represented. The coal is used to manufacture metallurgical coke and gas, as a steam coal in industrial plants, for domestic heating, and for thermal power generation. About half the annual production is consumed in the Atlantic Provinces and the rest is shipped to Quebec and Ontario. The Sydney coalfield on Cape Breton Island and the Minto coalfield in New Brunswick are the main producers. Other main fields are those of Pictou and Springhill on the mainland of Nova Scotia, but their output has greatly diminished since 1956.

Coalfields of the Riversdale Group

The oldest productive coal measures belong to the Riversdale Group and are of Westphalian A age. They occur in the St. George's coalfield of southwest Newfoundland and in the St. Rose – Chimney Corner and Port Hood coalfields of Nova Scotia.

The *St. George's* field occupies a small saucer shaped basin with a diameter of about 10 miles. The coal-bearing section is 2,200 feet thick and contains four seams. These seams, which are not more than 4 feet thick, are dirty and contain shale partings. The strata are strongly faulted, resulting in truncated, steeply dipping seams with shattered coal; commercial exploitation has not been undertaken.

Near *Chimney Corner* on the west coast of Cape Breton Island, two seams of possible economic value lie near the top of a 500-foot-thick coal-bearing member of the Riversdale Group. The seams, which have been mined very little, are only 3 to 3½ feet thick at the outcrop, dip steeply seaward, and their reserves are largely confined to an area under the sea.

The *St. Rose* field, 2 miles south of Chimney Corner, contains the same seams, but a small down-faulted block separates the two areas. The seams are 5 and 8 feet thick and occupy a land basin that measures only 1 mile by 2 miles. A small, but efficient and continuous operation has been in progress on the 8-foot seam since 1950.

The coalfield with the largest potential reserves in the Riversdale Group is situated at *Port Hood* on the west coast of Cape Breton Island. A syncline, striking parallel to the coast for about 5 miles, contains a coal-bearing section 2,000 feet thick. Two economical seams, 6 feet and 3½ feet thick, underlie 400 feet of coal measures that are concealed by the waters of Port Hood Harbour. Relatively small scale operations, on the "Six-Foot" or "Main" seam only have been carried out to date.

Coalfields of the Cumberland Group

The productive coal measures of the Cumberland Group are Westphalian B. They occur in the northern part of the Cumberland Basin of Nova Scotia and are mined at the Joggins, River Hebert, Springhill coalfields.

Near *Joggins,* in the seacliffs along the eastern shore of the Bay of Fundy, 5,000 feet of coal measures with no less than sixty-five separate coal horizons are exposed (Fig. VII-33). This section was first described in 1845 by Sir William Logan who pointed out the phenomenon of cyclic sedimentation so well displayed in these rocks. The coals are thin; only five seams attain mineable thickness; most average 3 feet or less. Inland the seams have been traced for a distance of about 19 miles, but they decrease in number and deteriorate in quality. At *River Hebert* only four seams are present, and two seams outcrop at Maccan and Chignecto. All seams have been mined at various times during the past 120 years, especially in the Joggins area where nearly all readily accessible coal has been extracted. By 1962 virtually all mining came to an end, except at River Hebert where one small mine is still operating.

The *Springhill* coalfield lies on the southern limb of the syncline that separates it from the Joggins–Chignecto coal area, but the exact stratigraphic relationship between the coal-bearing members of the two fields is not known with certainty. As a result of recent studies based on correlations made with fossil plant spores, they are considered to be contemporaneous, although individual seams may be different (Fig. VII-33). At Springhill the section is 2,600 feet thick and contains thirty-nine coal horizons, five of which attain an economical thickness of between

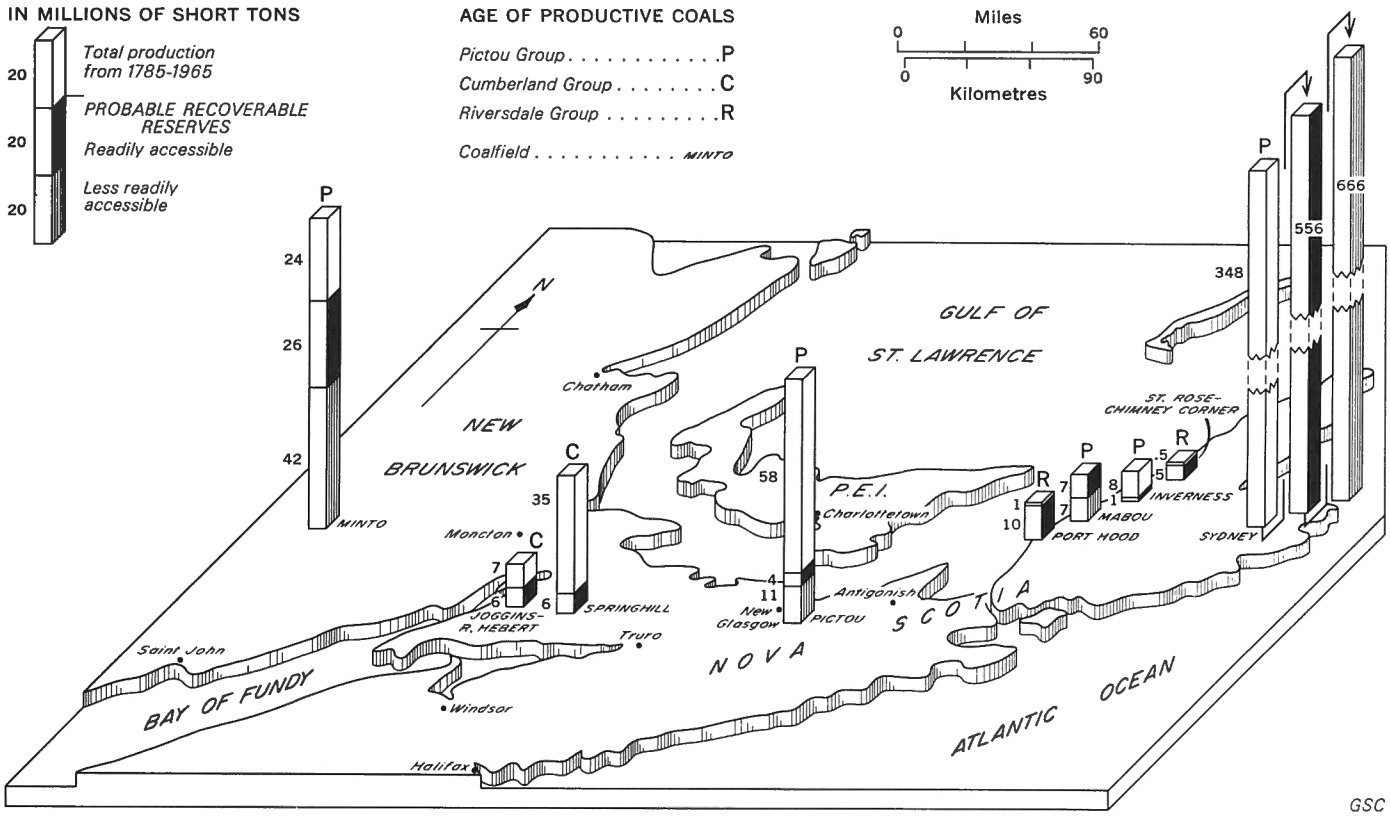

FIGURE VII-32. Location, production, and recoverable reserves of the coalfields of Southeastern Canada (by P. A. Hacquebard).

4½ and 10 feet in the area of good coal. This area is restricted to the centre of the field and extends westward from the outcrop. Along strike the seams deteriorate rapidly and become unworkable to the north and south. Coal has been mined in the area since 1870, and reached a maximum annual production of 733,329 tons in 1942. The largest producer was the No. 2 mine, which had the distinction of being one of the deepest coal mines in the world. It eventually reached a vertical depth of 4,340 feet below the surface. Production declined abruptly in 1957 and 1958 when the last two remaining collieries were closed after disastrous rock outbursts. In 1960 a small mine was opened in a hitherto unexploited part of the field. This part, however, has reserves of only a few million tons. Other large reserves still remain in the area of good coal, but these lie at a depth where extraction would be hazardous because of the outbursts or bumps. In the now-abandoned mines the bumps occurred below a depth of 1,000 feet and increased in magnitude in direct proportion to the increase in depth of overburden.

Coalfields of the Pictou Group

The youngest productive coal measures of the Appalachian Region belong to the Pictou Group and are Westphalian C and D. They occur in the Sydney, Inverness, and Mabou coalfields of Cape Breton Island; the Pictou coalfield of the mainland of Nova Scotia; and the Minto coalfield of New Brunswick.

The *Sydney* coalfield underlies a narrow fringe of lowland extending 31 miles along the coast of the Atlantic Ocean. It covers a land area of about 200 square miles and a large area where mining is carried out beneath the sea. The extent of this area is not known, but the farthest mine workings now extend nearly 4 miles from the shore. Nearly all the reserves lie in the submarine area and amount to 1,222 million tons of recoverable coal when calculated to a 5-mile limit from shore. These reserves represent 90 per cent of the total remaining reserves of the entire Appalachian Region (Fig. VII-32). The Pictou (Morien) Group in the Sydney coalfield reaches a maximum thickness of about 6,450 feet and contains twelve productive seams in the upper half of the section. The seams, 4 to 7½ feet thick, are thrown into broad open folds that trend northeasterly to easterly and pitch seaward at low angles. The prevailing dips throughout the field vary from 4 to 15 degrees, and serious structural difficulties have not been encountered in mining. Coal has been recovered at this field since 1720 when it was utilized by the French during the construction of the fortress of Louisbourg. Continuous, large-scale mining has been in progress during the past 140 years, reaching a peak annual production of 5,875,038 tons in 1940.

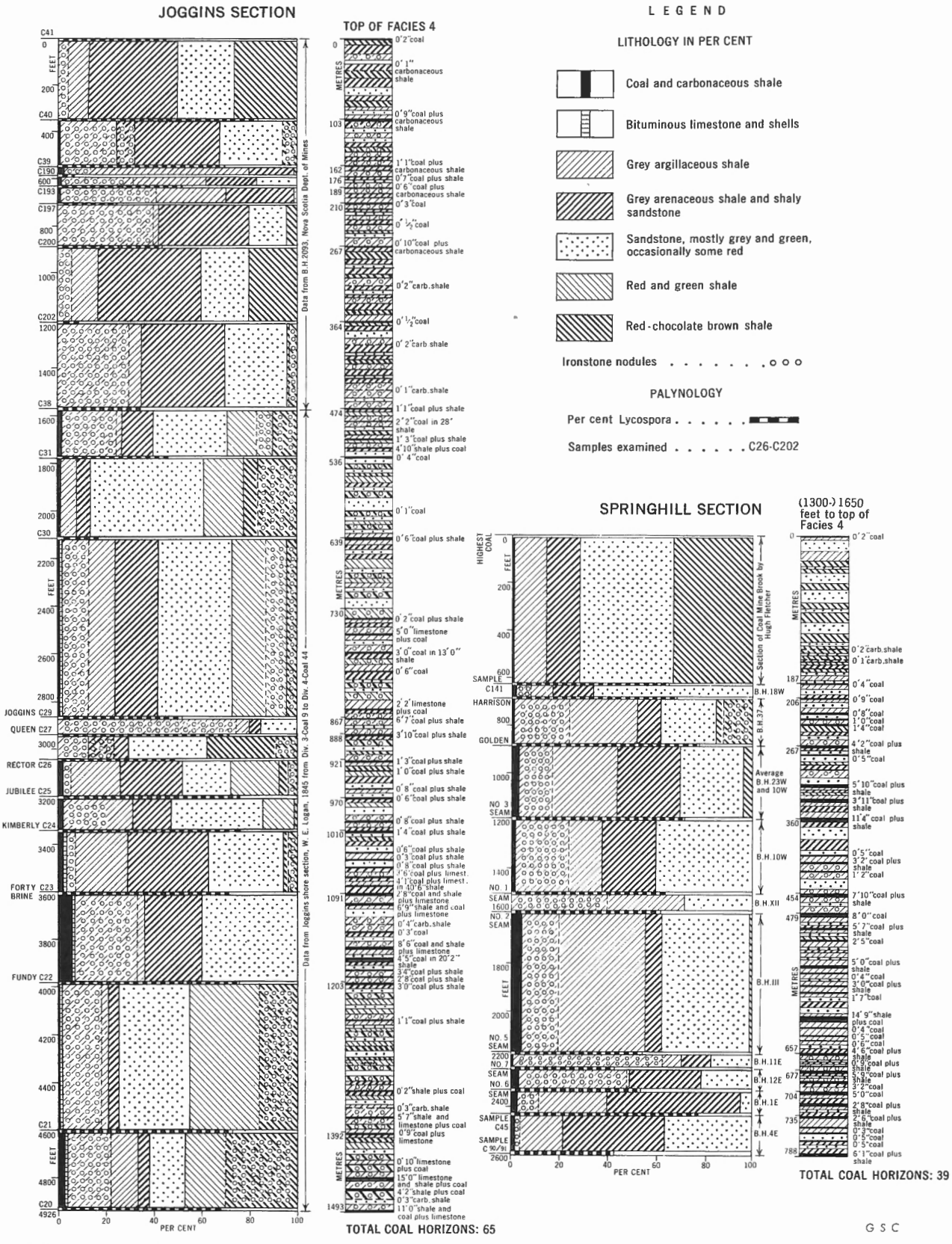

FIGURE VII-33. Lycospora distribution in several coal seams, and the lithology and stratigraphy of the coal-bearing sections at Joggins and Springhill, Nova Scotia (Hacquebard and Donaldson, 1964).

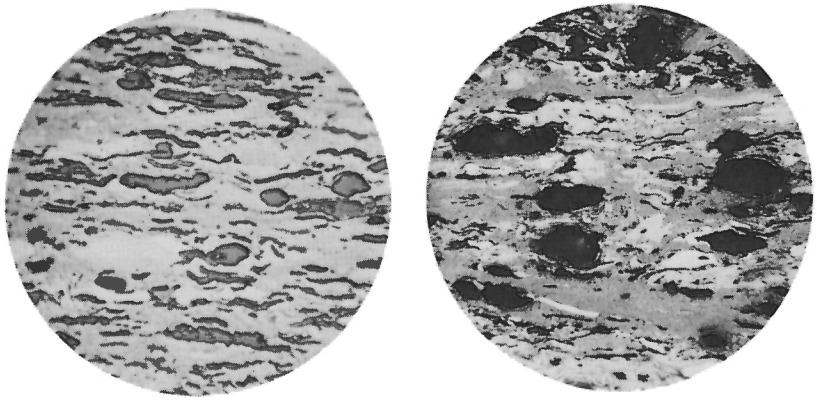

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PLATE VII-8

Photomicrographs of Carboniferous coal. (a) Well-preserved plant spores (dark grey, elongated bodies) in autochthonous coal from the Sydney coalfield. (Polished section perpendicular to bedding plane; magnification ×300; dry objective.) (b) Poorly preserved plant spores (thin, dark grey, linear forms) and detrital quartz grains (black, oval areas) in hypautochthonous coal from Pictou coalfield. (Polished section perpendicular to bedding plane; magnification ×215; oil-immersion objective.)

The deposition of the coal seams and accompanying fluvial and fluviolacustrine sediments of the Sydney coalfield took place in a valley-flat or flood-plain environment (Fig. VII-34). Normal-banded autochthonous coals accumulated in extensive peat bogs that were formed and buried where the vegetation grew. In such coals the plant entities are usually well preserved (Pl. VII-8a). The coal seams all rest on underclays with *Stigmaria* and persist for long distances with little variation in the thickness of individual coal benches. There is typical splitting and digitation of the coal by partings of clastic rocks that result from the interaction of fluviatile and peat bog depositional environments. Figure VII-34a shows the widespread distribution of five successive seams and their termination by splitting and gradual pinching out of individual coal benches. Changes in the groundwater table also had a pronounced effect on the type of peat that was formed because it affected the kind of vegetation that grew and the mode of its preservation. This is now evident in the vertical changes of petrographic composition and spore distribution within each seam (Fig. VII-34b). The presence of thin layers of durain (dull coal) that can be traced over long distances is considered a characteristic feature of coal seams that were laid down in a flood-plain environment.

The coals of *Inverness* and *Mabou mines* belong to two successive zones of the Pictou Group, those of Mabou mines being the oldest. In both fields the coal measures occupy small land areas but continue underneath the waters of the Gulf of St. Lawrence. The submarine areas may be continuous and may belong to one larger coal basin that extends along the coast for at least 13 miles. The section at Inverness measures 2,400 feet in the land area and contains four commercial seams that are from 3 to 13 feet thick; these seams have been largely depleted of recoverable coal. In only one seam hitherto not mined to any extent is a small reserve thought to be present. In 1961 virtually all mining came to an end. Peak annual production of 154,628 tons was reached in 1936. The known section at Mabou mines is 1,200 feet thick and contains six seams that are from 3 to 15 feet thick. There are also five seams, from 3½ to 11 feet thick, present in a faulted block that underlies the known portion. The land portion of this field is only a tenth of a square mile, but considerable reserves may occur offshore because of the great number of seams present. Only two seams are known that can provide ready access to this area. The workings of one seam were flooded in 1909 after only about 60,000 tons of coal had been extracted, although much development work had been carried out. On the other seam, which is 15 feet thick, very minor operations have been in progress from time to time; it constitutes the most important accessible reserve of the field.

The *Pictou* coalfield underlies an area of about 3 by 10 miles near New Glasgow, Nova Scotia, and in relation to its size, has been the most extensively exploited coalfield in Canada. Some 58 million tons have been mined over a period of about 150 years. A peak annual production of 828,368 tons was obtained in 1940. This has dropped considerably since 1956, and because of depletion the field has now reached its final stage of commercial operation. Structurally the Pictou coalfield is confined to a graben that is bounded by faults of large displacement. Within this fault block the coal measures lie in a number of open folds and become progressively younger from west to east. They reach a total thickness that is probably in excess of 8,500 feet. The section has as many as forty-five coal seams, eleven of which have been mined in three separate coal-producing districts; they range from 3 to 44 feet thick.

The Pictou coalfield is an excellent example of coal deposition in a narrow, intermontane lake basin. In this basin the fine clastic rocks with finely dispersed organic debris, such as the black shales, occupy a central position. They are flanked by areas of grey shales, which in turn are bordered by more sandy sediments. The coal seams reveal a similar picture, with clean, low-ash coal in the centre, and coarse high-ash coal at the periphery. The seams reach a great thickness, but are of limited extent

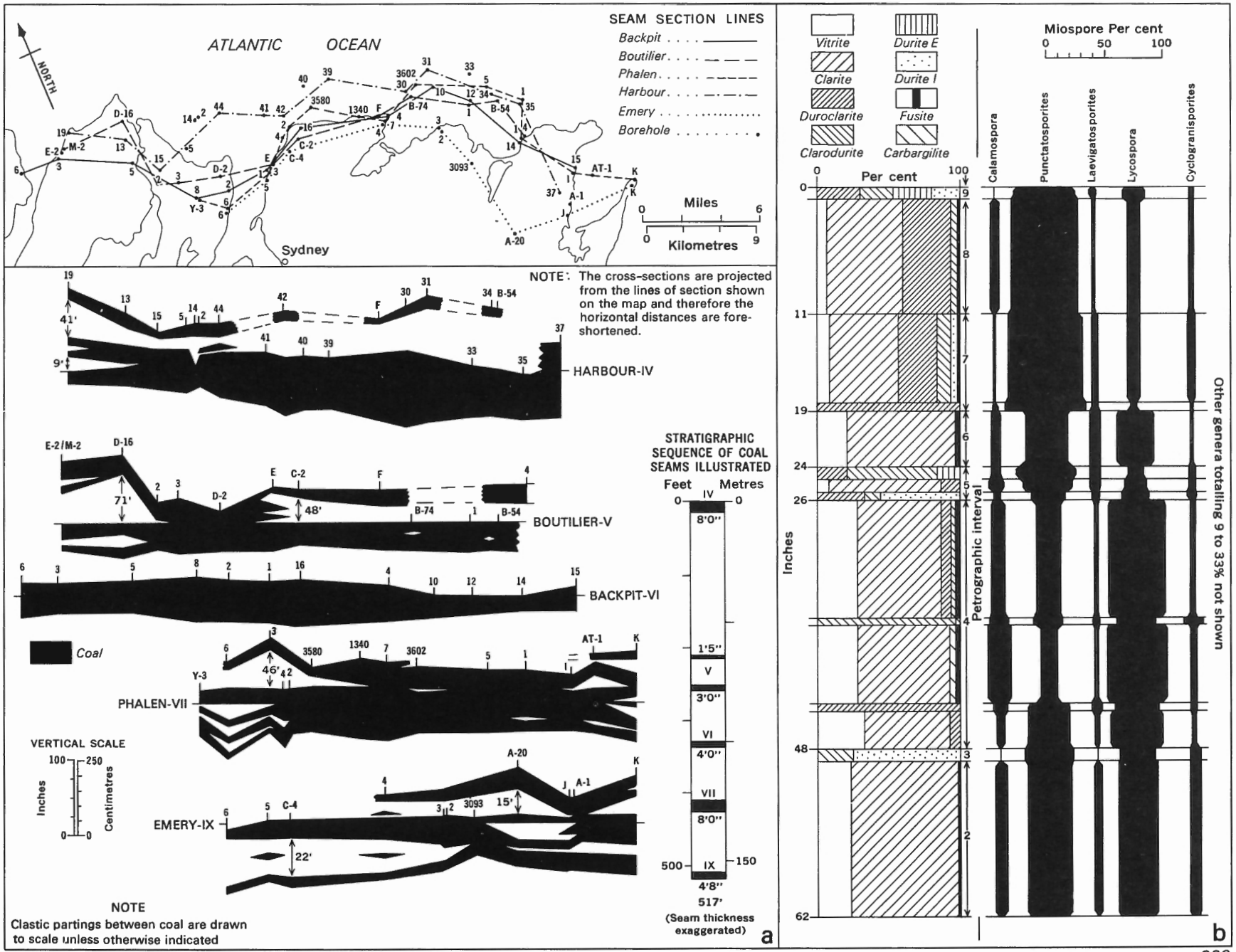

FIGURE VII-34. Flood-plain environment of coal deposition (by P. A. Hacquebard). (a) Cross-sections through five successive seams of the Sydney coalfield. (b) Palyno-petrographic seam profile, typical of the Sydney coalfield.

and terminate by lateral transition of coal into shale through zones of shaly coal and coaly shale (Fig. VII-35a). This type of facies change is typical of a limnic environment of coal deposition and is in marked contrast to the termination by splitting and digitation characteristic of a flood-plain environment. The Pictou coals are probably of drift origin, but the movement of the vegetal matter was likely restricted to the general area of plant growth, rather than being transported from outside the lake basin. Such coals are classed as hypautochthonous. They usually contain plant entities that are not well preserved and that often occur in association with detrital mineral matter (Pl. VII-8b). They are typically micro-banded, and apart from the pronounced variations in mineral matter, are of uniform petrographic composition between roof and pavement. Distinct durain bands, such as occur in the Sydney coals, are not present. The mode of preservation of the vegetal matter remained uniform throughout the life of the peat bog signifying that there were no long-lasting variations in the groundwater table. That the ecological conditions remained the same is also reflected in the remarkable constancy of the spore florules of successive benches of the same seam (Fig. VII-35b).

The *Minto* coalfield of New Brunswick occupies an area about 20 miles long and 8 miles wide. There are two coal-bearing members present; only the lower one, 100 feet thick, contains one productive seam. The mineable thickness of this seam varies from 12 to 36 inches, averaging about 18 inches; it is thickest in the southwestern section of the field and thins out in the northern and eastern parts. It occurs at shallow depth over much

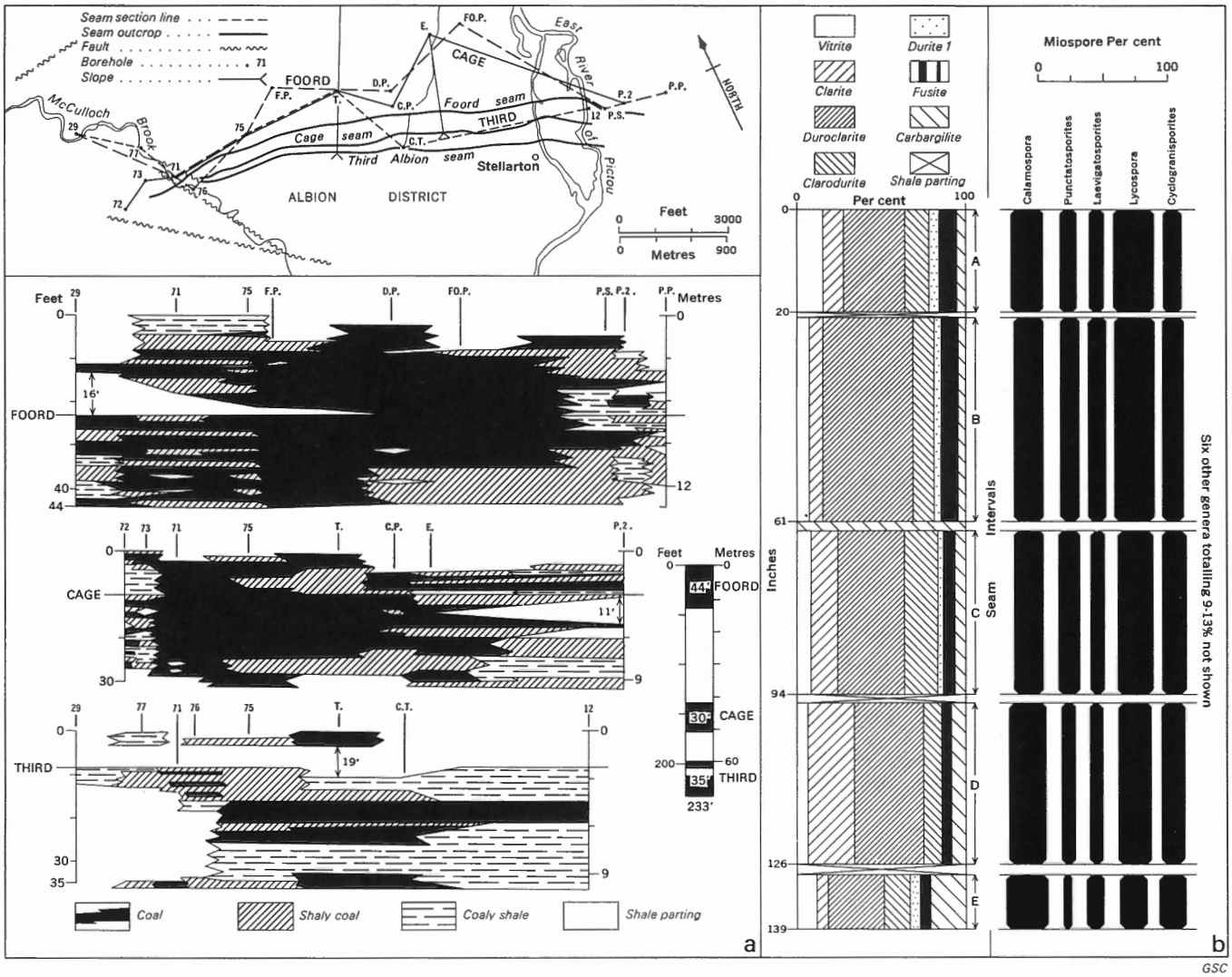

FIGURE VII-35. Limnic environment of coal deposition (by P. A. Hacquebard). (a) Cross-sections through three successive seams of Pictou coalfield. (b) Palyno-petrographic seam profile, typical of the Pictou coalfield.

of the area and is nearly flat lying. Approximately 85 per cent of the present production is by strip mining to a depth of about 70 feet. The remaining 15 per cent is extracted in underground workings to a depth of 125 feet. Since 1955 the annual output has remained fairly steady at between 800,000 and 1 million tons.

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VIII. Geology of Western Canada

Introduction 366

Precambrian 369

Paleozoic 375

Ultramafic Rocks 423

Plutonic Rocks 423

Mesozoic 428

Cenozoic 464

Selected References 481



INTRODUCTION

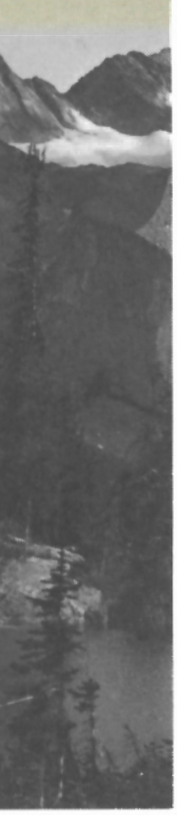
Western Canada is the part of Canada that lies west of the Precambrian rocks of the Canadian Shield. It includes the geological regions known as the Interior Platform, the Cordilleran Orogen, and parts of the Pacific and Arctic Continental Shelves and Coastal Plains (Fig. VIII-1).

The Interior Platform, underlain by Phanerozoic sedimentary rocks, is the northwestern part of the North American craton, the stable interior region of the continent. The nearly horizontal bedrock is covered by a thick mantle of glacial drift to form the plains and plateaux of the Interior Plains physiographical province. It is a region of grassland, forest, and tundra, 775,000 square miles in extent, that embraces parts of the prairie provinces of Manitoba, Saskatchewan, and Alberta, the northeast corner of British Columbia, and much of western District of Mackenzie, Northwest Territories. The southern part is composed of a relatively thick blanket of Mesozoic and Tertiary clastic rocks that overlie carbonates and evaporites of Paleozoic age, with most systems represented. They rest on a basement of Precambrian crystalline rocks. The region has been extensively drilled and the geology is well known. The northern part, however, is less well known and partly unknown; few wells penetrate the entire Phanerozoic succession. Devonian and older clastics, carbonates, and evaporites lie on Precambrian sedimentary and crystalline rocks and are covered by a thin veneer of late Lower Cretaceous rocks. Other systems are either unrepresented or are present only near Cordilleran Geosyncline and Arctic Continental Shelf. Interior Platform is linked to the St. Lawrence Platform of Southeastern Canada through the central United States of America and merges with Arctic Platform in the north, beyond the mainland.

The mainland part of Arctic Coastal Plain comprises the Cretaceous and Cenozoic rocks of Mackenzie Delta and a narrow belt that truncates the Richardson Mountains or borders the British Mountains of Eastern Cordilleran Orogen. The geology of the bordering part of Arctic Continental Shelf, 100 miles wide, is but little known.

The Cordilleran Orogen forms part of the circum-Pacific orogenic belt. It is a region, some 500 miles wide, of mountains and plateaux, trenches, valleys, and fiords. Parts are heavily forested dry grassland or tundra. Numerous peaks project above timberline or grassline and many bear glaciers. The eastern part of the orogen embraces the parts of Cordilleran Geosyncline that generally received miogeosynclinal and exogeosynclinal sediments and that underwent surficial or *décollement* deformation with little or no metamorphism, volcanism, or plutonism

[1] The accounts of the geology of the northern and southern parts of the Western Cordilleran Orogen have been prepared by H. Gabrielse and J. O. Wheeler, respectively. The stratigraphy of the Paleozoic rocks of the southern Interior Platform has been prepared by H. R. Belyea. The Mesozoic stratigraphy of the Interior Platform and Eastern Cordilleran Orogen has been written by D. F. Stott. R. J. W. Douglas is responsible for general coordination and the geology of the remaining parts of the Interior Platform and Eastern Cordilleran Orogen. When acknowledging, reference should be made to individual author and page number.



VIII

Geology of Western Canada

R.J.W. Douglas, H. Gabrielse, J.O. Wheeler, D.F. Stott, and H.R. Belyea[1]

Valhalla Range, southern Selkirk Mountains, British Columbia. Looking south to Mount Gladsheim (left) and serrated ridge carved from veined gneiss of the Shuswap Metamorphic Complex in the core of Valhalla Dome.

during the orogenic phases. The region is divisible into three main sectors. The southernmost sector, the Rocky Mountain Thrust Belt, was produced by the Columbian and Laramide Orogenies during the Cretaceous and early Tertiary. The dominant structures are subparallel, west-dipping thrusts that produce a series of narrow, linear mountain ranges formed of resistant Paleozoic carbonates. The mountains are bordered, on the east, by the Foothills belt of deformed Mesozoic and Tertiary clastic sediments. In the more southerly parts, very closely spaced thrust faults are characteristic, whereas in the central and northern Foothills they are less prominent and open and tight folds are more typical. The central sector comprises the arcuate Mackenzie Fold Belt that embraces Liard Plateau, Mackenzie, Franklin, and Wernecke Mountains, and the Taiga Ranges of Porcupine Plateau. The fold belt, characterized by Paleozoic carbonate and clastic rocks thrown into broad simple folds with intervening zones of complex folds and faults, was also produced during the Columbian and Laramide Orogenies. However, in some culminations the effects of the Hadrynian Racklan Orogeny on Helikian rocks are also evident. The northernmost sector, the Northern Yukon Fold Complex, includes Richardson and British Mountains, Porcupine Plateau, and Old Crow Range. A great variety of structures and trends has been produced as a result of several orogenies, mainly the Upper Devonian Ellesmerian Orogeny, the Columbian Orogeny, and the early and late phases of the Laramide Orogeny.

The western part of Cordilleran Orogen embodies the parts of the Cordilleran Geosyncline that received eugeosynclinal and epieugeosynclinal assemblages. The eugeosynclinal belt was at times dominated by northwesterly elongate troughs separated by geanticlines; the latter were repeatedly the site of regional metamorphism, granitic emplacement, and intense deformation. The deformation of the rocks in the troughs was controlled largely by their lithology; where volcanic rocks are predominant, broad folds, warps, and numerous steep faults developed, whereas the sedimentary rocks were deformed principally by folding and locally by thrusting. Ultramafic rocks are spatially restricted to the eugeosynclinal belt and temporally associated with basaltic volcanic rocks. They occur along thrusts or near the faults bounding major structural units. The younger epieugeosynclinal basins, which are separated by arches and geanticlines, are partly successor basins and partly synorogenic basins. While the adjoining geanticlines were being deformed, they received thick assemblages of rapidly deposited, marine and non-marine, clastic sediments and minor volcanic rocks. They were subsequently folded and faulted. Post-orogenic, slightly deformed, continental, clastic and volcanic rocks were deposited locally in fault troughs and are commonly associated with extensional zones of normal faults. Late Tertiary and Pleistocene clastic sediments occur along Pacific Continental Shelf and parts of the Coastal Plain.

The two belts of crystalline rocks, the Coast Plutonic Complex and the Omineca Crystalline Belt have been generally positive geanticlines since mid-Paleozoic time. They merge northwestward to form the Yukon Platform.

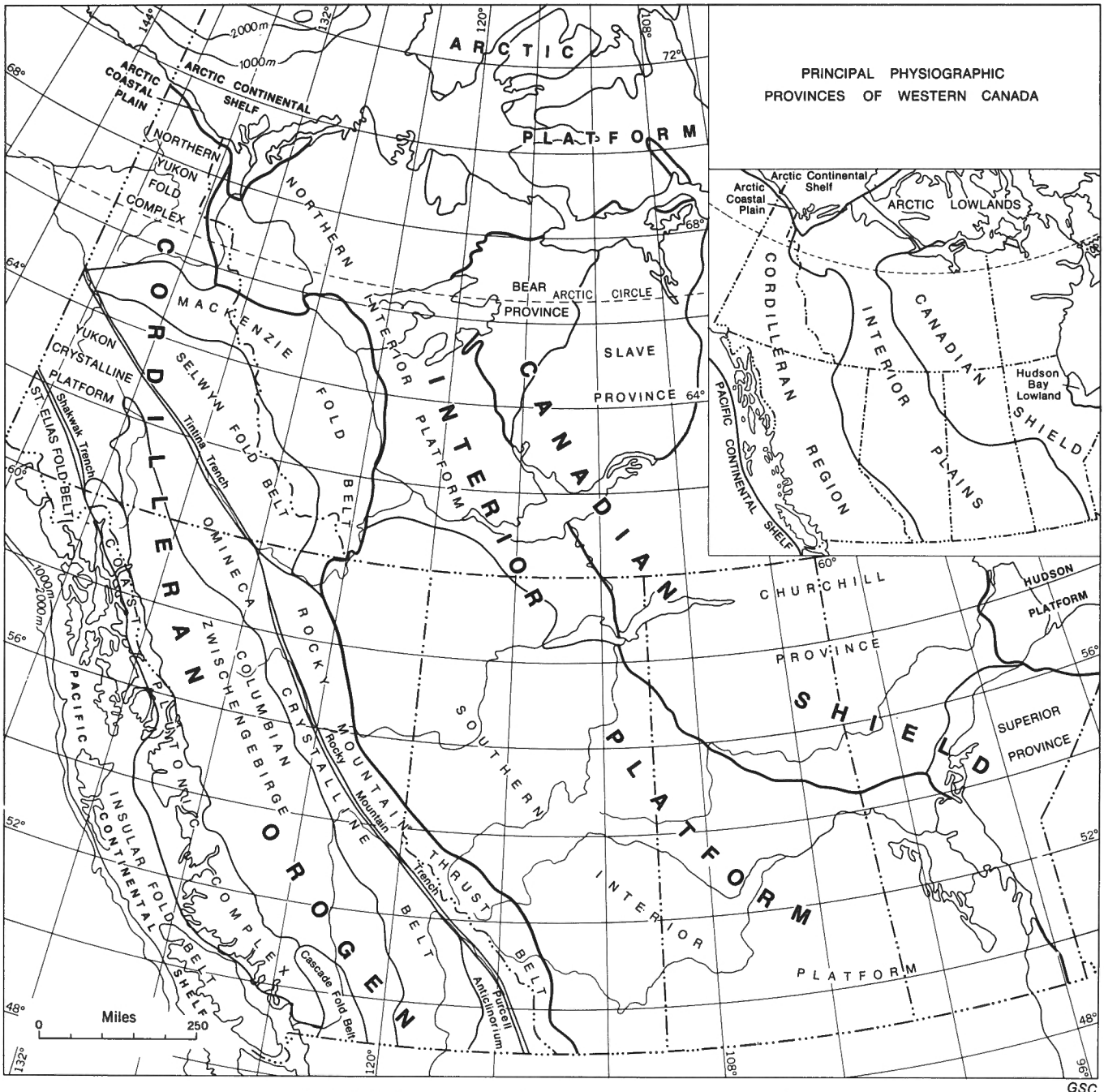


FIGURE VIII-1. Principal geological elements of Western Canada.

The western or Coast Plutonic Complex is composed predominantly of granitic rocks. It and the Cascade Fold Belt have gneiss–migmatite cores. Both belts were affected by a mid-Paleozoic orogeny, the late Jurassic to Cretaceous Columbian Orogeny, and by a late phase of the Laramide Orogeny. The northern part of Coast Plutonic Complex was also affected by the Middle Triassic Tahltanian Orogeny. The eastern or Omineca Crystalline Belt includes extensive areas of metamorphic rocks that form the core zone of an alpine-type orogen, the marginal zone of which is the Eastern Cordilleran Orogen. Metamorphic complexes occur as anvil-shaped culminations. Part of the belt was deformed during the Hadrynian East Kootenay Orogeny and the Upper Devonian to Mississippian Caribooan Orogeny. Most was redeformed during the Tahltanian and Columbian Orogenies. In the south, the eastern and western parts of the Cordilleran Orogen are separated by Rocky Mountain Trench, a linear erosional zone partly controlled by normal faults.

Columbian Zwischengebirge between the Omineca Crystalline Belt and the Coast Plutonic Complex is structurally heterogeneous, but dominated by an elongate

northwest-trending geanticlinal belt of disconnected horsts and by two broad transverse arches. Thick eugeosynclinal sedimentary and volcanic rocks were deposited during the late Paleozoic and early Mesozoic, partly synchronously with orogeny in the bordering crystalline belts. Deposition was locally interrupted by the Triassic Tahltanian Orogeny and the Jurassic Inklinian and Nassian Orogenies. As a result of the last orogeny, the region was segmented into three epieugeosynclinal basins, the sediments of which, partly non-marine, were folded and block faulted during the late phase of the Columbian Orogeny. In the Tertiary, the Zwischengebirge was subjected to transcurrent and normal faulting and extensive sheets of basic volcanic rocks were extruded. The early Tertiary volcanism was coeval with emplacement of small high-level granitic plutons.

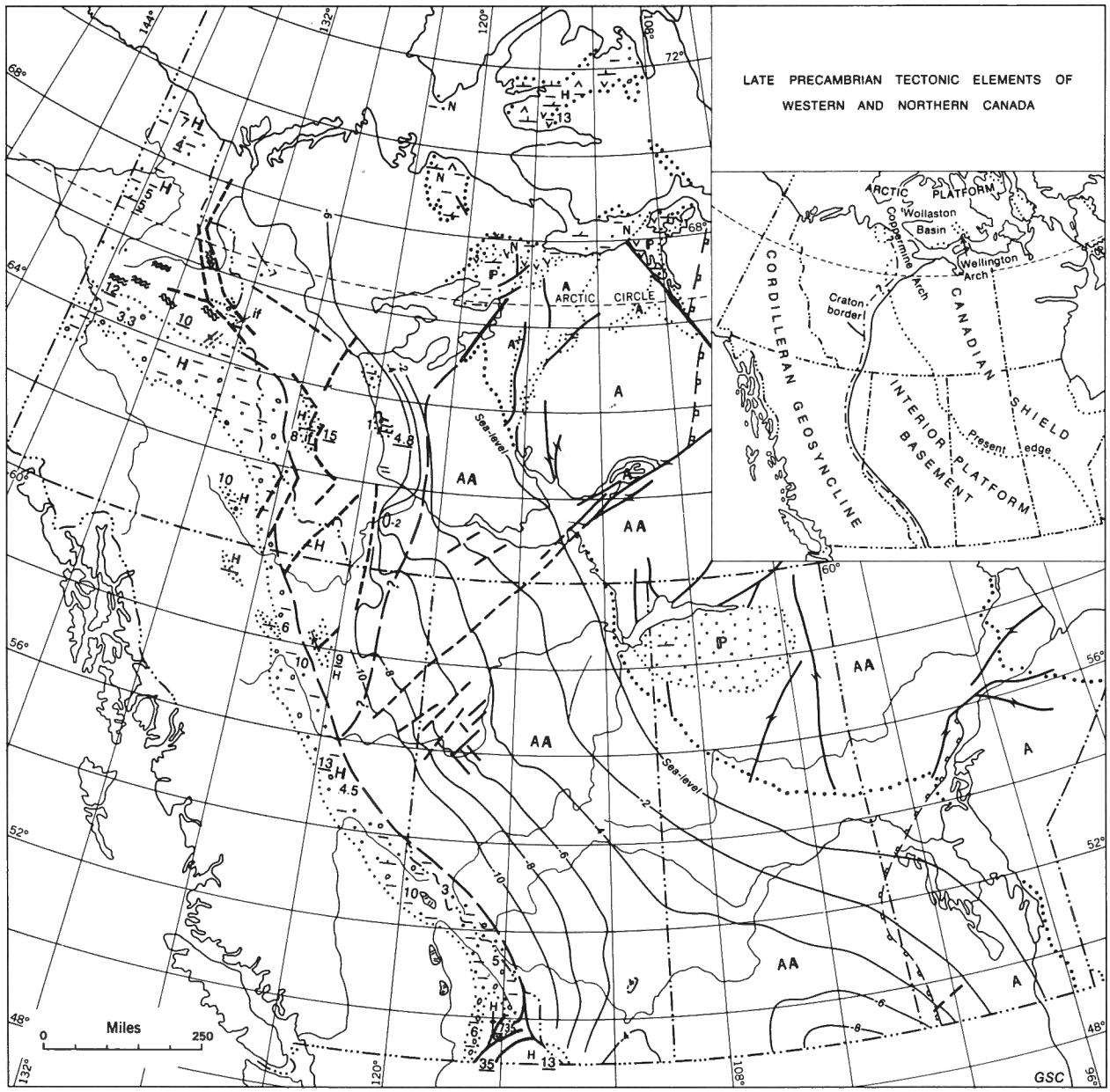
Bordering the Pacific Ocean are the St. Elias Fold Belt in southwestern Yukon Territory, the Insular Fold Belt embracing Vancouver and Queen Charlotte Islands, and the Pacific Coastal Plain and Continental Shelf. The St. Elias Fold Belt, bounding the Gulf of Alaska, contains rocks of Devonian to Pliocene age. It is composed of a core of crystalline rocks flanked by outwardly directed thrusts. The orogen was produced by late Paleozoic deformation, by the Columbian Orogeny, and by Tertiary deformation as young as late Pliocene. It is bounded on the northeast by the Shakwak transcurrent fault, part of the Denali system of Alaska. The Insular Fold Belt contains late Paleozoic to Tertiary rocks and is characterized by thick sequences of Mesozoic volcanic rocks that are only slightly warped and faulted. Deformation was not synchronous; on Vancouver Island it took place during the Middle Jurassic Nassian Orogeny whereas on Queen Charlotte Islands it occurred during the early Columbian Orogeny. Following the development in latest Cretaceous time of the epieugeosynclinal Georgia and Hecate Basins between the Coast and Insular Geanticlines, the region underwent local folding, faulting, and volcanism. The Pacific Continental Shelf is narrow, terminating a few miles west of Vancouver and Queen Charlotte Islands. Small parts of the islands underlain by flat clastic Tertiary sediments form the Pacific Coastal Plain. The Queen Charlotte fault, a transform fault related to the San Andreas–Denali linkage is steeply east-dipping with recent right-lateral movement. It lies near the foot of the steep Continental Slope and truncates the north to northeast magnetic trends of the ocean floor.

The Cordilleran Orogen of Canada is one of great geological variety and complexity that developed during the evolution of the Cordilleran Geosyncline from mid-Proterozoic to Recent. At present the region is tectonically quiet, relative to other parts of the circum-Pacific orogenic belt.

PRECAMBRIAN

Beneath the Phanerozoic strata of the Interior Platform are Precambrian sedimentary and crystalline rocks (Fig. VIII-2), the latter forming the basement of the western part of the North American craton. Potassium–argon age determinations (Burwash, *et al.*, 1962) on material obtained in the drilling of wells indicate that the structural provinces established on the Canadian Shield by Stockwell (Map 1251A) extend beneath Interior Platform. A small part in the southeast is an extension of Superior Province, formed of Archean rocks deformed by the Kenoran Orogeny, the major structural trends of which strike predominantly east-west. The greatest part however, represents a westward extension of the Churchill and Bear Provinces, composed of rocks of Archean and Aphebian ages deformed by the Hudsonian Orogeny. The major gneissic and structural trends, although predominantly northeast-southwest, vary to easterly, northerly, and northwesterly, and form complex arcs or bend around the borders of Slave and Superior Provinces. Major post-orogenic faults, on which displacements are transcurrent and normal, fall into three distinct sets: northeast, north, and northwest. On some, minor displacements occurred in the Phanerozoic.

Unmetamorphosed, slightly deformed Helikian and Hadrynian sedimentary rocks of Coppermine Subprovince of the Shield lie unconformably on the crystalline rocks of Bear and Slave Provinces and probably extend beneath the Paleozoic of the northern part of the Interior Platform. The thickness of these sediments is not known, nor whether they occur in geosynclinal development or only as cratonic cover. The depth and configuration of the crystallines accordingly are unknown and the position of the margin of the craton is uncertain. Helikian and Hadrynian sediments are common in the more easterly parts of the Cordilleran Geosyncline but have yet to be recognized in the more westerly parts. Known stratigraphic successions measure several tens of thousands of feet but are not known to be floored by basement. Some pre-Devonian and pre-Lower Cambrian crystalline basement occurs in the southern Cordillera but the time of its formation and the age of the constituent rocks have not been established. The little that is known about isopach and facies trends in the sedimentary rocks suggests that the loci of much of the margin of Cordilleran Geosyncline, and possibly also some of the major elements in the north, were established in the Helikian. Some prominent faults occurring within or bounding areas of Proterozoic rocks were in existence at the close of Precambrian time. They were active during the Phanerozoic and served to separate areas of strong depression from areas that were less depressed or uplifted.



DEPOSITIONAL FEATURES

(Hadrynian in Cordilleran Geosyncline, Helikian on craton)

- *Conglomerate*
- *Sandstone*
- *Shale, mudstone*
- *Limestone*
- *Dolomite*
- *Anhydrite*
- *Iron-formation* — if
- *Volcanic rocks* — V
- *Thickness in thousands of feet; Hadrynian, Helikian* — 8, 15
- *Core of gneiss domes*

PRE-PALEOZOIC GEOLOGY

Hadrynian	H	*Aphebian*	A
Neohelikian	N	*Archean*	A
Helikian	H	*Archean and Aphebian*	AA
Paleohelikian	P		

- *Truncated limit*
- *Outcrop limit*
- *Present edge of shield*
- *Fault, fault inferred*
- *Transcurrent fault*
- *Metamorphic front*
- *Structure contour, base of Paleozoic rocks on the craton, in thousands of feet below sea-level* — -2

RACKLAN AND EAST KOOTENAY OROGENS

- *Anticline*
- *Syncline*
- *Phyllite*
- *Granitic pluton (age in millions of years)* — 735

FIGURE VIII-2. Pre-Paleozoic paleogeology of Western Canada, geology of adjacent parts of the Canadian Shield, and some features of Helikian and Hadrynian sedimentation and tectonism.

Their trends are comparable to those cutting Archean and Aphebian crystalline rocks of the craton and the faults accordingly may also involve crystalline crust, as those in the geosyncline bordering the continent are more active and impart greater mobility to the crust than those in the interior.

The Proterozoic successions of the southeastern Cordillera are customarily referred to the Purcell and Windermere Systems. They are separated by a widespread unconformity and an interval of folding, metamorphism, and intrusion of granite stocks, these effects being considered to be the result of the East Kootenay Orogeny. The granites (Leech, 1962a) have yielded K–Ar dates of 675 to 745 million years which suggest that the orogeny occurred within the Hadrynian era, being somewhat younger than the Grenvillian Orogeny of the Canadian Shield. The Purcell is probably therefore mainly Helikian in age although part may be as young as early Hadrynian and, by virtue of its allochthonous position with respect to Hudsonian metamorphosed basement, part of the Purcell may be late Aphebian. The Windermere falls entirely within the Hadrynian and, as it is locally conformable with Lower Cambrian sediments, probably represents all or most of the late Hadrynian. Elsewhere, most Proterozoic sequences bear some lithological similarity to either the Purcell or Windermere, or may be metamorphosed by Phanerozoic orogenies. In the northern Cordillera, the sedimentary successions are likewise separated by a profound angular unconformity which resulted from the Racklan Orogeny that produced easterly and north-northeasterly trending folds and phyllitization, and subsequent taphrogenic activity, tilting, and volcanism. The Racklan Orogeny is presumed to be temporally equivalent to the East Kootenay Orogeny. The Moyie and other gabbro sills in Purcell Mountains have been dated at about 1,100 m.y., and the lead minerals of Sullivan orebody in the Aldridge Formation of the Purcell have yielded isotopic ages at about 1,340 m.y. (Sinclair, 1966). In adjacent northern parts of the Shield faulting and intrusion of diabase sills occurred very late in the Proterozoic and an interval of strong faulting, volcanism, and ultramafic extrusion, the Muskox disturbance, occurred at about 1,100 m.y or about mid-Helikian time.

Helikian (Middle Proterozoic)

Tectonic Summary

Helikian sedimentary strata, the oldest known in the Cordillera, comprise a thick sequence of fine-grained clastic and carbonate rocks predominantly of shallow-water marine origin (Fig. VIII-2). Andesitic flows and gabbro dykes and sills occur sporadically. Facies changes in the lower part of the succession from fine deep-water clastics to conglomeratic red-bed and shallow-water sediments and a general eastward convergence suggest that the sequence was deposited on a shelf or in a miogeosyncline bordering the western margin of the craton. Farther north, lateral changes in facies of Helikian rocks have not been recognized although, in general, the strata resemble the eastern facies of the south. Orange-weathering and varicoloured rocks and carbonates containing abundant stromatolites are particularly characteristic. It seems probable that the easternmost and northeasternmost exposures of these rocks fairly closely approximate the western and southwestern margin of the North American craton during Helikian time. The nature and extent of depositional troughs to the west is unknown.

Cordilleran Geosyncline

The succession and correlation of the Purcell rocks in the Clark, Galton, and Hughes Ranges of the *southern Rocky Mountains* and in the *Purcell Mountains* has become well established (Price, 1964a). The Purcell is divisible into two main parts by an hiatus, partly represented by the Purcell lava, a very extensive basalt flow several hundred feet thick (Fig. VIII-3). In Galton Range the lava is absent through erosion. In southern Purcell Mountains, the hiatus may be represented by 2,000 feet of strata with several thin flows. The lower part of the Purcell is divisible into three main units. The lowest includes the thin, varicoloured, laminated, and stromatolitic carbonates and the conglomerates with pebbles of quartzite and dolomite that comprise the near-shore Waterton and Altyn Formations in Clark Range, and which grade westward into various facies including that of the very thick Aldridge Formation. The Aldridge comprises rusty-weathering, fine-grained, laminated, argillaceous quartzite and dark argillite with graded bedding and scour and fill structures, indicating deposition in deep water. The middle unit contains red and green argillite and quartzite that comprise the Appekunny and Grinnell Formations in the east. These rocks contain such tidal flat and shallow-water features as channel fillings, torrential ripple-marks, mudcracks, and salt hopper and raindrop markings, features that are generally lacking in the equivalent but thicker Creston Formation of the west. The upper unit of the lower Purcell, represented by the Siyeh Formation in the east, contains much clean carbonate typically with "molar tooth" and stromatolitic and oölitic structures, interbedded with shale and siltstone in the lower and upper parts. The latter unit constitutes the Siyeh of the Purcell Mountains and the underlying rocks, argillaceous and silty dolomites, form the Kitchener Formation. The upper part of the Purcell, the Sheppard, and lower Gateway Formations contain much argillaceous and silty dolomite whereas the conformable higher units of alternating argillite and quartzite vary from green and grey to purple and red and exhibit shallow-water sedimentary structures. The Purcell is intruded by the Moyie gabbro sills and is unconformably overlain by the Hadrynian Windermere System in western and northern Purcell Mountains and by various Paleozoic formations in the eastern Purcell and southern Rocky Mountains.

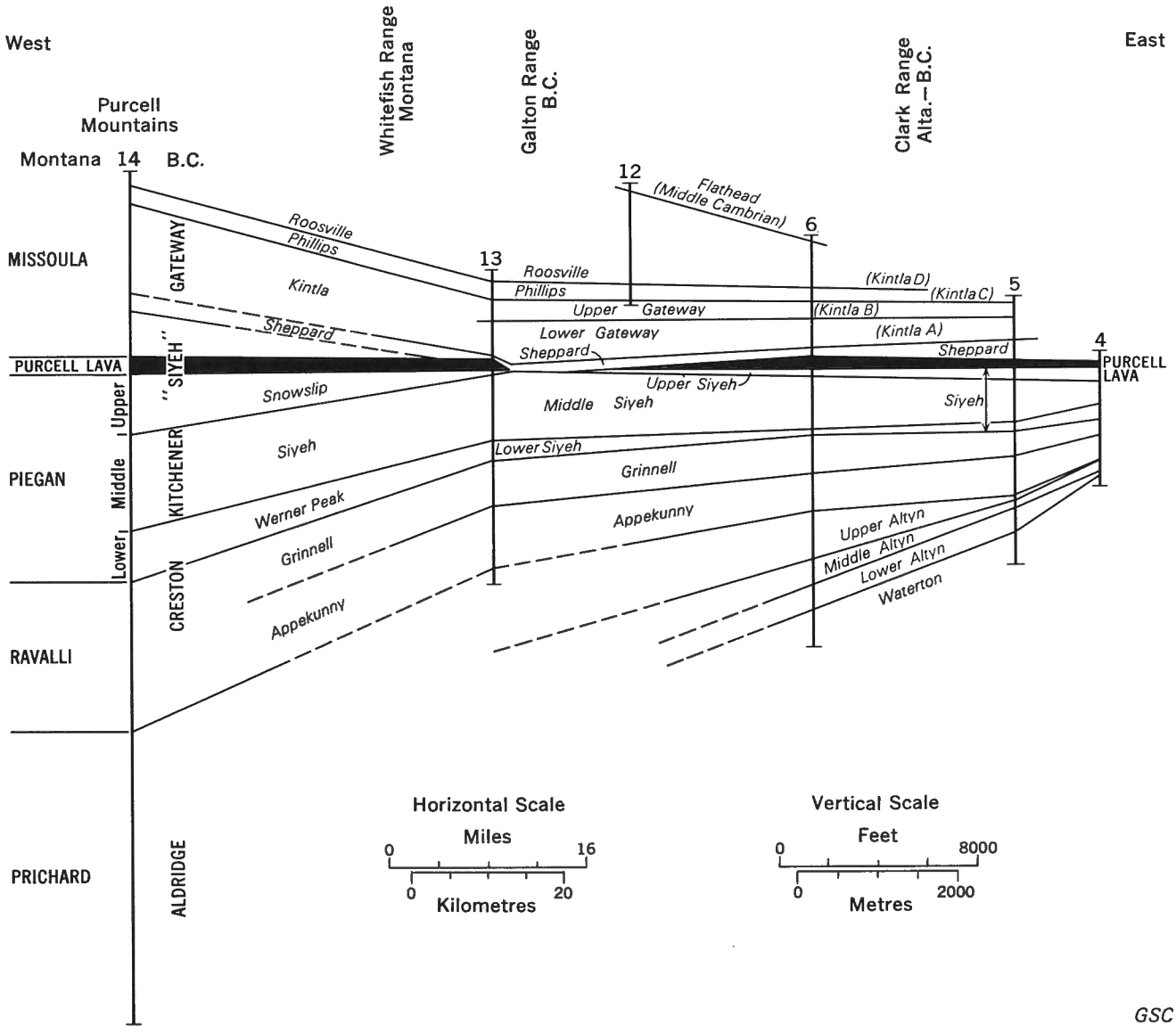


FIGURE VIII-3. Correlation of the Helikian (Purcell) rocks of southern Rocky Mountains and eastern Purcell Mountains, Alberta and British Columbia (Price, 1964a).

In *northern Rocky Mountains* rocks of probable Helikian age are divisible into a lower carbonate and quartzite part, 1,000 feet thick, and an upper pelitic part, 8,000 feet thick (Bell, 1968). The lowest unit in the core of an anticline, the Chischa Formation, is composed of pale grey, pastel-hued, aphanitic dolomite with stromatolites, desiccation cracks, ripple-marks, and "molar tooth" structures, interbedded with quartzite near the base and top. It is unconformably overlain by the Tetsa Formation, of dark siltstones and shales 1,065 feet thick. The grey carbonates of the George Formation are similar to those of the Siyeh Formation of the Purcell. It thins southwestward from 1,750 feet to 1,200 feet and is transitionally overlain by the Henry Creek Formation that thickens from 700 to 1,500 feet in the same direction. The Henry Creek consists of grey calcareous mudstone, siltstone, and conglomerate with pebbles of carbonate similar to the George Formation. The Tuchodi Formation is 5,000 feet of resistant brown-weathering feldspathic quartzite, dolomitic siltstone, and argillaceous dolomite with minor varicoloured shale. The uppermost formations, the Aida and Gataga are mudstone and siltstone with a general flysch aspect. The former is buff to grey weathering and the latter is dark olive-grey with a few poorly sorted and graded sandstones. The succession is intruded by gabbro dykes and overlain with angular unconformity by Lower Cambrian to Silurian formations. The Hadrynian and Helikian of the region are exposed in separate structures.

In *Omineca Mountains* (Roots, 1954) the Helikian may be included in regionally metamorphosed strata of the Tenakihi Group and the Wolverine Complex. The quartz–mica schists and quartzites include minor conglom-

erate with boulders as much as 8 inches in diameter of quartz, plagioclase, and potash feldspar, presumably derived from a granitic or metamorphic source. Part of the Shuswap Metamorphic Complex of Monashee Mountains may also be Helikian.

Helikian rocks near the headwaters of South Redstone River in *Mackenzie Mountains* comprise an apparently conformable succession more than 15,000 feet thick (Gabrielse, *et al.*, 1965). The basal unit, underlain by a thrust fault, is about 4,000 feet of recessive, dark grey and green shale interbedded with fine- to medium-grained quartzite and minor orange- and buff-weathering sandy dolomite. It is overlain by 4,300 feet of pink, white, and purple quartzites containing minor slate and orange-weathering dolomite, succeeded by stromatolitic and oölitic limestones and dolomites and orange-weathering sandy dolomites as much as 6,500 feet thick. A basalt flow or sill up to 100 feet or more thick occurs within the carbonates. In places, the carbonate strata are overlain by, and may be in part stratigraphically equivalent to, pink-weathering gypsiferous siltstones, locally 1,100 feet thick, that contain disseminated copper minerals. East of Keele River beds of gypsum more than 100 feet thick are conspicuous. Locally the siltstones are overlain by 675 feet of limestone, siltstone, slate, and limestone conglomerate. The Helikian strata in Mackenzie Mountains are overlain with angular unconformity by the Hadrynian Rapitan Group or the lower Paleozoic.

In *Franklin Mountains* (Douglas and Norris, 1963) the Helikian is probably represented by 4,800 feet of fine-grained, grey, quartzose sandstone and siltstone interbedded with dusty red and olive-green shale in the upper 3,100 feet and with cryptocrystalline green and brown dolomite in the basal 1,700 feet. The sequence is underlain by a thrust and overlain with moderate angular unconformity by the Lone Land Formation considered to be Hadrynian.

The oldest rocks exposed in *Ogilvie, northern Selwyn, and northern Mackenzie Mountains* are dark weathering slate, argillite, and phyllite, probably more than 5,000 feet thick (Green and Roddick, 1962). In most places the rocks are folded and sheared, mainly as a result of the Hadrynian Racklan Orogeny. The overlying, deformed, but not metamorphosed carbonate succession is variable. The most complete sequence, in southern Ogilvie Mountains, contains a basal unit, about 1,000 feet thick, of dark grey, thinly laminated dolomite, a middle unit of thin-bedded, orange-weathering dolomite, black shale, quartzite, limestone, and argillite, apparently more than 4,000 feet thick, and an upper unit of massive cherty and quartzose grey dolomite about 2,000 feet thick. To the east, in northwestern Selwyn Mountains, basal Helikian strata are slates and argillites overlain mainly by brilliant orange-weathering, thin-bedded dolomites. Still farther east, in northern Mackenzie Mountains, quartzites are dominant.

Helikian strata are present in the lower part of the Tindir Group of *southwestern Porcupine Plateau*. Dolomite, limestone, quartzite, shale, and slate, possibly 5,000 feet thick, are locally intruded by diabase dykes and sills. Succeeding strata of the group are referred to the Hadrynian. The Neruokpuk Formation of *British Mountains* consists of red and green argillite, feldspathic quartzite, shales, and phyllite. It possibly includes Helikian and Hadrynian rocks and also the lower Paleozoic as Silurian graptolites were obtained from Barn Mountains and the formation is overlain with angular unconformity by the Mississippian.

East Kootenay and Racklan Orogenies

Helikian sedimentation was brought to a close in southern Cordilleran Geosyncline by the East Kootenay Orogeny (White, 1959; Leech, 1962a). It is manifested by uplift, gentle folding, tilting, faulting, granitic intrusion, and regional metamorphism to greenschist facies and locally to sillimanite grade. In the southeastern Cordillera Hadrynian Windermere sediments and the Lower or Middle Cambrian rocks rest unconformably on Purcell strata. Fold trends related to the East Kootenay Orogeny are subparallel with those of younger orogenies. Muscovites from three coarse-grained to pegmatitic leucocratic granodiorite stocks in Purcell Mountains have given K–Ar ages of 675, 705, and 745 m.y. Micas from metamorphic rocks have given ages of 710 and 790 m.y. Whether these effects are only part of a much more intense orogeny farther west is unknown.

The Racklan Orogeny brought to a close the distinctive miogeosynclinal phase of Proterozoic sedimentation in the eastern part of the northern Cordilleran Geosyncline. Folding, block faulting, tilting, and uplift of the Helikian rocks resulted in one of the most spectacular unconformities in the northern Cordillera (Pl. VIII-1). In northern Selwyn Mountains Helikian strata form tight north-northeasterly trending folds and are overlain unconformably by the Rapitan Group. In the Taiga Ranges of southern Porcupine Plateau they form tight east-west folds and are unconformably overlain by Cambrian strata. Low-grade regional metamorphism of argillaceous rocks in northern Wernecke, Ogilvie, and Richardson Mountains possibly also took place at this time. In Mackenzie Mountains uplift, tilting, and block faulting followed deposition of Helikian rocks. The base of the Hadrynian Rapitan Group is marked by an angular unconformity, beneath which the degree of truncation varies greatly from one fault block to another. It is probable that the faulting was accompanied by the widespread emplacement of mafic dykes and flows that constitute a large proportion of the boulders in the Rapitan conglomerates.

Hadrynian (Late Proterozoic)

Tectonic Summary

A thick sequence of Hadrynian rocks, dominantly clastic and sedimentary, extends as an almost continuous

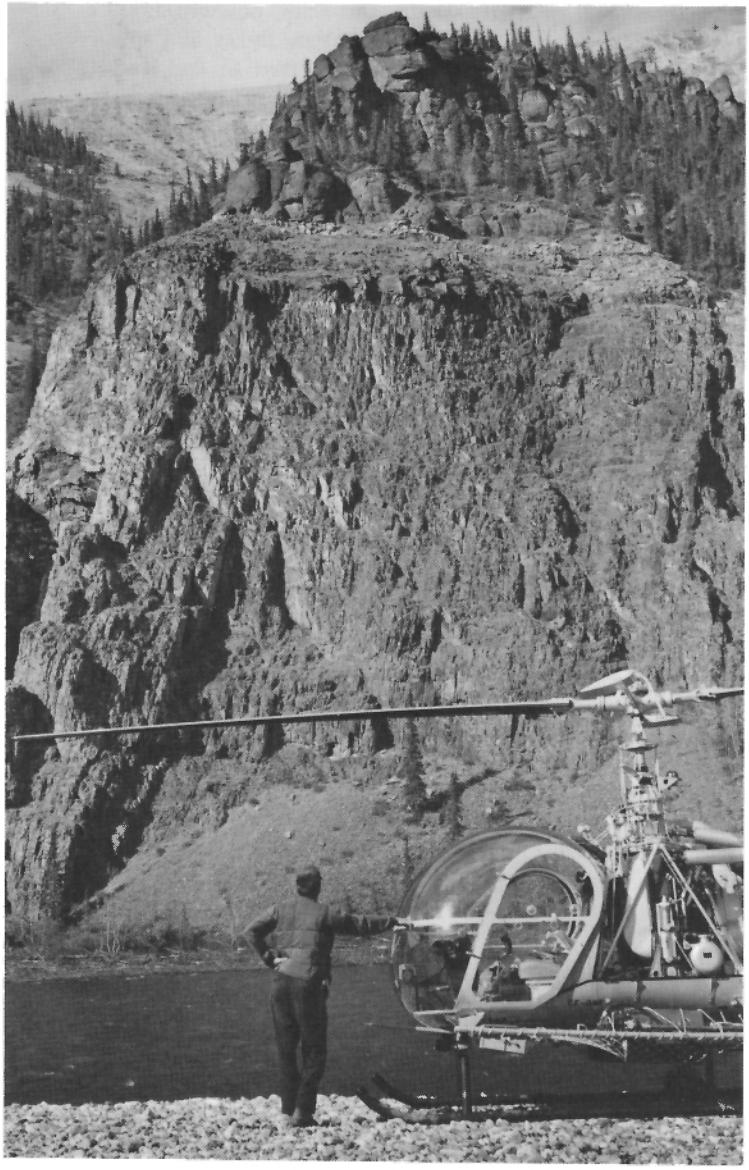


PLATE VIII-1. Angular unconformity on northwestern Mackenzie Arch between flat-lying Upper Cambrian dolomites and steeply dipping, easterly trending Helikian phyllite on Hart River, northern Yukon Territory.

belt through the eastern part of the Cordilleran Geosyncline (Fig. VIII-2). The sediments are of marine and shallow-water origin and probably partly miogeosynclinal, flanking a cratonic mass of Hudsonian crystallines with remnants of Helikian sedimentary cover. The source in the south is from uplift of the southern Purcell Geanticline following the East Kootenay Orogeny, and partly from the east as suggested by sedimentary features of the Miette Group which indicate westward transport. Part is locally derived from fault scarps, particularly the conglomerates of the Rapitan Group in the northern Wernecke and Mackenzie Mountains. The grits and pebble-conglomerates so extensive in the northern Cordilleran Geosyncline contain abundant potash feldspar and point to source areas containing granitic or metamorphic rocks. It remains for future work to determine whether the source was mainly the craton to the east, or a crystalline area to the west that was uplifted following the Racklan and East Kootenay Orogenies.

Cordilleran Geosyncline

Several thousand feet of amygdaloidal basalt, shale, tuff, conglomerate, iron-formation, dolomite, and limestone of probable Hadrynian age form the upper part of the Tindir Group in *southwestern Porcupine Plateau*. Possibly correlative rocks in *Ogilvie Mountains* consist of coarse conglomerate, 300 to 2,000 feet thick, overlain by dark shale and platy sandstone about 1,300 feet thick.

The Rapitan Group in *Mackenzie Mountains* (Gabrielse, *et al.*, 1965) consists typically of as much as 5,000 feet of dark weathering, fine-grained, feldspathic clastic rocks. Bedded chert–hematite iron-formation and poorly sorted conglomerate and conglomeratic mudstone form the lower part. Tuffaceous sediments are common and volcanic flows are present at several localities. Near the headwaters of Redstone River two unconformities, the lower one strongly angular in places, divide the Rapitan into three units: a lower unit, 0 to 1,285 feet thick, of reddish brown, maroon, and green mudstone, conglomeratic mudstone, and iron-formations; a middle unit, 1,215 to 2,705 feet thick, of grey-green to dark grey conglomeratic mudstone; and an upper unit, 1,870 to 3,845 feet thick, of dark shales, siltstones, and fine-grained sandstones. The Rapitan is overlain unconformably by orange-weathering dolomitic sandstones and sandy dolomites having a maximum thickness of 2,500 feet. The uppermost member of this unnamed formation is a widespread buff-weathering dolomite less than 100 feet thick, and locally, a basalt flow, 70 feet thick, occurs just below the top. The uppermost formation is a distinctive sequence of recessive dark weathering shale, siltstone, and sandstone, 2,000 to 3,000 feet thick.

In *Franklin Mountains* the Lone Land Formation, 965 feet thick, consists of grey shale with a basal ferruginous quartzose conglomeratic sandstone 200 feet thick that rests with angular unconformity on the Helikian and is overlain also with angular unconformity by the Lower Cambrian.

Large areas in *Selwyn Mountains* and northern *Rocky Mountains* (Bell, 1968) are underlain by a great thickness of impure clastic sediments consisting mainly of argillite, phyllite, slate, and siltstone but characterized by beds of gritty, feldspathic sandstone or feldspathic pebble-conglomerate, and by the presence of maroon, red, and green slates with minor limestone in the upper part. These rocks, informally termed the 'grit unit' are similar to Kaza, Miette, and Windermere strata of the southern Cordillera. The base of the sequence has not been observed and in most places no definite break with Lower Cambrian strata has been recognized.

The oldest exposed rocks in *Cassiar Mountains* (Gabrielse, 1963) are the Good Hope Group which consists of typical 'grit unit' rocks overlain by several thousand feet of well-bedded limestones, locally pisolitic and oölitic and commonly containing particles of ferrodolomite. The uppermost strata include crossbedded, sandy limestones, calcareous sandstones, and red and green shales and slates, overlain apparently conformably by Lower Cambrian.

In *Cariboo Mountains,* the Kaza Group of quartzose and feldspathic sandstone, grit, and phyllite is overlain by the argillaceous Isaac Formation, in turn succeeded by the Cunningham Limestone (Sutherland Brown, 1957). Limestones mapped as the Cunningham form three belts, the western two of which merge along a fault. Only the central limestone contains fossils and where stratigraphic relations appear unequivocal the fossiliferous limestone lies more than 4,500 feet above the unfossiliferous limestone. Thus, the eastern and western belts of limestone may be either unfossiliferous Lower Cambrian or equivalent to the carbonate member at the top of the Hadrynian Miette Group in the Rocky Mountains (Campbell, 1967).

In *Purcell and westernmost Rocky Mountains,* the Toby Formation forms the base of the Windermere. It is extremely variable in thickness, up to 2,000 feet, and also in lithology, being either a breccia or a conglomerate with a great range in ratio of matrix to clasts. Near the Rocky Mountain Trench, gneissic granite boulders as much as 2 feet in diameter are at the top. Reesor (1957) concluded that the material must have had a source nearby, probably Archean rocks, now buried beneath Paleozoic cover in the Rocky Mountains. The formation has been considered a fanglomerate by J. F. Walker but H. M. A. Rice thinks that it was deposited along the shoreline of a sea that spread rapidly over a subsiding landmass reworking loose surficial deposits. The Toby is conformably succeeded by the Horsethief Creek Group. The basal unit, 3,000 feet thick, is principally of quartzose and feldspathic grit, sandstone, and some varicoloured quartz-pebble conglomerate. It is overlain by a grey, green, and purple slate more than 1,000 feet thick, and succeeded by a limestone 650 feet thick. A unit of slate and quartzite at the top varies in thickness from zero to 1,000 feet. The group thickens westward and becomes generally finer grained and more abundant in carbonate in Selkirk Mountains and in the northern Purcell Mountains and Dogtooth Range (Reesor, 1957). Southwest of Kootenay Lake, the Irene Formation overlies the Toby Formation conformably. The Irene is a fine-grained, sheared, andesitic greenstone with interbeds of conglomerate and limestone. Considerable thicknesses of breccia and minor calcareous tuff may be present locally.

In *southern Rocky Mountains* the Miette Group is composed of a basal unit, 3,000 feet thick, of argillite and argillaceous sandstone, succeeded by 2,000 feet of sandstone, grit, conglomerate, and argillite, which is overlain by 3,000 feet of argillite and, locally at the top, by a carbonate 0 to 1,000 feet thick (Mountjoy, 1962). The Hector and Corral Creek Formations farther south are lithologically similar to the Miette and Horsethief Creek Groups.

Hadrynian rocks are possibly present in the regionally metamorphosed Yukon and Horseranch Groups and the Wolverine and Shuswap Complexes.

PALEOZOIC

Cambrian

Tectonic Summary

The Lower Cambrian (Fig. VIII-4) is restricted to the Cordilleran Geosyncline, and is generally thickest in linear troughs marginal to the craton. It is of miogeosynclinal facies. Thick extensive sheets of orthoquartzite and conglomerate overlain by limestone, commonly archeocyathid-bearing, grade westward into thinner calcareous shales and argillaceous limestones. The orthoquartzites of Alberta Trough are locally feldspathic and were derived from the Hudsonian crystallines of the craton or possibly its former Helikian cover. Their maturity relative to the underlying Windermere clastics, with which they may be locally conformable, suggests reduction of relief and a general tectonic stability of the craton by the beginning of and during the Early Cambrian. Northeast-trending Eager Trough, possibly bordered on the southeast by the Moyie fault, separates Purcell Arch from an emergent block to the south (Montania). Quartzose sandstone and conglomerate derived from adjacent uplifts transgress the arch and lie unconformably on Purcell and Windermere strata. Within Lardeau Trough the Lower Cambrian reaches a thickness of 9,000 feet and contains intercalated basic flows and pyroclastics.

Several tectonic elements of the northern Cordilleran Geosyncline, more evident in some later periods, appear to be established in the Early Cambrian. Polymictic conglomerates on the west flank of the MacDonald Platform were derived from a west-dipping homocline of Helikian sediments, the uplift bounded, in part at least, by faults active during deposition. Shallow-water orthoquartzite and limestone on the Cassiar Platform contrast with thick, deeper water, dominantly pelitic sediments in eastern Selwyn Basin, the latter grading easterly to shallow-water sandstone and carbonates of the western flank of Mackenzie Arch. On the arch they are absent through convergence and post-Lower Cambrian truncation. *Olenellus-*

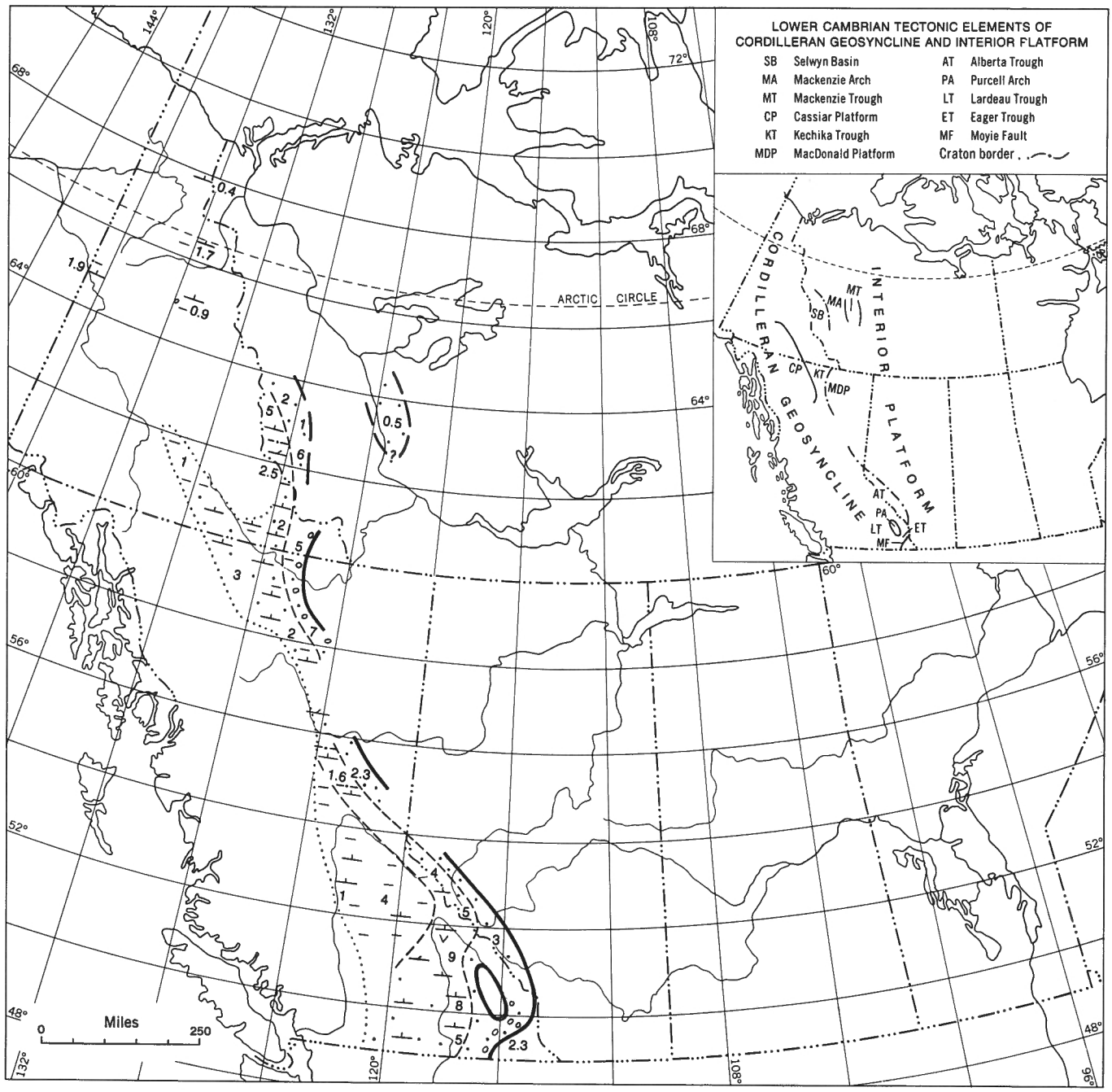
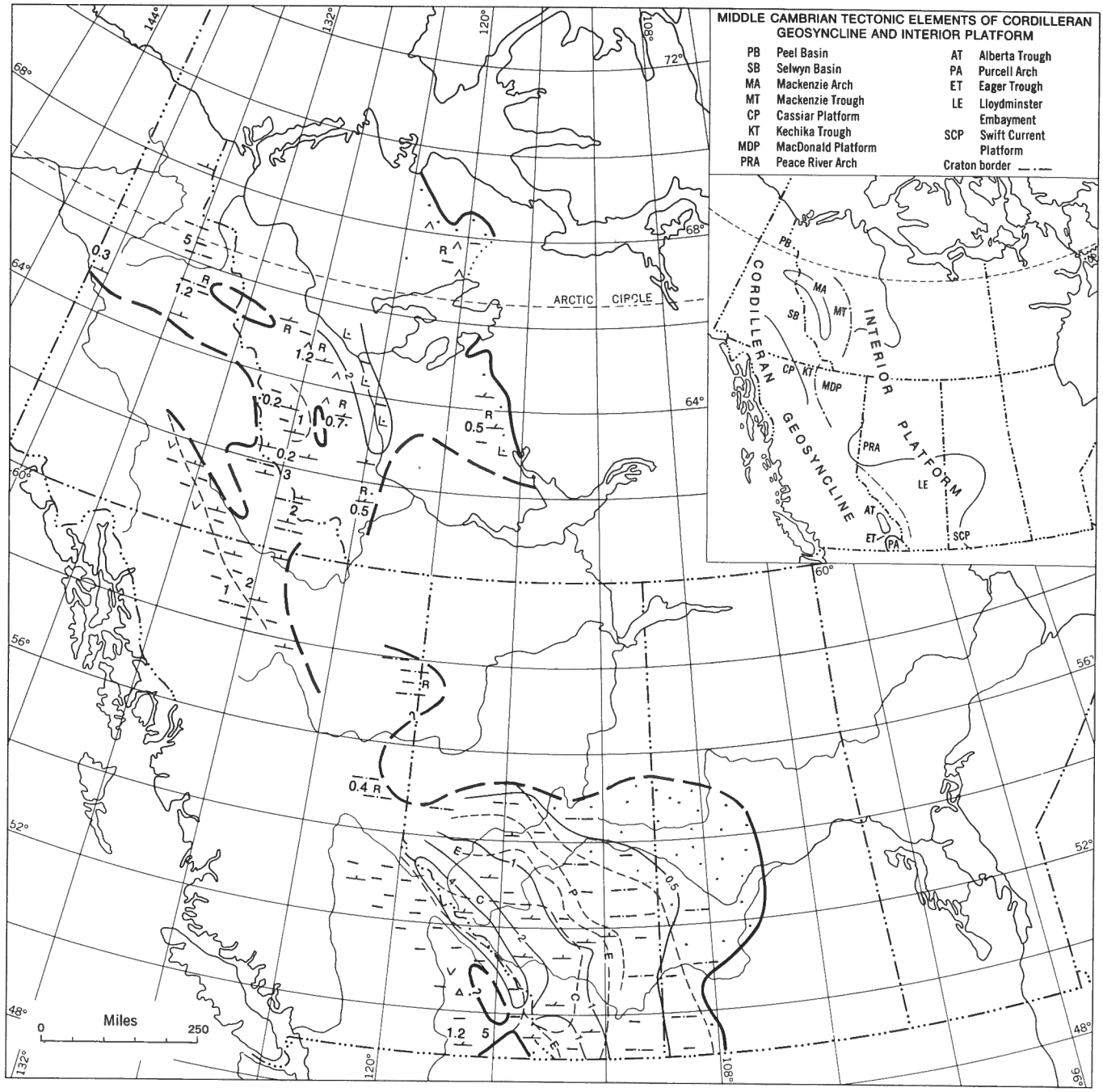


FIGURE VIII-4. Lower Cambrian sedimentation and tectonism in Western Canada.



DEPOSITIONAL FEATURES

Sandstone
Greywacke
Volcanics
Siltstone
Shale
Limestone
Dolomite
Anhydrite
Salt
Red beds

Depositional limit
Truncated limit
Facies boundary
Pika Formation
Eldon Formation
Cathedral Formation
Outcrop limit
Isopach (thousands of feet)
Thickness (thousands of feet)

PHYSIOGRAPHIC FEATURES

Blank areas are lowlands and plains developed on crystalline and sedimentary rocks of the platform and where Middle Cambrian is now absent through erosion, or unknown. In the geosyncline the blank areas may be sea.

FIGURE VIII-5. Middle Cambrian sedimentation and tectonism in Western Canada.

bearing limestone occurs in Peel Basin and orthoquartzites with overlying green shale in Mackenzie Trough; these strata lie unconformably on the Proterozoic.

Middle Cambrian sediments (Fig. VIII-5) in the southern Cordilleran Geosyncline conformably overlie the Lower Cambrian in the deeply depressed Lardeau and Alberta Troughs and overlap parts of Purcell Arch and the depressed margin of the craton. Lardeau Trough received shale, thin beds of limestone and quartzite, with intercalated basic pillow lavas and pyroclastics. In Alberta Trough several thousand feet of massive carbonate grade westward into thick shale and argillaceous limestone which in turn grade into westerly thinning carbonates that lie unconformably on the Proterozoic along the eastern edge of Purcell Arch. Shale was probably deposited in Eager Trough, the land to the south covered in part by thin sandstone, shale, and dolomite overlapping from the east, and fringed on the west by thin carbonates. East of the strong hingeline bounding the craton (Fig. VIII-7) Middle Cambrian rocks occupy Lloydminster Embayment, a broad northeast-trending depression within which the carbonates progressively grade eastward to shale and siltstone and basal overlapping sandstone and conglomerate. The depression is subparallel with the grain of the Hudsonian crystalline basement and is flanked on the southeast by Swift Current Platform and on the northwest by Peace River Arch. Thin Upper Cambrian (Fig. VIII-6) shaly siltstones and sandstones overlap the Middle Cambrian and Precambrian crystallines of Swift Current Platform. They extend to truncated margins and are deeply eroded by post-Lower Ordovician uplift along the margin of the craton that later developed as Alberta Arch. Thick, bedded carbonates occur in Alberta Trough, the upper part intertonguing westward with shale and the lower part, more persistent westward, merges with the late Middle Cambrian carbonates along the east flank of Purcell Arch.

In the central part of the craton, Middle and Upper Cambrian rocks are absent probably mainly through nondeposition. Thin undated shales occur west and north of Peace River Arch, but it is not known whether a depositional trough existed between the craton and MacDonald Platform, where Middle and Upper Cambrian rocks are missing, joining Alberta and Mackenzie Troughs. In southern Kechika Trough thick calcareous phyllite and argillaceous limestones grade westerly into calcareous phyllite and dark shale on Cassiar Platform. Thick carbonates extend along the southern and western flanks of Mackenzie Arch. Along the crest of the arch the Upper Cambrian is absent and the Middle Cambrian locally absent, mainly through pre-Upper Ordovician truncation. The Middle Cambrian on the arch consists of a thin red-bed facies of shale, sandy dolomite, and gypsum with sandstone and conglomerate in the west, where unconformably overlain by Upper Cambrian limestones that thicken northward into argillaceous limestone and shale in Peel Basin. In Mackenzie Trough the Middle Cambrian is mainly salt and the Upper Cambrian is dolomite. An undated red-bed facies borders the Canadian Shield but is absent on Coppermine Arch; it is Middle Cambrian east of the arch.

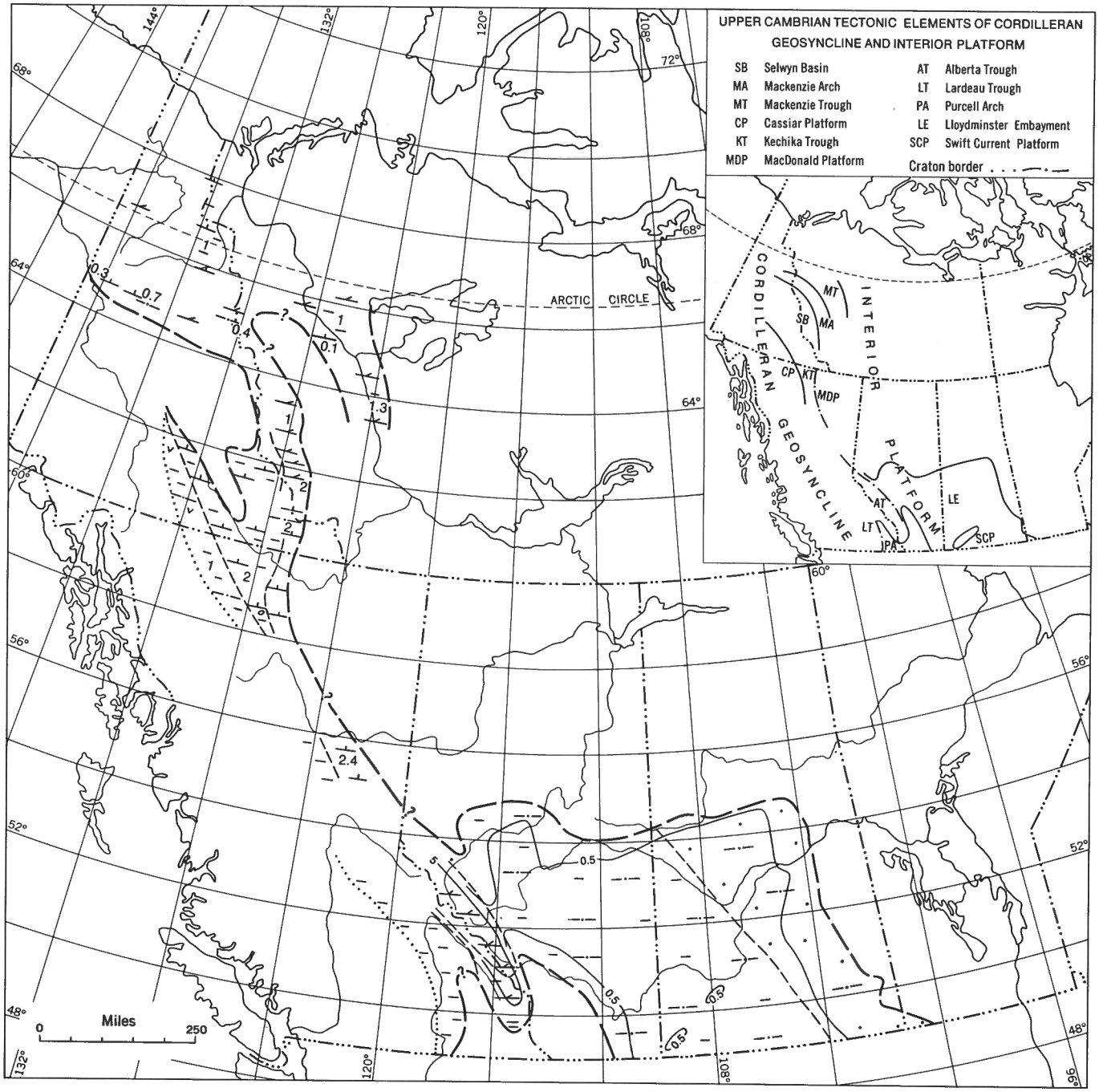
Lower Cambrian

In *Richardson Mountains* R. M. Proctor reports well-bedded grey limestones, 1,200 feet thick, with *Olenellus.* They are overlain, apparently conformably, by rocks of probable Middle Cambrian age, but their base is not exposed (Brabb, 1967). Less than 1,000 feet of limestone and limestone-conglomerate outcrops along the Alaskan boundary north of Yukon River. These rocks are overlain by 900 feet of argillite, siltstone, and minor quartzite and sandy limestone, in part of Early Cambrian age. In northern *Wernecke Mountains* about 900 feet of pelletoid limestone with interbedded reddish weathering sandstone and conglomerate lie unconformably on the Helikian. In *Franklin Mountains* 500 feet of fine-grained, white quartzose sandstone of the Mount Clark Formation lies with angular unconformity on Proterozoic strata, and is conformably overlain by 200 feet of grey, green, and red shales of the Mount Cap Formation.

The Lower Cambrian is as thick as 6,000 feet in eastern *Selwyn Mountains* and western *Mackenzie Mountains* (Gabrielse, *et al.*, 1965) but decreases eastward to zero as the result of depositional thinning on Mackenzie Arch and truncation beneath the regional unconformity at the base of Middle Cambrian rocks. In western Mackenzie Mountains the Lower Cambrian is mainly sandstone but includes minor pebble-conglomerate, siltstone, and shale. The base is probably a regional unconformity with little angular discordance. In northeastern Selwyn Mountains, limestones, dolomites, sandy dolomites, sandstones, and argillites, characteristically buff, mauve, cream, and brown weathering, range in thickness from 1,600 to 3,200 feet. In southeastern Selwyn Mountains the Lower Cambrian sequence consists of: calcareous siltstone and impure limestone, 200 feet; limestone, 200 feet; sandstone, sandy and silty dolomite, 2,000 feet; orange-weathering silty and sandy dolomite, 1,000 feet (Blusson, 1968). Basic volcanic flows and breccias, locally 200 feet thick occur in the upper part of the member below the orange-weathering beds.

In *Pelly Mountains* and *Cassiar Mountains* the Lower Cambrian Atan Group is represented by a basal sandstone or quartzite 1,000 to more than 2,000 feet thick, a middle shale about 150 feet thick, and an upper limestone unit as much as 1,500 feet thick (Gabrielse, 1963). Typically the limestones are massive, relatively pure, partly pelletoid, and yield abundant well-preserved archeocyathids. Strata of the Atan Group are structurally conformable with the underlying Proterozoic rocks and the overlying Middle or Upper Cambrian sequence.

In *northern Rocky Mountains* is a highly variable sequence of clastic and carbonate rocks, according to G. C. Taylor, more than 5,000 feet thick locally. In the east, the base rests, with marked angular unconformity (Pl.



DEPOSITIONAL FEATURES

Conglomerate . . . . . o
Siltstone . . . . . —·—
Shale . . . . . —
Volcanics v
Limestone . . . . . ⊥
Dolomite . . . . . ⊥

Truncated limit . . . . . ⌒
Facies boundary . . . . . ⌒
Outcrop limit . . . . . ·····
Isopach (thousands of feet) 0.5
Thickness (thousands of feet) 0.4

PHYSIOGRAPHIC FEATURES

Blank areas are lowlands and plains developed on crystalline and sedimentary rocks of the craton, and where Upper Cambrian is now absent through erosion, or unknown. In the geosyncline the blank areas may be sea.

GSC

FIGURE VIII-6. Upper Cambrian sedimentation and tectonism in Western Canada.

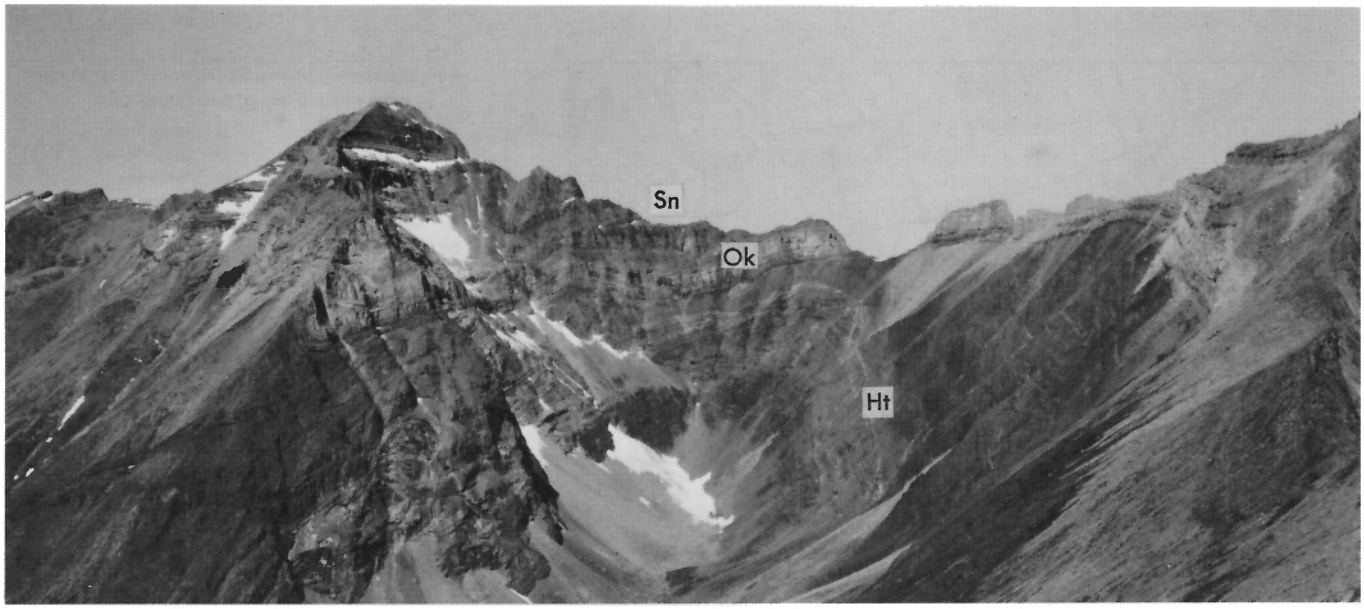


PLATE VIII-2. Angular unconformity on MacDonald Platform between the Ordovician and Helikian, Yedhe Mountain near Alaska Highway, northern Rocky Mountains, British Columbia. Looking northwest along axis of anticline produced in Racklan Orogeny. Sn—Silurian Nonda Formation; Ok—Lower Ordovician Kechika Group; Ht—Helikian Tuchodi Formation.

VIII-2), low in the Helikian succession, whereas a few miles to the west no apparent break separates the Cambrian and Hadrynian strata. Thick, coarse, red-weathering conglomerates with pebbles of quartzite, dolomite, and greenstone are interbedded with and grade westward into sandstone, shale, and argillaceous limestone. The conglomerate occurs adjacent to northeast- and northwest-trending faults bordering the west flank of MacDonald Platform and demonstrates Lower Cambrian faulting. In Terminal Range to the west, a widespread conglomerate, 50 feet thick, containing slabs of archeocyathid limestone occurs within a shale and calcareous sandstone sequence. Relatively minor, coarse, polymictic conglomerates also occur locally within limestones and dolomites immediately east of Rocky Mountain Trench. The limestones are underlain by sandstone and in general aspect the sequence is much like that of Cassiar Mountains except for the presence of well-bedded oölitic and sandy dolomites. In southern Rocky Mountains, in Alberta Trough, as much as 7,000 feet of quartzite, conglomerate shale, and limestone comprise the Gog Group (Aitken, 1968). A limestone, the Mural Formation, occurs near the top and thins to the south and east.

Lower Cambrian strata occur in isolated remnants and fault blocks as part of thick lower Paleozoic sections northwest of Moyie fault in *Purcell Mountains* and north of Dibble Creek fault in Hughes Range east of Rocky Mountain Trench. The basal Cranbrook Formation, a quartzite and quartzitic conglomerate, lies unconformably upon various units of the Purcell System. Northwest of Moyie fault it contains coarse fragments of the Creston Formation. About 20,000 feet of strata was eroded from Purcell Arch prior to overlap by the Cranbrook westward from Alberta Trough and northward from Eager Trough (Reesor, 1957). The Cranbrook locally includes a middle unit of magnesite. The succeeding Eager Formation contains late Lower Cambrian fossils in the lower part. It is probably more than 6,000 feet thick near Cranbrook and indicates that considerable subsidence occurred in Eager Trough during the Cambrian. However, south of Moyie and Dibble Creek faults, the Lower Cambrian is absent and the Upper Devonian lies unconformably on the Purcell although locally south of Dibble Creek fault some Middle Cambrian beds intervene. The rapid northwesterly truncation of Purcell strata beneath the sub-Cambrian unconformity (Norris and Price, 1966) and the presence of coarse debris in the Lower Cambrian near Moyie fault indicate that Eager Trough was probably bounded by an uplifted block (Montania) during Early Cambrian time.

In *Selkirk* and *western Purcell Mountains* the Hamill Group generally lies gradationally and conformably upon the Horsethief Creek Group. Fossils have been found in the Hamill Group at only one locality, in Dogtooth Range, where mid-Lower Cambrian *Callavia* occurs in the uppermost 20 feet. The delineation of the base of the Cambrian in this region and adjacent metamorphic terrane is therefore a problem. The base of the Hamill Group is used rather than, alternatively, the first unconformity below Cambrian fossils, which here would be the base of the Hadrynian Windermere System. In Dogtooth Range the Hamill Group consists of two quartzite formations separated by an argillaceous unit. It thins southeastward towards Purcell Arch from 4,000 feet to 1,500 feet and overlies the Horsethief Creek Group with erosional unconformity. The Hamill Group is overlain conformably

by the Donald Formation, 1,500 feet of sandstone, slate, limestone, and at least one flow of basalt. The Donald contains archeocyathids, *Olenellus,* and *Bonnia.* Near Kootenay Lake, the lower two-thirds of the Hamill Group, the Quartzite Range, and Mount Gainer Formations are composed mainly of thick-bedded, commonly crossbedded orthoquartzite locally containing sodic plagioclase and potash feldspar. In northern Selkirk Mountains the group contains abundant greenstone flows, breccia, and tuff, and rare stretched quartzitic conglomerate. The upper third of the group, the Reno and Marsh Adams Formations, is thinly bedded and contains much argillaceous sediment. The Hamill Group is overlain conformably by phyllite and limestone of the Mohican Formation (Fyles and Eastwood, 1962) and then by archeocyathid-bearing limestone of the Badshot Formation. The equivalent archeocyathid-bearing beds of the Laib Formation in Salmo area are also limestone (Little, 1960). The group thickens westward from about 4,000 feet in northern Dogtooth Range to about 8,000 feet in the Selkirks but thins westward from 4,500 at Kootenay Lake to 2,000 feet 30 miles southwest near the International Boundary.

The presence and thickness of the Hamill Group in the Shuswap Metamorphic Complex is uncertain but it is probably represented by a highly deformed, relatively pure quartzite (Wheeler, 1965). The quartzite thins northwestward and is apparently represented in the *Cariboo Mountains* by the Yanks Peak Quartzite, 600 feet in the southeast and 1,600 feet in the northwest. The Yanks Peak Quartzite is overlain by 500 feet of shales of the Midas Formation succeeded by archeocyathid-bearing limestone similar to the Mural Formation of the Gog Group of Rocky Mountains. The Yanks Peak, which therefore must be a tongue of quartzite of the Gog Group, is underlain by 2,500 feet of shales of the Yankee Belle Formation, which in turn overlies the Cunningham Limestone (Campbell, 1967).

In *Omineca Mountains* archeocyathid-bearing limestone has been included in the Ingenika Group together with orthoquartzite of possible Early Cambrian age (Roots, 1954). The remaining lower part of the group, consisting of grit, quartzite, phyllite, and schist, is like the Horsethief Creek Group and presumed Hadrynian.

Middle and Upper Cambrian

In *southern Richardson Mountains* about 5,000 feet of platy and nodular limestone, siltstone, and shale, probably mainly Middle Cambrian, are overlain by more than 1,200 feet of platy limestone and argillite with an Upper Cambrian, Dresbachian fauna in its lower part. Along the Alaskan Boundary north of Yukon River the Middle Cambrian is represented by the upper part of a sequence 900 feet thick of argillite, siltstone, and minor quartzite and sandy limestone overlain by 400 feet of limestones and limestone-conglomerates partly of Upper Cambrian age. In *Taiga Ranges* and *Wernecke Mountains* well-bedded, commonly buff-weathering limestones and dolomites of Middle and Late Cambrian age are generally 1,000 to 2,000 feet thick and overlie Proterozoic rocks with regional unconformity.

In the *Franklin Mountains* the Middle Cambrian is probably represented by salt and green, red, and grey shales of the Saline River Formation. A thickness of 2,783 feet was penetrated by the Imperial Vermillion Ridge No. 1 well. H. G. Bassett considers the Franklin Mountain Formation to be of Late Cambrian age. It consists of 1,300 feet of light brown weathering, fine-grained, dull dolomite with minor variegated shale. The McDougall Group in *northern Mackenzie Mountains* is about 1,000 feet thick and consists of green, brown, and maroon shales and sandstones with gypsum and grey limestone in the upper part. The formation has yielded a single fossil of Middle or Late Cambrian age and rests with regional unconformity on quartzites of the Helikian Katherine Group (Hume, 1954).

Throughout much of *western Mackenzie Mountains* Middle Cambrian strata overlie older rocks with regional unconformity. Their distribution and thickness are also strongly influenced by major regional unconformity at the base of the Upper Cambrian. Near the headwaters of Redstone River the Middle Cambrian sequence comprises recessive, thin-bedded, argillaceous limestones, siltstones, and shales as much as 1,500 feet thick structurally conformable with underlying Lower Cambrian sandstones. Southeast of Sekwi Mountain the Middle Cambrian succession is reduced to only a few hundred feet. Franconian fossils have been collected from basal beds above the unconformity at several localities. A conspicuous silver-grey-weathering sandstone or dolomitic sandstone locally more than 300 feet thick occurs at or near the base of the Upper Cambrian sequence and is overlain by several hundred feet of buff- and orange-weathering limestone and dolomite conformable with overlying Lower Ordovician strata. The Upper Cambrian rocks are truncated on Mackenzie Arch to the east beneath the regional unconformity at the base of the Upper Ordovician.

In *southwestern Mackenzie Mountains* northeast of South Nahanni River, buff-weathering Middle and Upper (?) Cambrian rocks including dolomitic siltstones, silty dolomites, and stromatolitic dolomites and limestones are at least 1,500 feet thick and are overlain by massive limestones and dolomites of Late Cambrian age at least 600 feet thick (Gabrielse, *et al.,* 1965).

About 2,000 feet of well-bedded silty limestone of Middle Cambrian age in *eastern Selwyn Mountains* near the big bend in South Nahanni River is bevelled southerly to zero beneath Upper Cambrian (Franconian) strata. The basal Upper Cambrian rocks are maroon weathering sandstones 200 feet thick and are overlain by grey weathering carbonates also of Late Cambrian age that grade upward into Lower Ordovician carbonates. Near the Canada Tungsten mine and to the south Middle and Upper (?) Cambrian to Lower Ordovician silty limestones are more than 4,000 feet thick. They are commonly

argillaceous in the upper part. The base of the sequence is marked by a regional unconformity, or, locally by an angular unconformity.

In *Pelly Mountains* thin-bedded and recessive glossy calcareous shales and argillaceous limestones are dated only by their stratigraphic position between the Lower Cambrian and Lower Ordovician (Green, *et al.*, 1960a, 1960b). Locally minor green volcanic breccia and tuff are intercalated. Middle (?) and Upper Cambrian rocks in *Cassiar Mountains* are dominantly calcareous in the east and argillaceous in the west. In *Kechika Ranges* several thousand feet of thin-bedded argillaceous limestones and calcareous phyllites locally including intrusive bodies of diabase and minor bodies of green volcanic breccia appears to lie conformably between Lower Cambrian and Lower Ordovician rocks (Gabrielse, 1962b). Farther west rocks in the same stratigraphic interval are in part of the Upper Cambrian (Trempeleau). They are dominantly dark weathering, thin-bedded shales and phyllites, variably calcareous and about 1,000 feet thick (Gabrielse, 1963).

In Terminal Range of *northern Rocky Mountains* as much as 1,000 feet of well-bedded calcareous sandstone overlies archeocyathid-bearing conglomerate and is in turn overlain by a thick assemblage of thin-bedded, nodular, argillaceous limestone mainly of early Ordovician age. The latter assemblage may include strata of Cambrian age, particularly in westernmost exposures.

The Middle and Upper Cambrian rocks of *Purcell Mountains* were deposited on the eastern flank of Purcell Arch which probably remained emergent until some time in latest Early or Middle Cambrian time. They occur at intervals within Rocky Mountain Trench, and along its west side, consisting of the unfossiliferous dolomite of the Jubilee Formation and the lower part of the overlying thin-bedded limestone and shale of the McKay Group (Reesor, 1957). West of the trench the two formations total no more than 300 feet whereas to the east they thicken to nearly 5,000 feet and fossiliferous mid-Upper Cambrian strata overlie the Jubilee. On both sides of the trench, near Columbia Lakes, the Jubilee lies disconformably on the Horsethief Creek Group, but to the northwest and southeast it lies on late Lower Cambrian strata.

In southernmost *Selkirk Mountains* fossiliferous Middle and Upper Cambrian strata in Lardeau Trough are the limestone, dolomite, and calcareous shales of the Nelway Formation (Little, 1960) and its equivalent in Washington, the Metaline Formation (Park and Cannon, 1943). These rocks accumulated as carbonate banks fringing the southern part of Purcell Arch and the uplifted

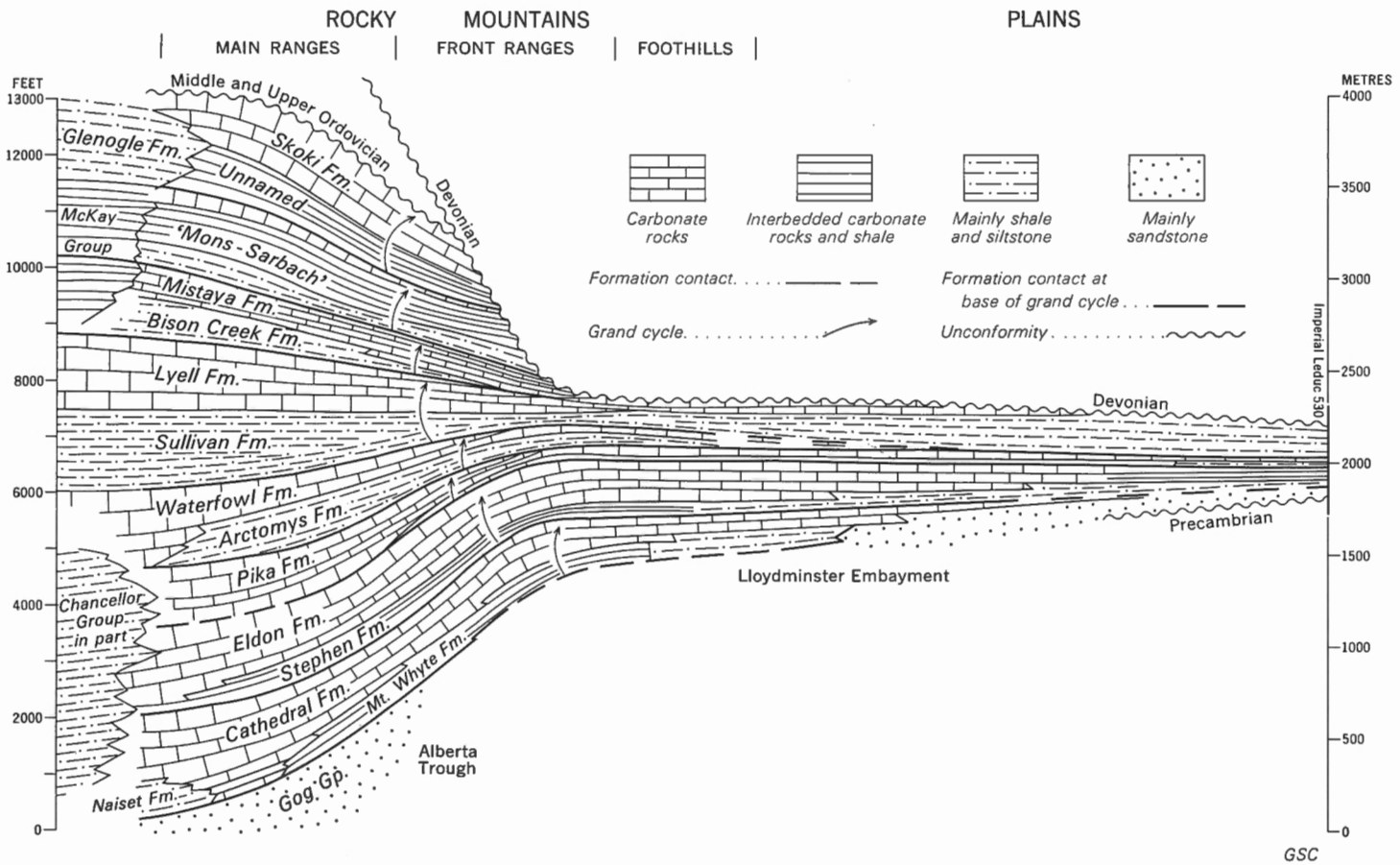


FIGURE VIII-7. Schematic restored stratigraphic section of Cambrian and Ordovician rocks, southern Rocky Mountains and Interior Plains of Alberta and British Columbia (Aitken, 1966).

block south of Eager Trough. The Nelway Formation lies gradationally and conformably upon the Laib Formation and is overlain with structural conformity by the Active Formation. The Nelway Formation contains no diagnostic fossils but the Metaline has yielded two Middle Cambrian faunas, the higher of which resembles the one in the Stephen Formation. Elsewhere in Lardeau Trough the fossiliferous Lower Cambrian limestone is conformably succeeded by the unfossiliferous, predominantly argillaceous Index Formation (Fyles and Eastwood, 1962) that also contains thin beds of limestone, quartzite, and amygdaloidal and pillowed greenstone flows and breccia.

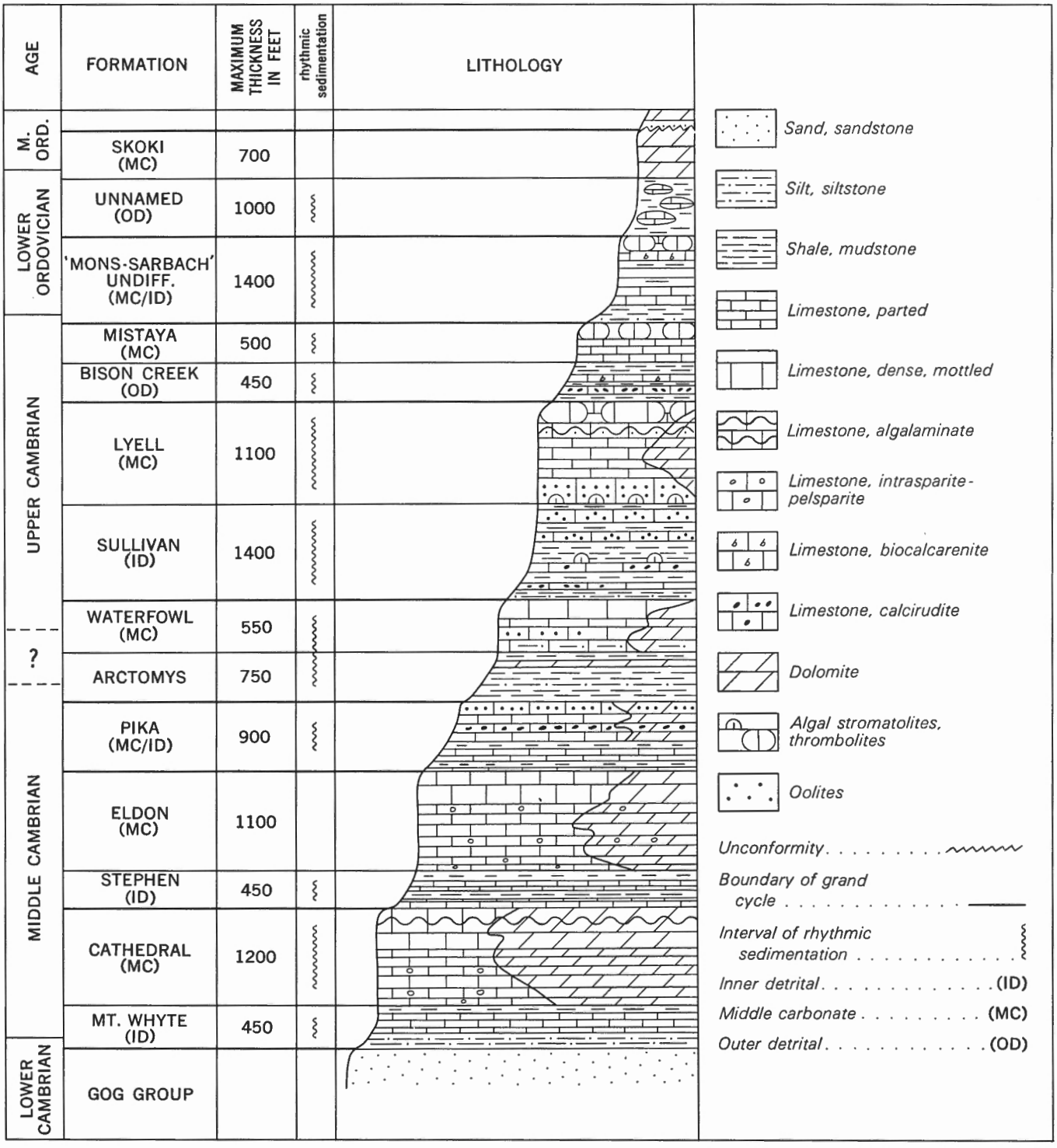


FIGURE VIII-8. Composite stratigraphic column of Middle Cambrian to Middle Ordovician rocks of southern Rocky Mountains (Aitken, 1966).

Cambrian Lithofacies of Alberta Trough and Lloydminster Embayment

The Cambrian rocks of Alberta Trough and Lloydminster Embayment (Aitken, 1966) record continuous marine conditions of deposition that began in the early Cambrian or before and continued into the Ordovician. During this interval, as the margin of the craton was depressed to form Lloydminster Embayment, the seas gradually trangressed eastward. No major unconformities are known to interrupt the deposits produced during this prolonged transgression, but a series of lesser regressions and transgressions also occur so that the second order and even higher order movements are evident as cycles or alternations of facies (Fig. VIII-7). Typically, in the carbonate succession for example, each cycle begins with the abrupt appearance of shale and siltstone lying with sharp contact upon carbonates, continues as interbedded clastic and carbonate, and ends with a sequence of uninterrupted carbonates. The clastic strata that were derived from the craton to the east and north generally reflect conditions of rapid regression and shallowing of the sea whereas the building of the carbonate banks reflects a gradual transgression. Many formation boundaries coincide with the beginning and mid-points of the cycles, eight of which are recognized in the Middle and Upper Cambrian (Fig. VIII-8).

Three major facies belts arose during the major transgression, each of which is also characteristic of a general tectonic environment; an inner detrital belt of thin sandstone and siltstone that occupies most of Lloydminster Embayment; a middle carbonate belt of thick carbonate and shale mainly in Alberta Trough but extending also into the embayment; and an outer detrital belt occupying the axis of Alberta Trough and extending westward to the flank of Purcell Arch. However, as a result of the second order cycles, the dominant facies of each belt also occurs in the other belts.

The inner detrital belt of Lloydminster Embayment is characterized by shallow-water and near-shore clastic facies. Those of Middle Cambrian age are unnamed and those of the Late Cambrian comprise the lower part of the Deadwood Formation which is lithologically inseparable from the upper part of Early Ordovician age. The Deadwood reaches a maximum thickness of 1,100 feet, some 500 to 700 feet of which are assigned to the Upper Cambrian. The basal or near-shore facies of Middle Cambrian age are glauconitic quartzose sandstone and fine conglomerate with minor shale, grading into offshore facies of fissile, dark green and maroon shale and thin-bedded glauconitic siltstone with subordinate high glauconitic fine sandstone and green sands. Mudcracks are prevalent. The sandstones and siltstones are commonly intensively reworked by burrowing animals and are ripple-marked and crossbedded. The Upper Cambrian shales, sandstones, and siltstones are similar to those of the Middle Cambrian. The Middle Cambrian rests directly on Precambrian crystalline rocks and is overlapped by the Upper



PLATE VIII-3. Bedding surface with algal mounds in Upper Cambrian Mistaya Formation, Chaba Creek, southern Rocky Mountains, Alberta.

Cambrian on the south and north sides of the embayment. Numerous knobs on the Precambrian surface, particularly on Swift Current Platform, cause local thinning and irregular isopach trends. On part of the north side of the embayment the Upper Cambrian is less extensive than the Middle Cambrian, but the relationship of facies trends to truncated edges suggests that the original depositional margin was subparallel with the present margin and that Peace River Arch probably existed during the Cambrian, forming the northern flank of Lloydminster Embayment. Subsequently uplift and erosion also took place.

In western Lloydminster Embayment clastic facies alternate with carbonate tongues, the sequence reaching a maximum thickness of 1,500 feet. The carbonates represent the easternmost extremities of the Cathedral, Eldon, Pika, and Lynx Formations and are predominantly the dense limestone facies. The clastic units are composed mainly of fissile shale and greenish grey to brown, predominantly calcareous mudstone, interbedded with thin siltstone beds, particularly near the base, and with coarsely particulate and glauconitic limestones, pelmatozoan–trilobite calcarenite, oölite, and conglomerate near the top.

The middle carbonate belt of Alberta Trough in the Middle Cambrian is dominated by dense limestones, largely the micrite, pelsparite, and intrasparite types (Pl. VIII-2). Where thinly bedded, the limestones are characterized by partings, and where not bedded by mottlings of tan-weathering, argillaceous, very fine crystalline dolomite. Pale birdseye limestones, oölite, and algal pisolites also occur predominantly near the base of the carbonate parts of a cyclic unit. In the Upper Cambrian the rocks are more varied and display a greater abundance of features indicative of very shallow water and emergence. Stromatolites and stromatolitic limestones are prominent (Pl. VIII-3), silty and sandy carbonates with minor beds of non-glauconitic siltstone and sandstone are common, and many thin beds of oölite and limestone–pebble con-

glomerate occur. Mottled and dense limestones are relatively unimportant. Biocalcarenites are minor and non-glauconitic.

The western margin of the middle carbonate belt has many of the attributes of a reef. Possible reef-builders are blue-green algae (for whose activities there is abundant evidence), *Stromatactis*-like structures, and an encrusting organism suggestive of a stromatoporoid or encrusting calcareous algae that forms reefal masses at Mount Assiniboine. Also found near the margin are large, reef-like masses of megacrystalline white dolomite, which appear to have shed coarse debris that was incorporated into adjoining beds at the same stratigraphic level. No trace of the organism that formed these masses has been found, if indeed they do represent organic accumulations.

Dolomitized equivalents of all limestone lithofacies in the middle carbonate belt are recognized; the bulk is considered secondary but several lithofacies are known only as dolomites. The dolomites thought to be of early diagenetic origin are common in the Waterfowl, Lyell, and Mistaya Formations. They occur as laterally persistent beds, fine to microcrystalline, argillaceous, laminated with silt or sand, and are associated with indicators of very shallow water and emergence such as mudcracks, ripple-marks, breccia, and pebble-conglomerates. A widespread zone of dolomite at the top of the Pika Formation is attributed to dolomitization during deposition of the overlying evaporitic Arctomys Formation. Great masses of secondary dolomite, probably of post-diagenetic origin, are characteristic of the Cathedral, Eldon, and Pika Formations, and are found also in the Upper Cambrian. They are medium to coarsely crystalline and commonly vuggy, and in many instances the lateral contacts are steep and cut across the bedding so that the masses have a greater height than width.

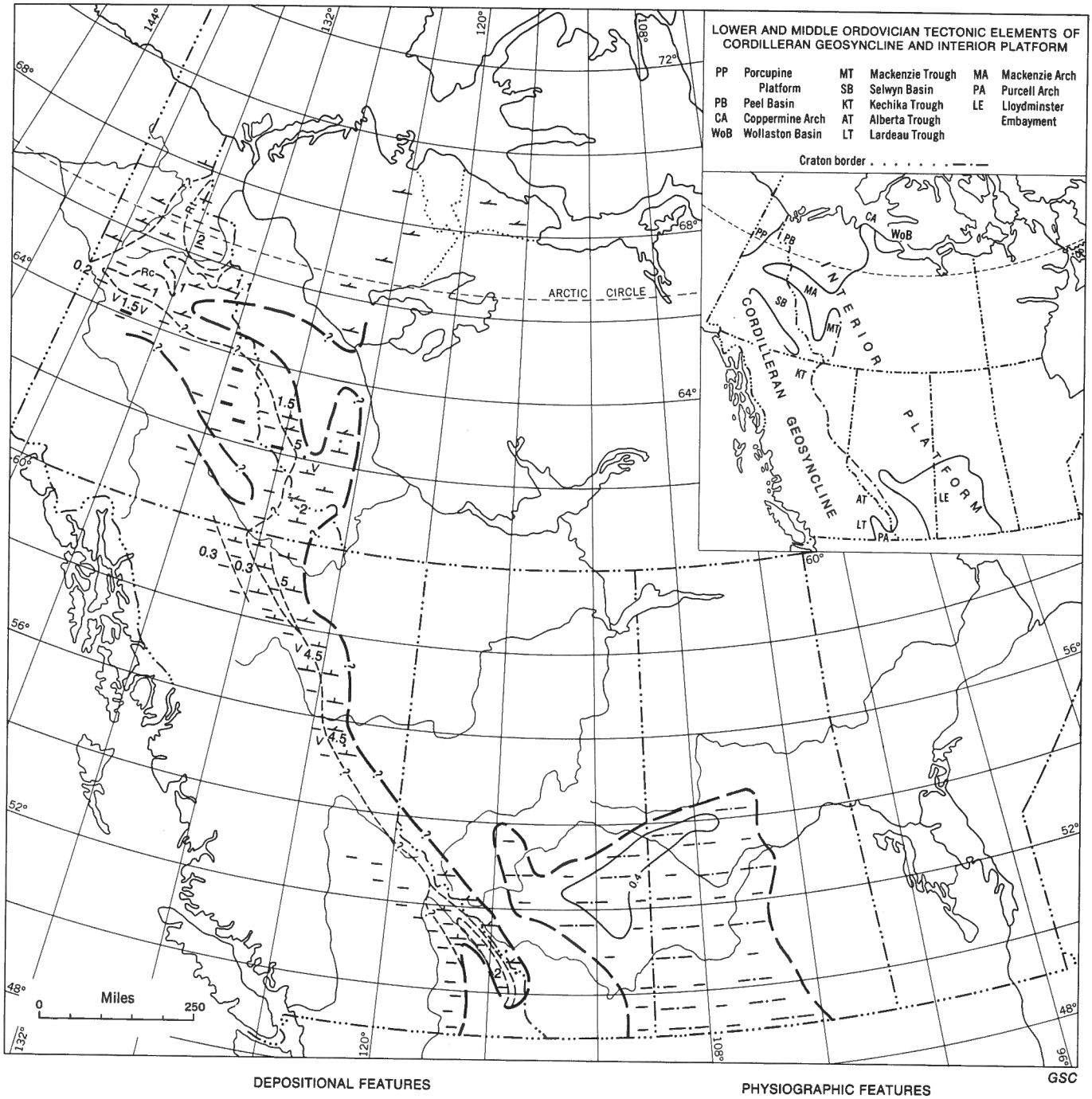
The outer detrital belt is characterized by thin-bedded and laminated, green to dark grey, calcareous and silty shales and by thin-bedded argillaceous limestone. The rocks are poorly fossiliferous and generally strongly deformed or metamorphosed to phyllite. Indicators of shallow water are few, and evidence of slumping and sliding of the sediments on depositional slopes can be seen at several places. Glauconite is absent. The facies is represented by the Chancellor and Canyon Creek Formations and the lower parts of the Goodsir and McKay Groups. Tongues occur in the middle carbonate belt as the Naiset Formation at Mount Assiniboine and as a thin shale in the Eldon Formation at Vermilion Pass and Field. The facies changes to most formations of the middle carbonate belt occur approximately along the same line and form a fairly abrupt front. That of the Cathedral Formation is readily seen near Field and Mount Assiniboine. The early Upper Cambrian carbonates, the Ottertail or Jubilee Formation, extend much farther west across Alberta Trough, and the latter overlaps the older rocks of Purcell Arch.

Ordovician and Silurian

Tectonic Summary

Throughout the eastern Cordilleran Geosyncline, Lower and Middle Ordovician rocks form a conformable sequence, conformably succeeding the Cambrian. The Upper Ordovician unconformably overlies all older systems and is conformably succeeded by the Silurian. In the more northerly and internal parts of the geosyncline deposition locally appears to have been continuous from the late Cambrian into the Devonian. Remnants of Lower and Middle Ordovician carbonates border the craton (Fig. VIII-9). They are truncated along their eastern limit beneath the Upper Ordovician and Devonian, but thicken rapidly westward into the geosyncline and grade abruptly either into relatively thinner graptolitic shales or into an intervening calcareous shale and argillaceous limestone facies. In Peel Basin shales are interbedded with limestone turbidites and are interpreted as a deep-water facies lying between the thick carbonates covering Porcupine Platform to the west and Mackenzie Arch to the south. The shales are oldest in the north and possibly represent the southwestern continuation of a similar facies in the Franklinian Geosyncline of the Arctic Archipelago. In Selwyn Basin bedded cherts and shale accumulated. Basic flows are intercalated in the northwest and are also present in Middle Ordovician carbonates near southern Mackenzie Arch. Black shales occupy the axis of Alberta Trough and occur in Lardeau Trough. Calcareous shale and limestone overlie at least part of Purcell Arch suggesting deposition in shallow water and relatively less depression of the arch. Tongues of orthoquartzite, possibly derived from uplift of Peace River and Alberta Arches, are intercalated in the Middle Ordovician carbonates of eastern Alberta Trough and west of Peace River Arch, indicating some uplift of the arches. In eastern Lloydminster Embayment thin, shallow-water Lower Ordovician shales and siltstones are preserved. During retreat of the sea gentle uplift of bordering arches initiated development of the weakly negative Williston Basin, centring in Montana.

In the Late Ordovician the craton was generally uniformly depressed, the seas transgressing the truncated margins of earlier Paleozoic rocks and the Precambrian of the Shield (Fig. VIII-10). Although now absent, Upper Ordovician carbonates probably covered parts of the central-western part of the craton, Severn Arch, and the northern Shield, as the Upper Ordovician is faunally similar throughout western and northern Canada, and also Hudson Platform. Facies relationships in the northern Cordilleran Geosyncline are similar to those of the Lower and Middle Ordovician rocks with which, in Peel and western Selwyn Basins, Upper Ordovician and Silurian carbonates, shales, and minor pyroclastics form a conformable sequence. Differences are mainly in the greater extent of the graptolitic shale facies, particularly in southern Mackenzie Trough which was deeply depressed. MacDonald Platform was possibly not covered by the Upper Ordovician; Middle Silurian dolomites lie with



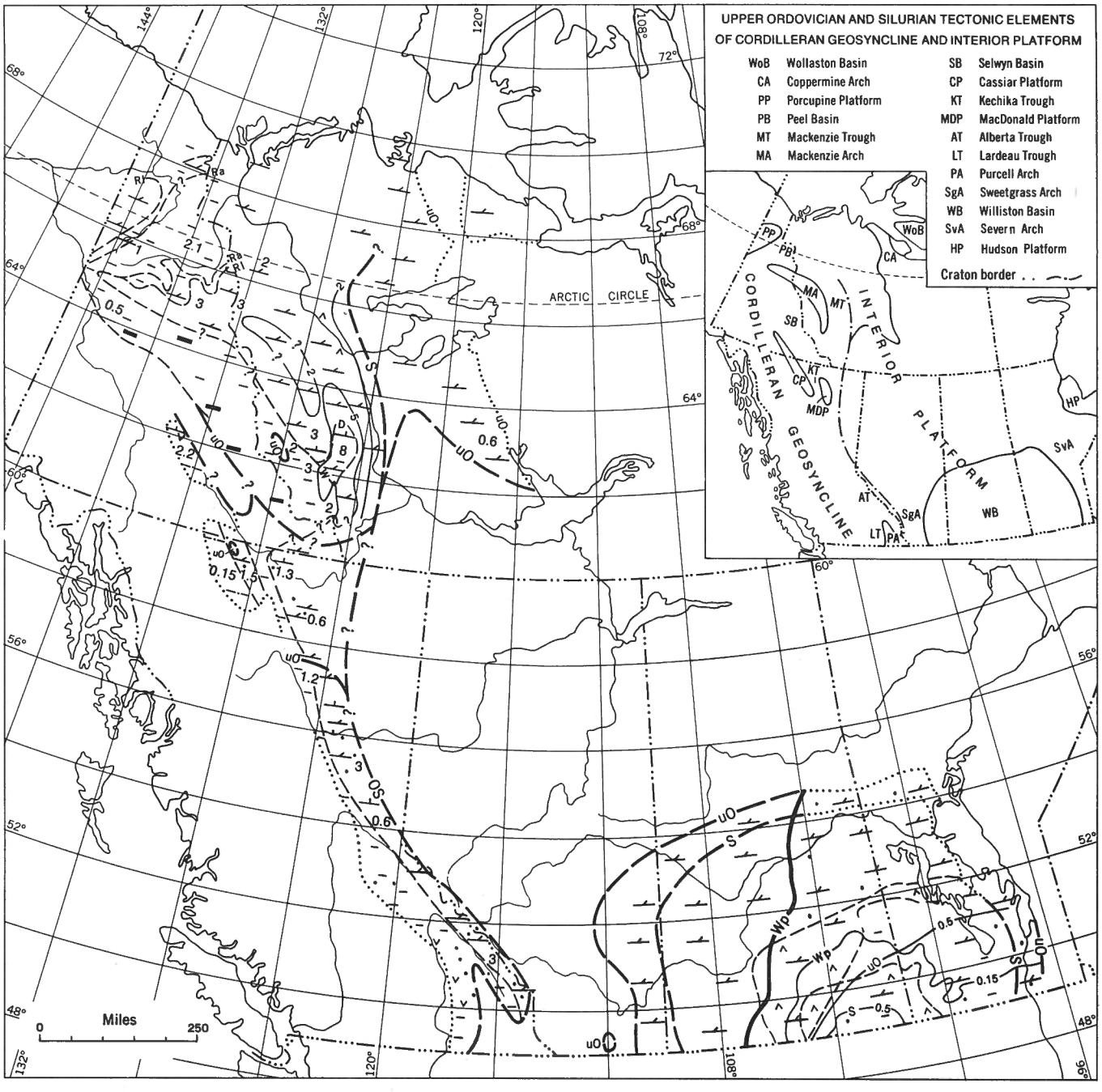
DEPOSITIONAL FEATURES

- Sandstone
- Siltstone
- Shale
- Chert
- Volcanics
- Limestone
- Dolomite
- Truncated limit
- Facies boundary
- Road River Formation
  - Caradocian . . . Rc
  - Tremadocian . . . Rt
- Outcrop limit
- Isopach (thousands of feet)
- Thickness (thousands of feet) . . . 4.5

PHYSIOGRAPHIC FEATURES

Blank areas are lowlands and plains developed on crystalline and sedimentary rocks of the craton, and where Lower and Middle Ordovician is now absent through erosion, or unknown. In the geosyncline the blank areas may be sea.

FIGURE VIII-9. Lower and Middle Ordovician sedimentation and tectonism in Western Canada.



DEPOSITIONAL FEATURES

Sandstone
Siltstone
Shale
Limestone
Dolomite
Anhydrite
Chert
Volcanics

Depositional limit
Winnipeg Fm . . . . . Wp
Truncated limit
Silurian . . . . . S
Upper Ordovician . . . . . uO
Ordovician-Silurian . . . . . OS
Facies boundary
Delorme . . D Whittaker . . W
Road River, Ashgillian . . . . Ra
Road River, Llandoverian . . . Rl

Outcrop limit
Upper Ordovician . . . . . uO
Isopach (thousands of feet)
Silurian . . . . . S
Upper Ordovician . . . . . uO
Winnipeg Fm . . . . . Wp
Thickness (thousands of feet) 0.6

PHYSIOGRAPHIC FEATURES

Blank areas are lowlands and plains developed on crystalline and sedimentary rocks of the craton, and where Upper Ordovician and Silurian is now absent through erosion, or unknown. In the geosyncline the blank areas may be sea.

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FIGURE VIII-10. Upper Ordovician and Silurian sedimentation and tectonism in Western Canada.

angular unconformity on the Proterozoic. Purcell and Alberta Arches, earlier sources of quartz sand, were possibly covered by Upper Ordovician carbonate. In Alberta Trough, the latter was succeeded by Lower Silurian graptolitic shale and siltstone and in Lardeau Trough by black shale, basic flows, and pyroclastics. The seas transgressed from the southeast into the newly formed, weakly autogeosynclinal Williston Basin, on the northern flank of which thin evaporites were deposited.

Northern Cordilleran Geosyncline and Interior Platform

In the *Richardson Mountains* and *Porcupine Plateau* most of the Ordovician and Silurian is represented by the Road River Formation, a thick sequence of thinly bedded graptolitic shale, argillaceous limestone, limestone turbidite, and chert, the lower part being Upper Cambrian and the upper part being Lower Devonian (Norford, 1964). It is overlain, generally conformably, by Lower Devonian shales or unconformably by the Middle Devonian. Graptolites are rather common, ranging from Lower Arenigian to Lower Devonian, one of the most continuous successions known (Jackson and Lenz, 1962). Those representing the Wenlockian are absent, suggesting a possible hiatus, although there is no stratigraphic evidence. The turbidites, thick graded-bedded limestone breccias and calcarenites, are thought to have been derived from banks of carbonate bordering Peel Basin on the south and west. The carbonates of northern Wernecke Mountains and western Porcupine Plateau are thin to massive-bedded, grey weathering, fine-grained dolomites and some pelletoid and bioclastic limestones, and are generally unfossiliferous. The facies change from shale to carbonate is abrupt, the shales in general being less extensive in the Ordovician than in the late Silurian, when carbonates were deposited only in Keele and Iltyd Ranges (Figs. VIII-9 and -10). The carbonates are thickest adjacent to the shale facies and thinner on Mackenzie Arch and probably also Porcupine Platform.

A sequence, uniquely condensed but apparently complete, of Ordovician and Silurian shale and chert only 400 to 900 feet thick occurs in *western Ogilvie Mountains* near the Alaskan border just north of Yukon River (Churkin and Brabb, 1965). It is overlain disconformably by Midde Devonian chert and shale. In *central Ogilvie* and *Wernecke Mountains,* Lower Ordovician to Upper Silurian strata are dominantly carbonate with minor interbedded shale. Basal Lower Ordovician or Upper Cambrian (?) carbonates unconformably overlie rocks as old as Helikian. In southern Wernecke Mountains the Ordovician is represented by as much as 1,900 feet of dolomite, argillaceous limestone, tuffaceous argillite, limestone, shale, and very minor sandstone and chert-pebble conglomerate. Brown tuffaceous argillites and thin beds of limestone ranging from 580 to possibly 1,000 feet thick grade south and southeast into massive volcanic rocks, in part vesicular (Green and Roddick, 1962).

In *southern Ogilvie* and *Selwyn Mountains* graptolitic shales, siltstones, and cherts are widespread throughout Selwyn Basin. The facies change to carbonate rocks of the western flank of Mackenzie Arch is fairly abrupt and coincides roughly with the boundary between Mackenzie and Selwyn Mountains (Fig. VIII-11a). In the few places where thicknesses can be measured the graptolitic sequence seems to be considerably thinner than the correlative carbonate sequence. Great thicknesses of graptolitic shales, cherts, and siltstones outcrop in the MacMillan Plateau region (Roddick and Green, 1961a). Locally, the oldest strata of this assemblage may be Silurian but in most places Ordovician strata are included. The rocks appear to overlie unconformably clastic rocks of Proterozoic and possibly Cambrian age and may grade upward conformably into the Devono-Mississippian shales and chert-pebble conglomerates.

Throughout northern and central Mackenzie Mountains, Franklin Mountains, and the northern Interior Plains the Upper Ordovician overlaps the Lower and Middle Ordovician and rests unconformably on rocks as old as the Helikian. Beneath the *northern Interior Plains* the Ronning Group, according to E. J. Tassonyi, is divisible into a lower part comprising about 300 feet of microcrystalline, partly silty, and anhydritic dolomite, a middle part of 635 to 1,500 feet of pale grey, fine to coarse, porous dolomite with milky chert, and an upper part of as much as 1,400 feet of light grey and buff, finely crystalline dolomite. The Mount Kindle Formation of *Franklin Mountains* is formed from thickly bedded, grey, porous dolomite of Late Ordovician age and succeeding, finely bedded, silty dolomites of possible Silurian age. The Chedabucto Lake Formation between Great Bear and Great Slave Lakes is similar but only 580 feet thick compared with the 900 feet for the Mount Kindle.

In *southwest Mackenzie Mountains* thick sequences of Ordovician and Silurian carbonates and shales are included in Sunblood, Whittaker, and Delorme Formations (Gabrielse, *et al.,* 1965; Douglas and Norris, 1961). In Backbone Ranges the Sunblood Formation is entirely Middle Ordovician in age and comprises about 4,000 feet of distinctively banded, grey, cream, dark grey, and wine-coloured dolomite and limestone with minor orange-brown-weathering sandstone. Locally, the formation includes one or two andesite or basalt flows as thick as 150 feet. To the north and east the Sunblood appears to be arenaceous and is commonly absent beneath the unconformity at the base of the Upper Ordovician Whittaker Formation. To the southwest, the Sunblood is underlain by more than 1,000 feet of Upper Cambrian and Lower Ordovician limestones and dolomites. The Whittaker Formation consists of banded light and dark grey weathering, massive, dark grey coarsely crystalline dolomite, thinly bedded argillaceous limestone, and grey graptolitic shale. The dolomites comprise the middle part of the formation and carry the coralline Upper Ordovician fauna; the upper part has yielded graptolites as young as Ludlovian. The

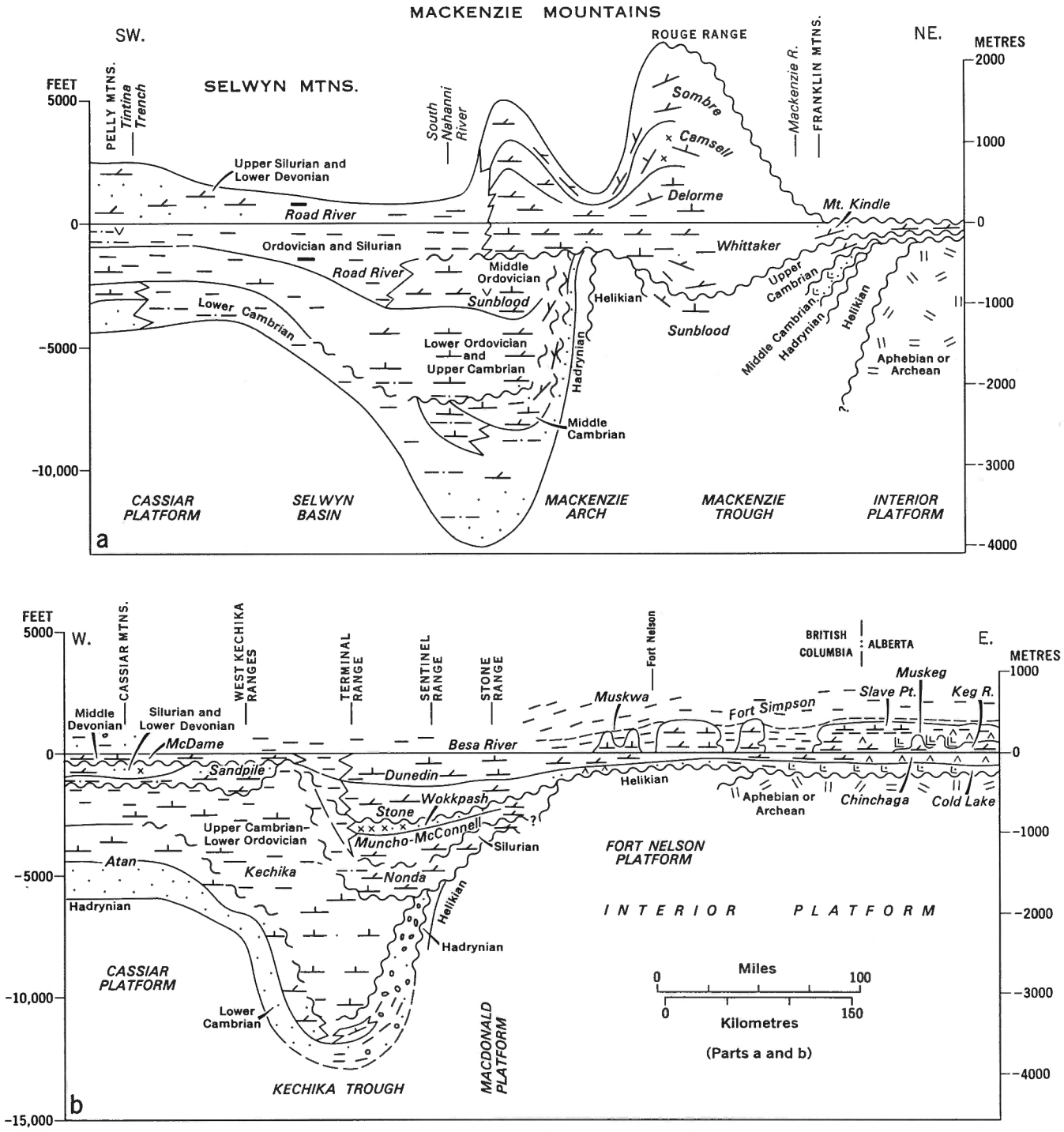


FIGURE VIII-11. Diagrammatic restored sections of lower Paleozoic rocks between Cassiar Platform, Cordilleran Geosyncline, and northwestern Canadian Shield: (a) Pelly Mountains, Yukon Territory, to southern Franklin Mountains, District of Mackenzie (by R. J. W. Douglas and H. Gabrielse); (b) Cassiar Mountains to Fort Nelson, British Columbia (by H. Gabrielse and H. R. Belyea).

conformably overlying Delorme Formation may include Lower Devonian in its uppermost part. It is typically banded buff and brown weathering, thinly bedded shale, limestone, and dolomite that grade southeastward into silty dolomite. The succession as a whole attains a thickness of about 10,000 feet in the axial part of southern Mackenzie Trough (Pl. VIII-4) thinning westward to less than 1,000 feet on southern Mackenzie Arch, partly through convergence in the Whittaker and Delorme Formations, and partly by removal of the Whittaker strata. To the east, the succession, together with the thick conformably overlying beds referred to the Lower Devonian, is rapidly bevelled beneath the Middle Devonian. To the southwest the Ordovician and Silurian change facies to graptolitic shales, siltstones, cherts, and argillaceous limestones in Selwyn Mountains.

Southwest of Tintina Trench graptolitic shales of



PLATE VIII-4. Whittaker and Iverson Ranges, southern Mackenzie Mountains, Northwest Territories. Looking north at easternmost structure of Mackenzie Fold Belt and conformable succession from Middle Ordovician to Upper Devonian more than 15,000 feet thick. Width of Whittaker anticline is 10 miles.

Ordovician age, ranging from a few hundred feet to a maximum of about 750 feet in thickness, outcrop in *Pelly* and *Cassiar Mountains*. They appear to be conformable with underlying Cambrian rocks and are overlain unconformably by latest Ordovician or Silurian rocks. In Cassiar Mountains the upper part of the Kechika Group of Cambro-Ordovician age includes dominantly argillaceous strata in the west that change to a much thicker, dominantly calcareous facies in the east (Gabrielse, 1963). Highly altered greenstones are common and may be in part extrusive. Upper Ordovician (Richmond) cherty dolomites are present locally beneath the Silurian Sandpile Group. Near Rocky Mountain Trench Lower Silurian graptolites occur in the siltstones of a thick, recessive assemblage that includes strata of the Cambro-Ordovician Kechika Group and is unconformably overlain by Mississippian carbonates. Cherty and sandy dolomites of the Sandpile Group, about 1,500 feet thick, disconformably overlie Lower and Middle Ordovician graptolitic shales in Cassiar Mountains. The westernmost exposures are 100 to 200 feet of graptolitic siltstone disconformably overlying Lower Ordovician shales and overlain by laminated dolomites and dolomitic sandstones possibly in part of Late Silurian age. In Pelly Mountains, andesitic tuff and breccia as much as 300 feet thick are interbedded with Silurian graptolitic siltstones overlain by sandy and silty dolomites of Middle and Late Silurian age (Green, *et al.*, 1960a, 1960b).

Southern Cordilleran Geosyncline

In *northern Rocky Mountains* the upper part of the Kechika Group includes two facies: the upper Cloudmaker Formation as much as 1,255 feet of calcareous shale, siltstone, and cherty carbonate mainly of Middle and Late Ordovician age, and the lower Mount April Formation, more than 4,500 feet of argillaceous limestone mainly of Early Ordovician age (Jackson, *et al.*, 1965). The contact is diachronous, the lower or carbonate facies rising stratigraphically eastward and the whole bevelled beneath the late Lower Silurian Nonda Formation (Norford, *et al.*, 1966). The Nonda is mainly a thin dark grey dolomite with basal and upper quartz sandstones. It lies with angular unconformity on the Helikian on the crest of MacDonald Platform (Pl. VIII-2), and unconformably on the Ordovician as far south as Peace River Arch. It grades westward into thicker limestone and thence into black graptolitic shales in Kechika Trough.

In *central* and *southern Rocky Mountains* the Lower and Middle Ordovician are represented by the Chushina, Survey Peak, and Outram Formations of shale, argillaceous limestone, and silty dolomite (Aitken and Norford, 1967). The succeeding Monkman and Tipperary Formations are easterly derived, fine quartz sandstone or equivalent brown shales and nodular limestones. Overlying beds are the finely crystalline silty dolomites of the Skoki Formation and the locally present Owen Creek dolomite. These rocks grade westward into the upper part of the Cambro-Ordovician McKay Group and the black graptolitic shales of the Glenogle in the axial region of Alberta Trough. The Upper Ordovician lies unconformably on the older rocks, which are basal pure quartz sandstones, easterly derived, and termed the Mount Wilson or Wonah quartzites, the latter possibly also derived from Purcell Arch to the west. Succeeding strata are massive grey Beaverfoot dolomites as thick as 3,000 feet with a shelly fauna, and the thin, early Silurian Tegart dolomites and graptolitic shales. In eastern and northern *Purcell Mountains* and in the Rocky Mountain Trench, the McKay Group, possibly entirely Upper Cambrian, is disconformably overlain by thin unfossiliferous dolomite typical of the Beaverfoot Formation.

The predominantly argillaceous Active Formation of southern *Selkirk Mountains* contains Lower Ordovician graptolites (Little, 1960). The Active lies with structural conformity on the calcareous shales of the Nelway Formation and probably extends through the Ordovician and into the Silurian or even Middle Devonian, as the lithologically similar Ledbetter slate, just south in Washington, contains Lower and Middle Ordovician graptolites and an overlying formation includes Silurian graptolites and Middle Devonian fossils. Elsewhere in Lardeau Trough, formations lithologically similar to the Active, but unfossiliferous, are in the middle part of the Lardeau Group, particularly the sooty black argillaceous Triune and Sharon Creek Formations that are separated by the Ajax Quartzite (Fyles and Eastwood, 1962). The Ajax and Triune may be equivalent to the lithologically similar Ordovician Wonah and Glenogle Formations of the Rocky Mountains.

Northern Flank of Williston Basin

The Lower Ordovician of Williston Basin (Fig. VIII-12) is represented by the upper part of the Deadwood Formation, described previously, as it is partly of Late Cambrian age. The Middle Ordovician is missing. The Upper Ordovician and Silurian (Porter and Fuller, 1964) form a conformable succession. The basal unit is the unfossiliferous Winnipeg Formation, which may be of Middle Ordovician age. It has a maximum thickness of 350 feet in North Dakota but thins northward and lies with regional unconformity on Lower Ordovician and Cambrian strata and the Precambrian crystalline rocks of the Canadian Shield. It consists of basal light grey to white, fine- to medium-grained, well-sorted, porous quartzose sandstone with calcareous or ferruginous cement, overlain by green to greenish grey, non-calcareous, locally waxy, pyritic shales, siltstones, and sandstones. Hematitic or limonitic oölites, phosphatic and siliceous pellets, and pyrite are found at some horizons. The northwest termination is abrupt possibly due to regression and erosion before deposition of the succeeding Red River Formation. Its present limit, however, may mark a shoreline against an underlying linear topographical feature.

The overlying strata show a general upward progression from fossiliferous limestones through rhythmic alternations of carbonates and evaporites into cryptocrys-

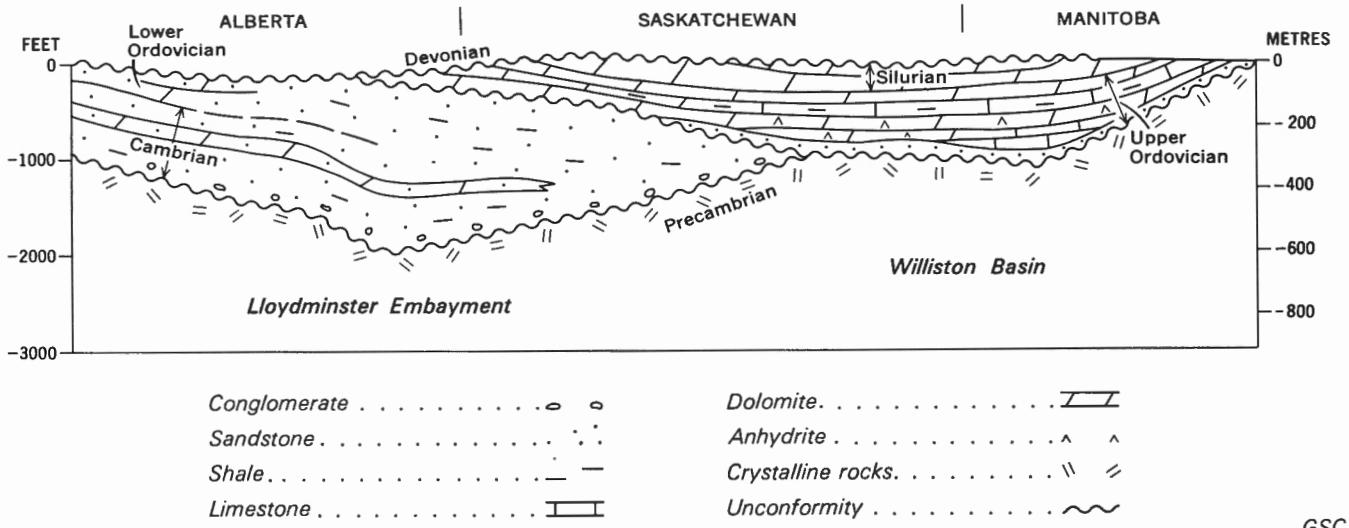


FIGURE VIII-12. Stratigraphic section of Cambrian, Ordovician, and Silurian rocks of Lloydminster Embayment and Williston Basin, Alberta, Saskatchewan, and Manitoba (by H. R. Belyea).

talline dolomites with sandy layers. A gradual shallowing of the basin and upwarp of adjacent land areas is indicated. Also, each formational unit shows a progressive change shoreward from fossiliferous limestones, interpreted as marine shoals, through a rhythmical carbonate–evaporite facies into cryptocrystalline dolomites interpreted as littoral to sub-littoral deposits. These facies relationships are less pronounced in the predominantly shallow, evaporitic Silurian rocks.

The basal unit of the Red River Formation, the Yeoman Member, is 350 feet thick in southern Saskatchewan and thins to 100 feet before truncation by post-Silurian erosion. It consists predominantly of mottled tan and brown dolomitic limestone carrying bryozoa, brachiopod, echinoderm, and coral fragments and grades north and west to cryptocrystalline and finely crystalline varicoloured dolomite, locally cherty. The overlying Herald Member consists of a rhythmical sequence of dolomitic fossiliferous limestone, dolomite, and anhydrite grading southeast offshore to a predominantly dolomitic, fossiliferous limestone shoal and shoreward to cryptocrystalline and finely crystalline evaporitic dolomite.

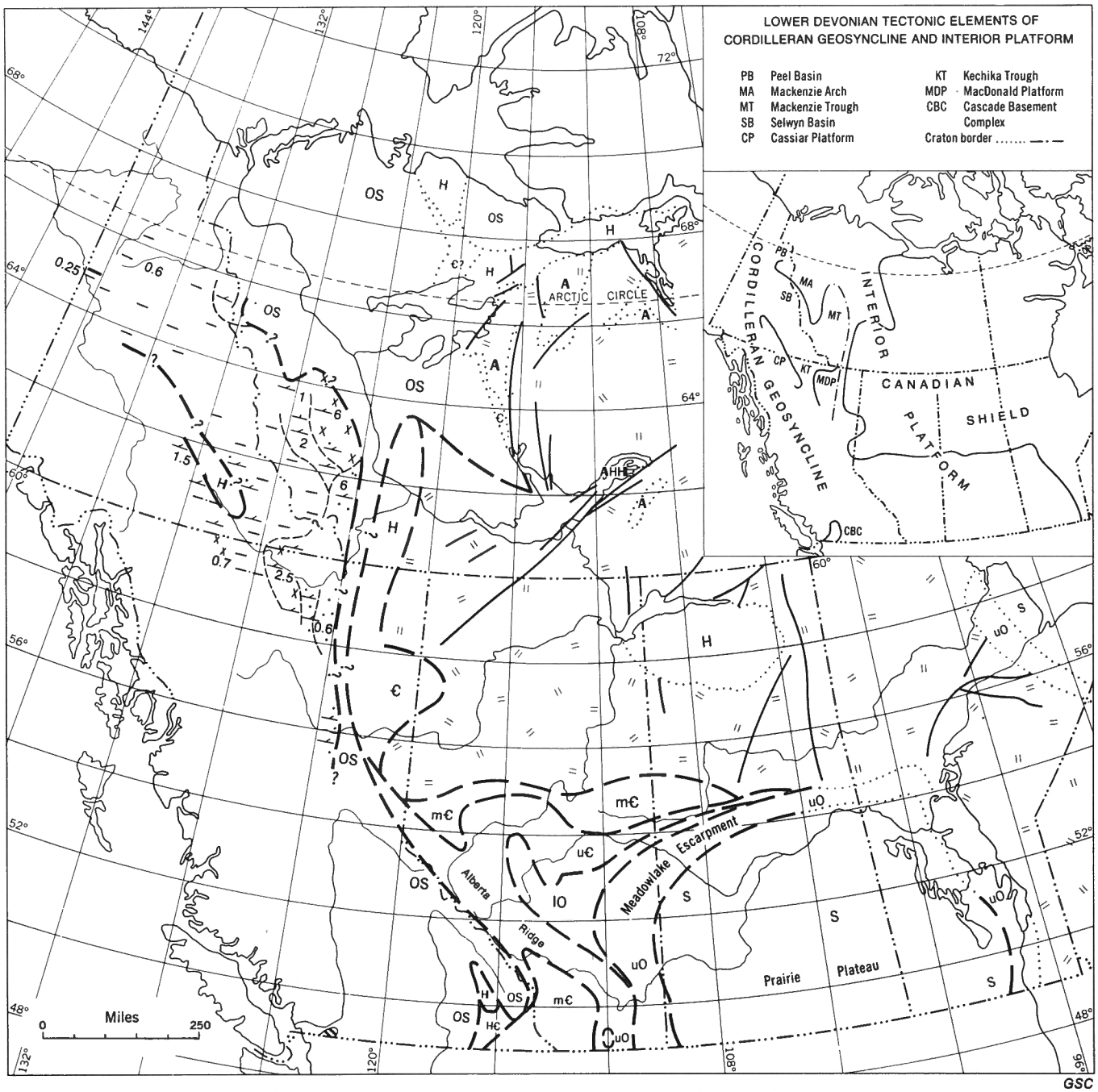
The Red River is sharply overlain, probably disconformably, by fossiliferous shaly limestone and shales of the Stoughton Member of the Stony Mountain Formation which is thin and grades to limestone indistinguishable from the Gunton. The Gunton, 55 feet thick, is mottled fossiliferous dolomitic limestone, dolomite, anhydrite rhythmically alternating like the Herald, overlain by a basin-wide grey to red-brown silty dolomite that completes the cycle. The succeeding Stonewall, 100 feet thick across the basin, comprises carbonate–evaporite rhythms interrupted by thin basin-wide beds of argillaceous dolomite and shale that contain lenses of coarse, rounded, frosted quartz sand, and fragments of the immediately underlying dolomite and shale. The top of the Ordovician in the subsurface is taken by Porter and Fuller (1964) as the uppermost grey and red arenaceous shale and is placed by Brindle on paleontological evidence at the "t marker", an argillaceous zone with floating quartz sand grains.

The Silurian Interlake Group (Stearn, 1956) consists of a sequence of dolomitized fossiliferous limestone, stromatolitic and oölitic dolomite, porous vuggy dolomite, and fossiliferous biostromal dolomite at the top. These units are not readily recognizable in the subsurface. Across Saskatchewan (Porter and Fuller, 1964) the lower Interlake consists of cryptocrystalline and finely crystalline dolomite with shaly, silty anhydritic layers, some with salt casts. The middle Interlake, preserved only in the south, consists of pale, cryptocrystalline, and very fine grained dolomite interbedded with fossiliferous dolomitic limestone and sandy layers, crossbedded and mudcracked. Dolomitized pelletoid limestone with nodules, possibly algal, and clotted dolomitic mudstones complete the Saskatchewan section although still younger light coloured, cryptocrystalline, anhydritic, and halitic pelletoid dolomite and vuggy fossiliferous dolomite are present in North Dakota.

Devonian

Tectonic Summary

Lower Devonian rocks, poorly fossiliferous, are restricted to the Cordilleran Geosyncline (Fig. VIII-13). Thin dark shales conformably overlie Silurian shales in Peel Basin and eastern Selwyn Basin. The latter grade eastward into thick carbonates and breccia bordering Mackenzie Arch, thin over the southern part of the arch through convergence and pre-Middle Devonian erosion, and thicken eastward into Mackenzie Trough (Fig. VIII-11a). There, the Lower Devonian is represented by massive dolomite and limestone breccia that grade southeast to sandy dolomite. Thin sand zones on MacDonald Platform grade westerly into carbonate breccia and thence



DEPOSITIONAL FEATURES

- Chert
- Sandstone
- Shale
- Dolomite
- Limestone
- Carbonate breccia
- Truncated limit
- Facies boundary
- Thickness (thousands of feet) 2.5
- Outcrop limit

PALEOGEOLOGY

- Silurian S
- Ordovician O
- Cambrian €
- Hadrynian Hl
- Helikian H
- Aphebian A
- Archean and Aphebian crystalline rocks
- Cascade basement complex
- Fault

- Lower l
- Middle m
- Upper u

PHYSIOGRAPHIC FEATURES

Except where indicated otherwise, the areas of platform underlain by sedimentary rocks probably form low-lying plains. The areas of Precambrian crystalline rocks are lowlands of moderate to little relief. In the geosyncline, blank areas are probably sea.

FIGURE VIII-13. Lower Devonian sedimentation and tectonism in Western Canada, and some features of the pre-Devonian geology and physiography.

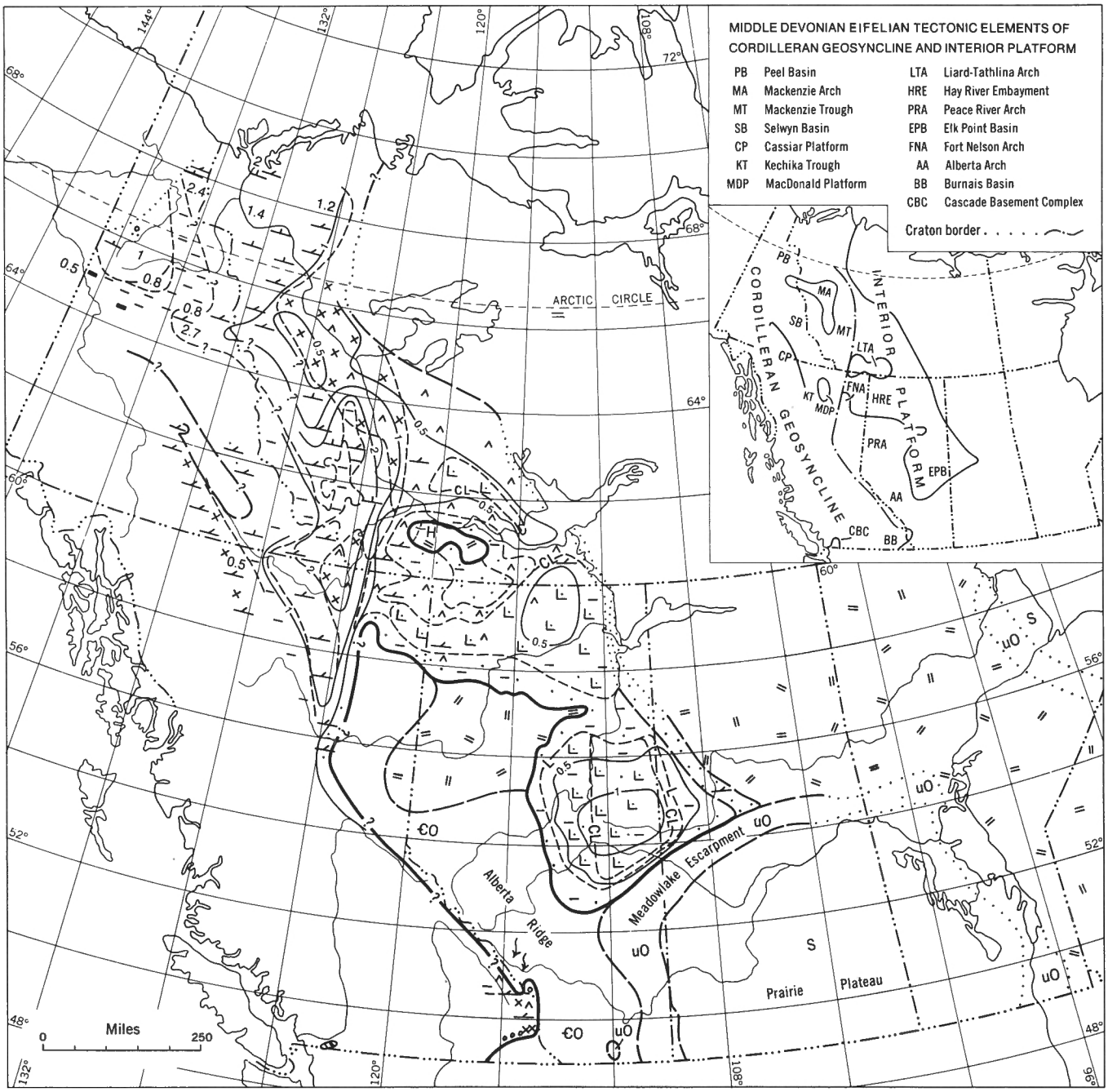
to carbonate and shale of Kechika Trough. Widespread but relatively thin sandstones, carbonates, and carbonate breccia occur on Cassiar Platform, the quartz presumably derived from an emergent western part. The craton was generally high during the Early Devonian and under active erosion. The sedimentary cover was stripped back from Precambrian crystalline lowlands and a topography of considerable relief established. Prominent features were broad valleys such as the one surrounding the Tathlina Upland and gentle cuestas carved in Paleozoic sediments such as Meadowlake Escarpment and Alberta Ridge. Possibly at this time the Ice River alkaline complex was emplaced into Cambro-Ordovician rocks on southern Alberta Arch.

Various products of erosion of Middle Devonian or late Early Devonian age, part non-marine, fill stream channels and low areas on the erosion surface (Fig. VIII-14). The streams on the west flank of Alberta Ridge drained into Burnais Basin where anhydrite, gypsum, and sandy dolomite were being deposited on the site of the axis of the former Alberta Trough. Burnais Basin was flanked on the west by the Purcell Arch that was partly overlapped by shale and carbonate and bounded on the south by an escarpment along Moyie fault from which conglomerate and breccias were derived. In Elk Point Basin during the early Middle Devonian, and possibly also in the late Early Devonian, clastics, salt, and anhydrite were deposited in land-locked lakes or tongues of the sea. The sea may have initially penetrated the continental interior via valleys in the northwest or crossed the Shield from the northeast and withdrawn at the close of the early Middle Devonian. The evaporites and clastics grade northwestward into thick carbonate, breccia, and anhydrite that flank islands and peninsulas along the margin of the craton and cover weakly positive elements in the northern Cordilleran Geosyncline—Mackenzie Arch and the MacDonald, Cassiar, and Porcupine Platforms. The carbonates grade into shale in adjacent weakly negative troughs and basins. In the present Cascade Mountains Middle Devonian carbonate and clastics were deposited unconformably on a metamorphic complex intruded by diorite.

In the late Middle Devonian (Fig. VIII-15), extensive limestone sheets blanketed the northern craton and adjacent geosyncline, grading northward and westward into thin shale. A continuous bank was formed north of Peace River landmass over and between the slightly depressed Fort Nelson, Liard, and Tathlina Arches, and probably extended along the west flank of Alberta Arch. Limestone at the front grades northwestward to shale and southeastward to dolomite and evaporites. The front is irregular and flanked by stromatoporoid reefs. The limestones near faults in the basement are recrystallized to coarse vuggy dolomite, locally containing galena and sphalerite. Within the barrier reef complex, and to the southeast within the evaporites, are areas of prolific small reefs and atoll-like bodies separated by salt, anhydrite, and dense dolomite. The anhydritic facies passes southward into halite and sylvite bordered by reefy carbonate on the northeast which suggests that formerly the basin extended beyond the present truncated limit and that influx of sea water in that direction was less restricted. Regression occurred at the close of the Givetian, producing an extensive unconformity.

The Ellesmerian Orogeny of the Arctic Archipelago is represented in northern Yukon Territory by granites dated at 355 and 370 m.y. (Fig. VIII-16). They intrude low-grade, southeast-trending metamorphic rocks of Proterozoic and possibly early Paleozoic age that are overlain with angular unconformity by the Lower Mississippian. The products of erosion of this orogenic belt, a thick clastic sequence, are coarsest and non-marine in the north, lie with minor unconformity on the Givetian, and interfinger southward with dark marine shales and siltstones. The bulk of the sediments accumulated in Peel Basin where the miogeosynclinal elements of the Franklinian and Cordilleran Geosynclines merge, and also in Mackenzie Trough which was strongly depressed. Possibly Mackenzie Arch was high, preventing most of the clastics from reaching Selwyn Basin. Thick black shale and ribbon chert accumulated in Selwyn Basin, intercalated with chert-pebble conglomerate derived from an adjacent uplift to the west. Andesite, basalt, shale, and limestone occur farther west. The sedimentary record of the southern Cordilleran Geosyncline may be contained in the extensive metamorphic complexes. Poorly sorted, feldspathic, quartzose clastics of uncertain age and locally having intercalated basic flows and pyroclastics may have been deposited in Lardeau Trough during the Devonian and, if of that age, must have been derived from uplifts farther west.

Early Late Devonian seas transgressed on the craton from the north, enveloping the landmass formed by the gradually subsiding Peace River Arch, and overlapped westward, covering Alberta Arch. Depression of the craton in the Frasnian was uniform over broad areas and greatest adjacent to the present truncated margin, suggesting a broad embayment. Early Frasnian shales of the north grade to carbonate banks and reefs. Shoreline clastics and evaporites flank the landmasses, the latter extending across the southern part of the embayment as a broad evaporite basin bounded on the south by Continental Arch and Central Montana Uplift in the United States. During the mid-Frasnian, carbonate deposition and reef growth kept pace with subsidence and coincided with earlier shoal areas or lines of lesser basement depression. The banks and reefs accumulated on shallow water shelves fringing Peace River Arch, over Alberta Arch, and across southern Alberta and Saskatchewan on shelves fronting land areas in Montana and North Dakota. Chains of reefs and banks extend into the basin where thin black shales and thick greenish grey shale accumulated. Behind the front the carbonates grade into anhydrite. In the late Frasnian, subsidence of the craton decreased and, with cyclical regression, the site of evaporite deposition shifted northwestward and thin sheets of carbonate, mainly clastic,



DEPOSITIONAL FEATURES

- Conglomerate
- Sandstone
- Shale
- Chert
- Limestone
- Dolomite
- Carbonate breccia
- Anhydrite
- Salt (halite)
- Depositional limit
- Truncated limit
- Facies boundary
- Cold Lake salt — CL
- Lotsberg salt — L
- Outcrop limit
- Isopach (thousands of feet) — 0.5
- Thickness (thousands of feet) — 1.2

PALEOGEOLOGY

- Silurian — S
- Upper Ordovician — uO
- Cambrian and Ordovician — €O
- Helikian — H
- Précambrian crystalline rocks
- Cascade basement complex

PHYSIOGRAPHIC FEATURES

- Stream channel

Blank areas are lowlands and plains developed on crystalline and sedimentary rocks of the craton, and where the Eifelian is now absent through erosion, or unknown. In the geosyncline the blank areas may be sea.

GSC

FIGURE VIII-14. Middle Devonian, Eifelian, sedimentation and tectonism in Western Canada, and some features of the pre-Eifelian geology and physiography.

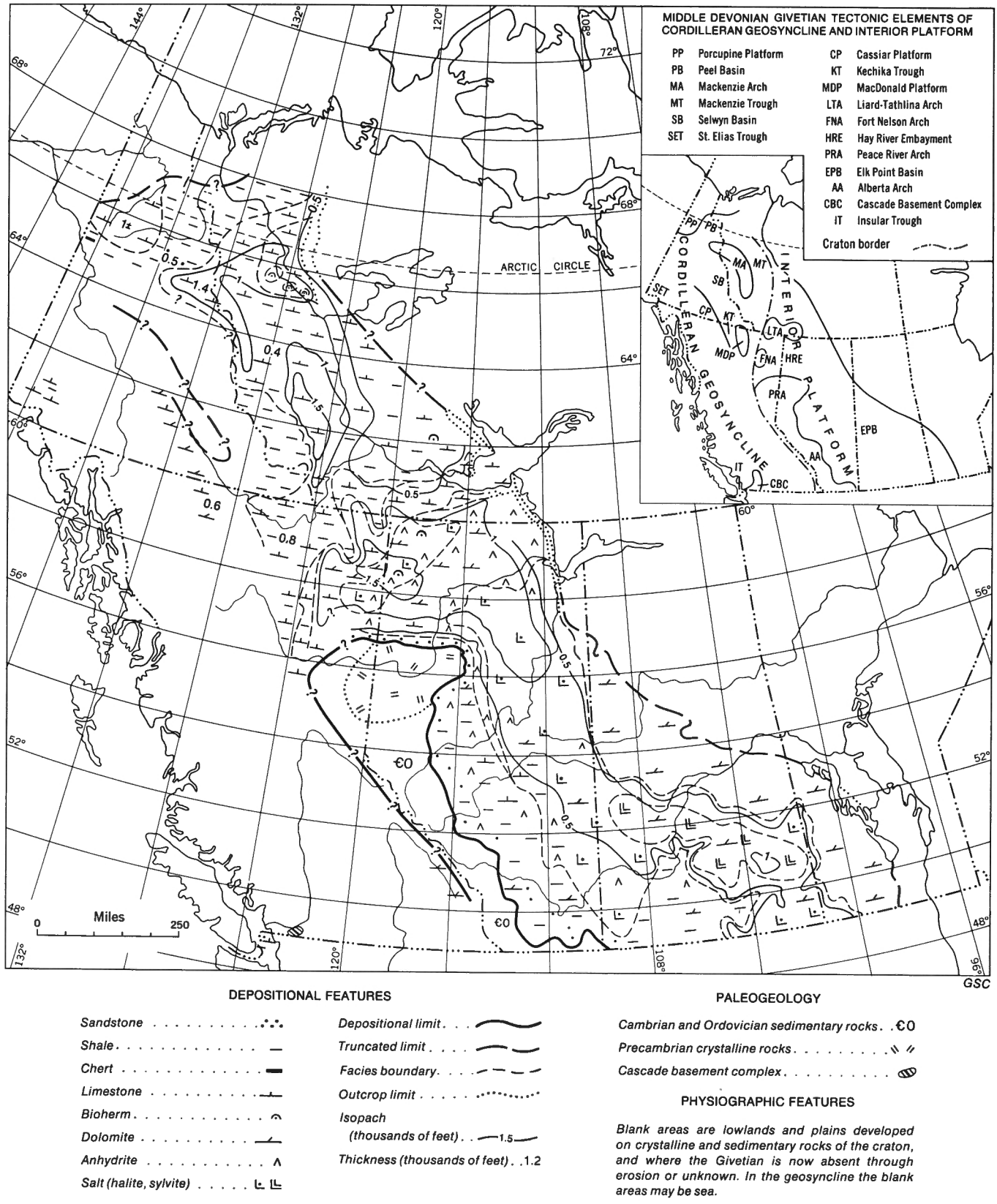


FIGURE VIII-15. Middle Devonian, Givetian, sedimentation and tectonism in Western Canada, and some features of the pre-Givetian geology.



DEPOSITIONAL FEATURES

Conglomerate
Sandstone
Siltstone
Shale
Chert
Volcanics v
Limestone
Dolomite
Bioherm
Carbonate breccia x
Anhydrite ^
Salt (halite)

Depositional limit
Truncated limit
Facies boundary
Beaverhill Lake carbonate-shale BL
Winterburn carbonate-shale W
Winterburn sandstone-shale R
Outcrop limit
Isopach (thousands of feet) 2.5
Thickness (thousands of feet) 1.5

ELLESMERIAN OROGEN

Structural front (approximate) ?
Trend of metamorphic rocks
Granite pluton (age in millions of years) 370

PALEOGEOLOGY

Precambrian crystalline rocks

PHYSIOGRAPHIC FEATURES

Blank areas are lowlands and plains developed on crystalline and sedimentary rocks of the craton, and where the Frasnian is now absent through erosion, or unknown. In the geosyncline the blank areas may be sea.

FIGURE VIII-16. Upper Devonian, Frasnian, sedimentation and tectonism in Western Canada, and some features of the Ellesmerian Orogeny.

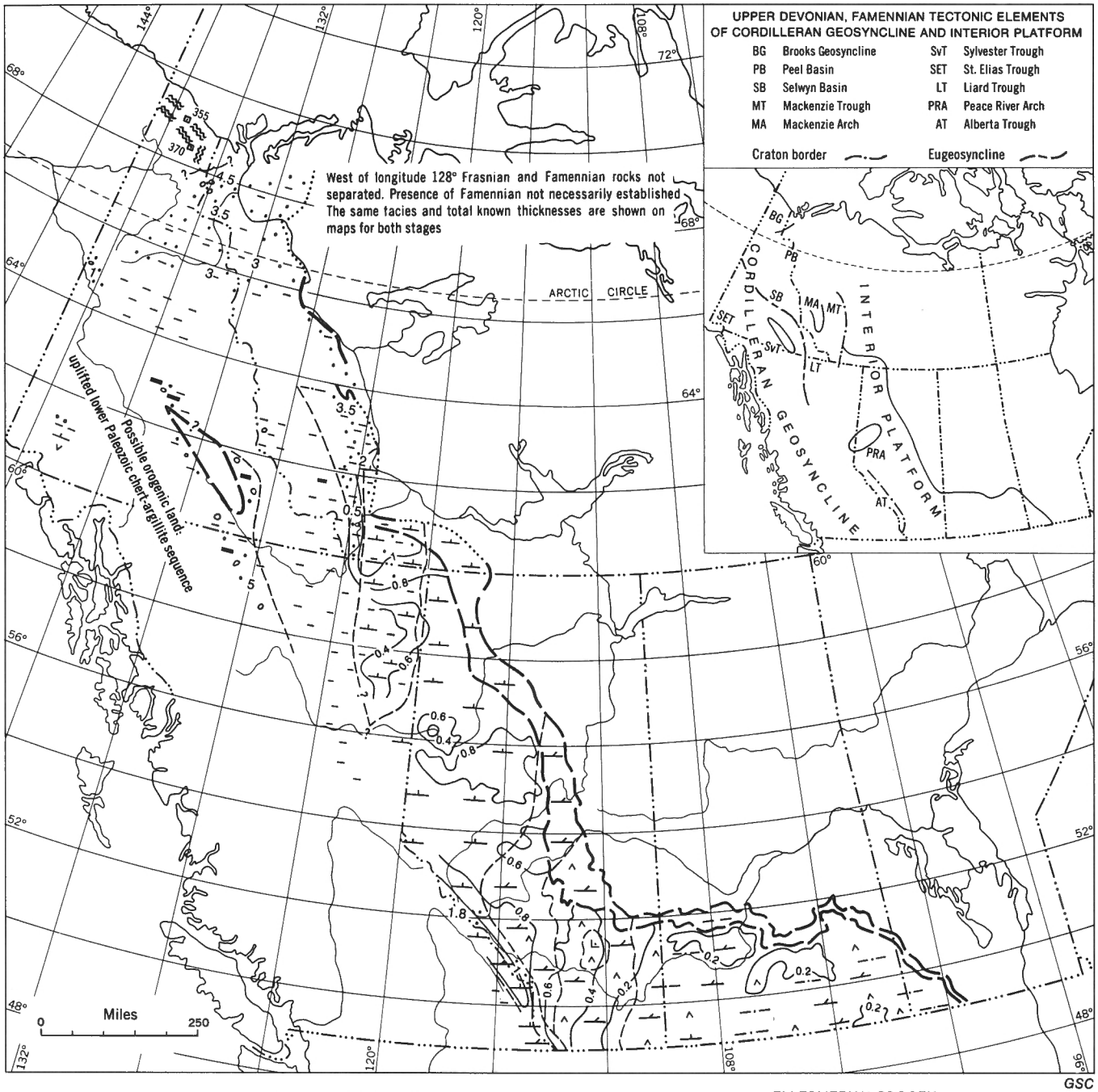


FIGURE VIII-17. Upper Devonian, Famennian, sedimentation and tectonism in Western Canada, and some features of the Ellesmerian Orogeny.

blanketed extensive areas, intertonguing westward with shale. An hiatus separates the Frasnian and Famennian stages, marked by a profound faunal change, slight erosion, and channelling.

Famennian seas transgressed on the craton from the west and north, dark shales grading eastward into dense bahamite-type limestones partly mottled with dolomite in Alberta Trough and covering the weakly negative central part of the craton (Fig. VIII-17). Alberta Trough was re-established as a strongly negative linear element. Peace River Arch persisted, an island existing until it was submerged near the close of the Famennian. MacDonald Platform was possibly static, as in Frasnian times, and was covered by extremely thin black shales. The limestone grades eastward to dolomite and thence to thin rhythmic repetitions of dolomite, anhydrite, shale, and siltstone. The latter, partly red, increases in abundance to the south and east. The latest Famennian is known only in southern Mackenzie Trough; non-deposition or regression occurs on the craton where the Famennian is succeeded by a thin transgressive black shale that may straddle the Devono-Mississippian boundary.

Lower Devonian of Cordilleran Geosyncline

In *Richardson, Ogilvie,* and *Wernecke Mountains* the Lower Devonian is represented by some black shales that form the upper part of the mainly Silurian and Ordovician Road River Formation and by the partly Middle Devonian Michelle and Gossage Formations (Norris, A.W., 1967). The Michelle, 200 to 600 feet thick, is black calcareous shale and aphanitic silty limestone with rich fauna. The trilobites from the lower third and the goniatites represent the Emsian, and the corals are possibly Eifelian. The Gossage Formation, the lower part of which is thin-bedded limestone and dolomite, is of Early Devonian age and has yielded fish remains, considered Gedinnian, from near the base.

The Lower Devonian in southern *Mackenzie Mountains* is represented by the upper Delorme, Camsell, and Sombre Formations (Douglas and Norris, 1961). These units have an aggregate thickness ranging from less than 1,000 feet in northern and central Mackenzie Mountains to more than 6,000 feet in the southeast, and are generally unconformably overlain by the Middle Devonian (Fig. VIII-11a). On southern Mackenzie Arch the Delorme and Camsell Formations comprise a single distinctive buff-weathering, well-bedded, cryptograined dolomite unit in places only a few hundred feet thick. In southern Mackenzie Trough, the uppermost beds of the Delorme Formation, recessive, creamy or reddish weathering, buff argillaceous limestone and shale, are possibly of Early Devonian age. The conformably overlying Camsell Formation consists of grey limestone breccia that grades west and south into bedded cryptograined limestone and dolomite, about 1,500 feet thick, and southeast on Nahanni Plateau, into thick-bedded laminated and crossbedded sandy dolomite. The overlying Sombre Formation comprises well-bedded, resistant, banded, light and dark grey dolomites, 4,000 feet thick in the axial part of the trough but thinning to about 1,500 feet on the flanks. It may be locally absent through pre-Middle Devonian erosion. In southwestern Mackenzie Mountains the Lower Devonian strata change facies to siltstones and shales, generally less calcareous than those of the underlying Ordovician and Silurian strata.

According to G. C. Taylor the McConnell Formation, in *northern Rocky Mountains,* comprises laminated and banded dolomites and minor buff-weathering dolomitic siltstone, ranges in thickness from 300 feet in the east to 2,000 feet in the west, and disconformably overlies the Middle Silurian Nonda Formation (Fig. VIII-11b). In Sentinel Range the overlying Lower Devonian Wokkpash Formation comprises 200 feet of sandstone formed of frosted sand grains. It grades westerly into 500 feet of limestone and dolomite breccia. The westernmost facies, in Terminal Range, are well-bedded, silty dolomites containing fish fragments (Gabrielse, 1962c).

In *Cassiar Mountains,* Upper Silurian and Lower Devonian rocks consist essentially of very pure sandstone or dolomitic sandstone with an average thickness of 500 feet overlain by a similar thickness of laminated dolomite. Locally, dolomitic breccias are abundant. In *Pelly Mountains,* these rocks are included in the Askin Group, the sandstones of which are characterized by extreme purity and exceptionally well developed crossbedding. To the northeast, across Tintina Trench, are sandstones, dolomitic sandstones, and dolomites containing Eifelian fossils near the top.

Ellesmerian Orogeny

In northern Yukon orogenic movements occurred that are herein related to the late Devonian and early Mississippian Ellesmerian Orogeny of Arctic Islands (Chap. X). Three main effects of the orogeny are evident on the mainland: 1) low-grade metamorphism of the Neruokpuk Formation which is probably mainly Proterozoic and lower Paleozoic. Southeast-trending folds were also probably produced, but these have not yet been distinguished from those of later orogenies. The early Mississippian Kyak Formation unconformably overlies the Neruokpuk; 2) the intrusion into the Neruokpuk of small stocks of coarse porphyritic granite at Mount Fitton in Barn Mountains and Mount Sedgewick in British Mountains. Biotite from the former was dated at 370 m.y. and hornblende from the latter at 355 m.y. by K–Ar methods. These dates represent the Givetian and Frasnian, respectively. Biotite from the light grey porphyritic granites of Old Crow Range has yielded ages of 265 m.y. and 220 m.y., possibly indicating younger orogenies; 3) indications of the presence of uplift in the British Mountains or land beneath the present Arctic Ocean as reflected in the northerly source of Upper Devonian clastic rocks constituting the Imperial Formation of the mainland and its temporal facies equivalents of Arctic Islands, the Melville Island Group.

Pre-Devonian Crystalline Rocks of Cascade Mountains

A complex of metamorphic rocks known as the Vedder "greenstone" comprises schists and gneisses of almandine amphibolite facies intruded by saussuritized diorite. It occurs as a fault wedge (W. J. McMillan) upthrust to the northwest into an unfossiliferous assemblage of volcanic arenite, argillaceous rocks, ribbon chert, limestone, and conglomerate locally carrying granitic clasts. This assemblage is on strike with similar beds in Washington with lenses of limestone containing Frasnian fossils. Misch (1966) considers that the lithology of the complex resembles that of the Yellow Aster Complex of Washington which he regards as a correlative of similar rocks unconformably underlying Middle Devonian clastic sediments, limestone, chert, and spilitic volcanics on San Juan Islands.

Ice River Complex

Intrusive into the shales and limestones of the Cambro-Ordovician Goodsir Formation and the underlying Ottertail limestone are two small laccoliths of alkaline rocks of the Ice River Complex (Allan, 1954). Skarn and marble occur at the upper and lower contacts. Petrologically, the complex comprises an intergrading series of nepheline- and sodalite-syenite, urtite, ijolite, and jacupirangite. K–Ar dates range from 327 to 392 m.y.

Middle and Upper Devonian of Cordilleran Geosyncline

The Middle Devonian of *Northern Yukon* (Fig. VIII-18) is represented by three main facies; the Gossage dolomites, the overlying Ogilvie limestones, and the equivalent Prong's Creek shales (Norris, A. W., 1967). The Gossage, partly of Early Devonian age, is generally 1,500 to 2,000 feet thick and consists of banded, grey to dark grey, fine-grained dolomite and dolomitic limestone. At Mount Burgess, it includes thin chert-pebble conglomerate and red shale that may represent a shoreline facies. The lower contact of the Ogilvie is sharp and probably disconformable. The formation consists of brown to grey fine-grained, partly bioclastic and reefal limestones, locally cherty or dolomitized. The thickness varies considerably being least on Mackenzie Arch and thickest where it represents all Givetian as well as upper Eifelian time, as on Porcupine Platform. The Prong's Creek Formation consists of 2,500 feet of black shale with argillaceous limestone and chert. Black, non-calcareous, rusty-weathering shales of the Canol Formation disconformably or unconformably overlie the Middle Devonian and grade upwards into the shales, siltstone, sandstones, and pebble-conglomerates of the Upper Devonian Imperial Formation. The latter attains a maximum thickness of more than 6,000 feet, and is in part non-marine and coarser in the northwest indicating a source in the Brooks Geanticline in the British Mountains or beneath the Arctic Ocean. Commonly the Imperial Formation has a distinctive mauve colour and the coarse beds contain abundant angular to subangular fragments of black chert and siltstone. Macrofossils are Frasnian in age, but spores in the upper part are possibly of Famennian age.

In *western Ogilvie Mountains* just north of Yukon River, in northwestern Selwyn Basin, the Middle and

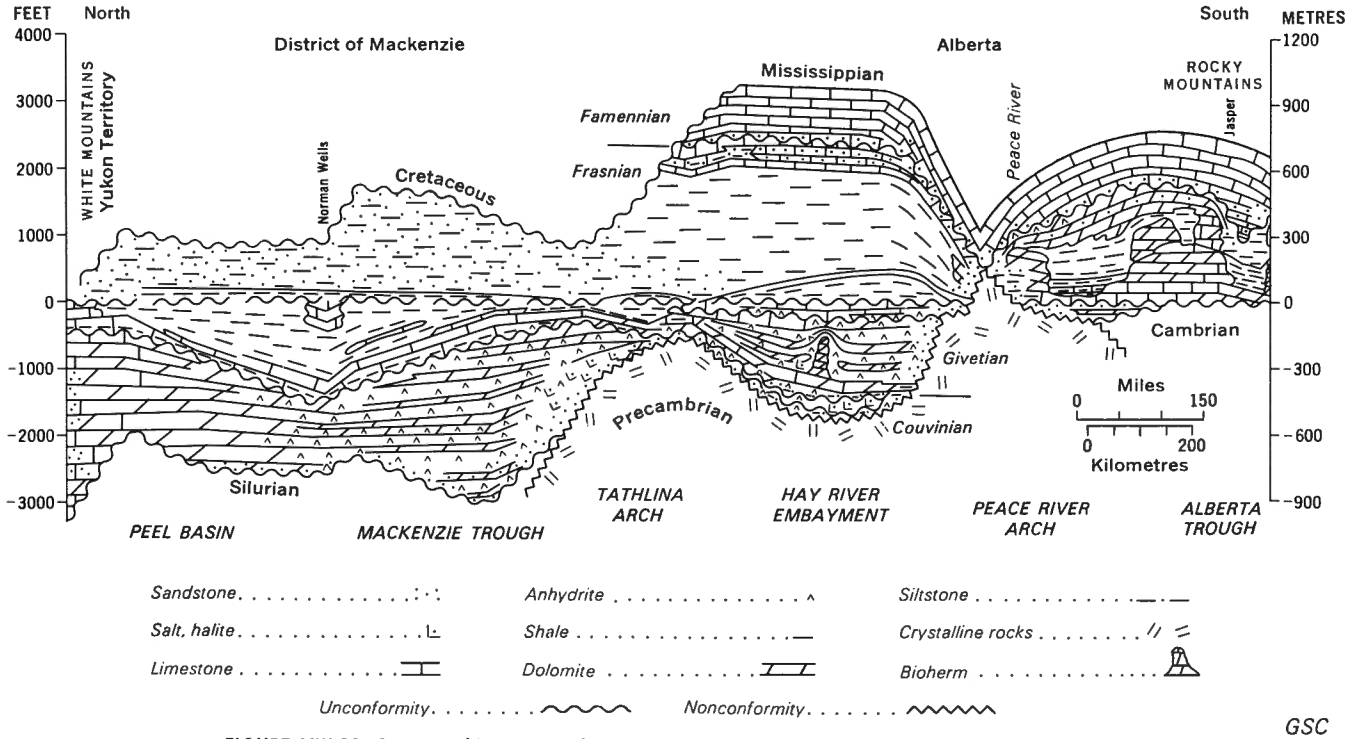


FIGURE VIII-18. Stratigraphic section of Devonian rocks between White Mountains, Yukon Territory, and southern Rocky Mountains, Alberta (by H. R. Belyea).

Upper Devonian McCann Hill chert and shale ranging from 200 to 800 feet in thickness are overlain by more than 3,000 feet of Upper Devonian mudstone, sandstone, gritstone, and conglomerate of the Nation River Formation (Churkin and Brabb, 1965). The latter contains plant fragments and spores and may be partly or mainly non-marine. A succeeding unit of shale and chert, about 3,000 feet thick, is partly Lower Mississippian.

In the subsurface of eastern Peel Basin, according to J. Tassonyi, the Gossage Formation comprises a lower 170 feet of buff aphanitic limestone with waxy green shale, a middle finely crystalline, finely porous dolomite 670 feet thick, and an upper pelletoid aphanitic limestone as much as 250 feet thick. It grades eastward and southward into the Bear Rock Formation of brecciated dolomite and anhydrite, which is orange weathering and hoodoo-forming where it outcrops in *Franklin* and *northeastern Mackenzie Mountains*. The disconformably overlying Hume Formation contains 145 feet of bioclastic and argillaceous silty limestones at the base overlain by 150 feet of grey and greenish grey shale succeeded by coral- and stromatoporoid-bearing bedded limestones as thick as 200 feet. The Hare Indian Formation comprises as much as 700 feet of light, greenish grey calcareous shale grading upwards into the basal, brown, bedded limestones of the Ramparts Formation that form a widespread platform upon which local reefs of the Kee Scarp Member have grown. These are composed of massive bedded, light buff and grey, coral and stromatoporoidal limestones as thick as 400 feet. Disconformably overlying are dark brown bituminous shales of the Canol Formation, which are thickest between the reefs. The Canol grades upwards into greenish grey marine shale, siltstone, and sandstone of the Imperial Formation which is unconformably overlain by Lower Cretaceous or Tertiary strata.

In *southern Mackenzie Mountains* (Douglas and Norris, 1961), banded dark grey dolomites of the Arnica Formation overlie Lower Devonian strata disconformably or unconformably. They range in thickness from less than 500 feet on Mackenzie Arch to about 2,000 feet in Mackenzie Trough (Pl. VIII-4) and eastern Selwyn Basin. The upper part exhibits several facies changes, grading laterally into the calcareous shales and argillaceous limestones of the Funeral Formation, the coarse-grained porous dolomites of the Manetoe Formation, the cryptocrystalline limestones of the Landry Formation, or the breccias and anhydrites of the Bear Rock Formation. To the southwest the Arnica banded dolomites grade through

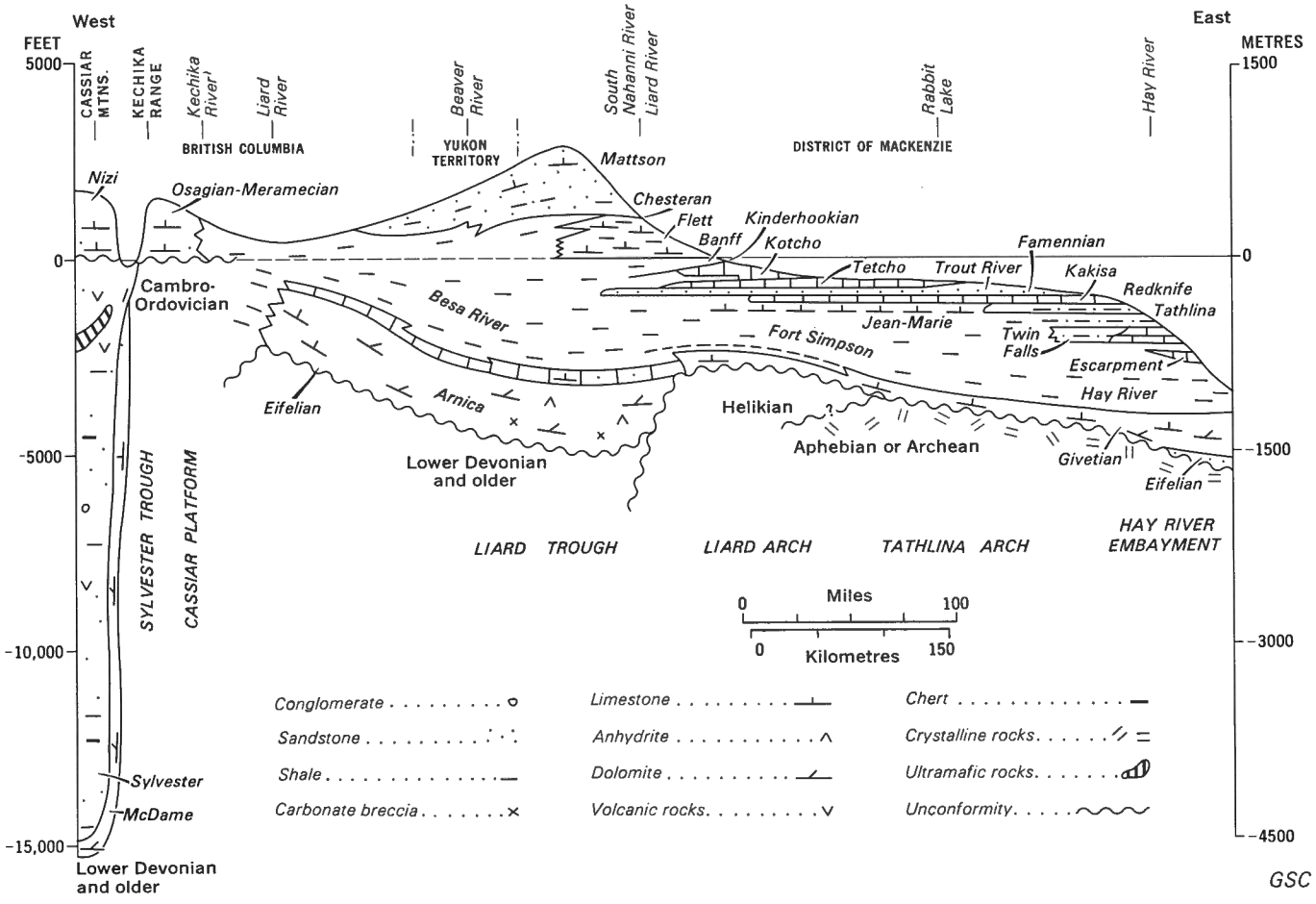


FIGURE VIII-19. Diagrammatic restored section of Devonian and Mississippian rocks between Cassiar Mountains, British Columbia, and Hay River, District of Mackenzie (by R. J. W. Douglas and H. Gabrielse).

diachronous facies of coarse dolomite, crinoidal limestone, and argillaceous limestone into a dominantly shale sequence (Fig. VIII-19). The Headless Formation disconformably overlies the Manetoe, Landry, and Funeral Formations. It consists of recessive argillaceous limestones and shales that grade upwards into the Nahanni Formation and also laterally eastwards into it. The Nahanni is a resistant, well-bedded, dense and bioclastic limestone, remarkably uniform in thickness at about 700 feet, but it may thicken to 1,500 feet near the line of facies change to the Headless shales. Upper Devonian rocks are commonly the youngest Paleozoic exposed. They may be extensively developed in Selwyn Basin also but have been rarely separated from underlying rocks of similar lithology. In Mackenzie Trough, 600 feet of hard black shales of the Horn River Formation is overlain by several thousand feet of grey shale of the Fort Simpson Formation and units of maroon and green siltstone, grey shale, and minor limestone. The latter beds grade eastward to thinly bedded, laminated, fine-grained, green, calcareous siltstone and sandstone with thin beds and isolated masses of reefy limestone.

In *Pelly Mountains* the upper part of the Askin Group includes about 1,000 feet of partly fetid, dark grey dolomite of Middle Devonian age containing *Stringocephalus,* overlain unconformably by early Upper Devonian chert arenite, quartz arenite, siltstones, shales, and pebble-conglomerates (Green, *et al.,* 1960a, 1960b). In the clastic rocks, fossil plant fragments have been found at several widely separated localities. Northeast of Tintina Trench on *MacMillan Plateau* (Roddick and Green, 1961b), a dominantly sombre weathering clastic sequence comprises quartz arenite, chert arenite, chert, quartzite, slate, limestone, and chert-pebble conglomerate. It contains sparse Famennian and early Mississippian faunas. The base of the sequence appears to be a regional unconformity. An assemblage of sandstone, dolomite, and argillaceous limestone, locally more than 1,000 feet thick and containing Eifelian fossils near the top, forms scattered outcrops in southwesternmost Selwyn Basin.

In *Cassiar Mountains* (Gabrielse, 1963) Middle Devonian carbonates comprise a lower member of dark grey, fetid, locally bituminous and vuggy dolomite, characterized by abundant *Amphipora,* 300 to 550 feet thick, and an upper platy limestone member containing *Stringocephalus,* about 200 feet thick. The lower member lies with angular unconformity on Silurian and Ordovician rocks of Kechika Ranges and locally includes a basal chert-pebble and chert-cobble conglomerate. Farther west in Cassiar Mountains the Middle Devonian strata are structurally conformable with underlying, possibly Lower Devonian rocks and are overlain by bedded chert, chert and quartz arenite (Pl. VIII-5), shale, siltstone and pebble-conglomerate forming the lower part of the Sylvester Group. The upper part of the group, which may be as much as 10,000 feet thick, consists of mafic saussuritized and uralitized volcanic rocks, serpentinized ultramafic

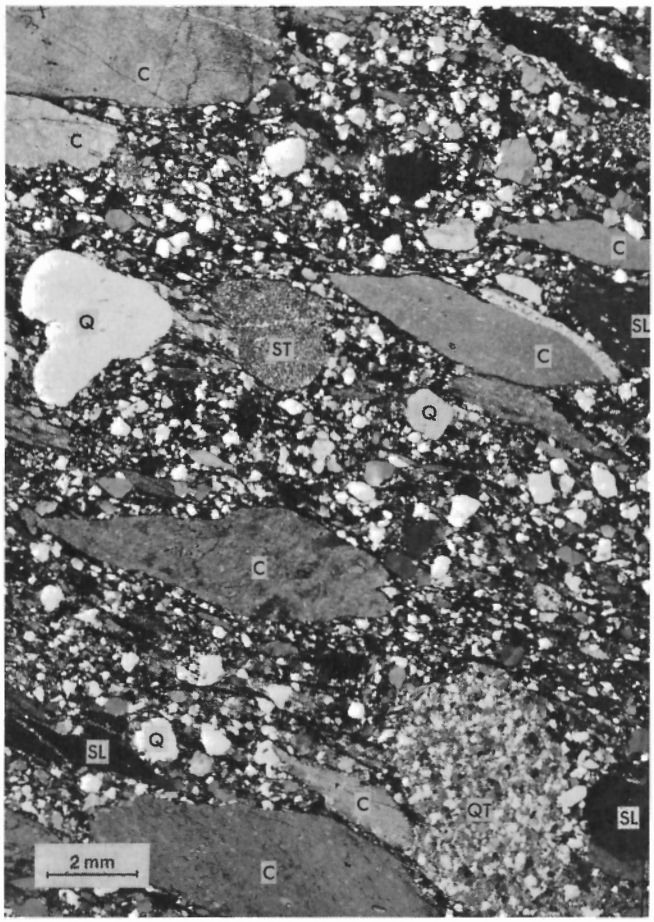


PLATE VIII-5. Photomicrograph of fine pebble-conglomerate of sub-greywacke composition from the Devono-Mississippian Sylvester Group, northern Cassiar Mountains, Yukon Territory. Q—quartz; C—chert; SL—slate; ST—siltstone; QT—quartzite.

rocks, greywacke, and limestone possibly entirely of Mississippian age. Similarly, in Pelly Mountains, the clastic Upper Devonian rocks are overlain by mafic volcanic rocks that are in turn overlain by felsic and mafic volcanic breccias. Locally associated are shattered syenites that may be possibly of Mississippian age. These great thicknesses of Devonian and early Mississippian strata in Cassiar and Pelly Mountains were deposited in the western part of the Sylvester Trough which lies on the site of the former Cassiar Platform, stable until the end of Middle Devonian time.

The Kaskawulsh Group in *St. Elias Mountains* consists of about 2,500 feet of Middle Devonian limestone, dolomite, limestone breccia, slate, and argillite overlying 2,000 feet of argillite, slate, and quartzite that may be older than Middle Devonian (Wheeler, 1963). An overlying assemblage including volcanic rocks and chlorite schist is partly of Late Devonian age.

In *northern Rocky Mountains* banded dark grey and grey, Middle Devonian dolomites and brecciated dolomites, locally barite-bearing, are probably correlative with the

Arnica Formation (Fig. VIII-11b). They increase in thickness westerly from about 1,000 feet in Stone Range to more than 1,500 feet in Terminal Range (Hughes, 1963) and thin southward, becoming sandy. They are overlain conformably by about 1,000 feet of limestone correlative in the north with the Nahanni Formation and become progressively younger at the top as the Peace River Arch is approached including equivalents as young as the Slave Point Formation near Halfway River. The limestone varies from brown calcilutite to coarse-grained bioclastic calcarenite, partly dolomitized. The Middle Devonian, if present farther west, is part of an undivided chert, shale, and siltstone sequence in the Kechika Trough. The Middle Devonian is also represented by the Besa River Formation that includes Upper Devonian and Carboniferous equivalents (Kidd, 1963). It is characteristically thin, about 1,000 feet thick, and composed of black, pyritic, non-calcareous shales and siliceous mudstones. Between Gataga and Kechika Rivers sandstone, shale, and conglomerate, locally with well-developed crossbedding, are more than 1,000 feet thick and lie unconformably on Cambro-Ordovician strata. The age of this assemblage is uncertain but it is believed to be correlative with the widespread Upper Devonian clastic rocks of Cassiar Mountains. The Besa River Formation extends south of Pine Pass where part grades into the carbonates of the Famennian Palliser Formation or the Mississippian Kindle Group. The shale facies extends still farther south, to Athabasca and Bow Rivers grading into the carbonate reefs of the Frasnian Fairholme Group. The basal Fairholme overlaps the Middle Devonian along Alberta Arch resting unconformably or disconformably on Ordovician or Cambrian strata.

The Upper Devonian in the *southern Rocky Mountains* is represented by the Frasnian Fairholme Group and the unconformably overlying Famennian Sassenach and Palliser Formations (McLaren, 1955). The Fairholme embraces two megafacies, the carbonate and clastic facies, which grade laterally into each other either abruptly or by intertonguing (Fig. VIII-20). The basal formation of the clastic facies, the Flume, is 100 feet of grey argillaceous limestone and dolomite, cherty and stromatoporoidal in the upper part. It is succeeded by 50 feet of argillaceous thinly bedded limestone, the Maligne Formation, which grades upwards into the black, nodular, and calcareous shales of the Perdrix. The contact with the overlying lower Mount Hawk Formation is transitional, the shales becoming interbedded with argillaceous limestone; the two formations commonly total about 1,000 feet. The Cairn Formation is the basal unit of the carbonate sequence, the lowermost beds being similar to the Flume and the upper part comprising as much as 600 feet of dark grey, thick-bedded, fine-grained, dolomitic limestone and dark brown, coarse, biostromal dolomite, stratigraphically equivalent to the Maligne and Lower Perdrix Formations (Fig. VIII-21). The succeeding Southesk Formation is divisible into three members, the Peechee and Arcs consisting mainly of light grey, bedded, aphanitic to finely crystalline limestone or dolomite separated by the Grotto Member of dark brown, finely crystalline, dolomite with abundant corals, stromatoporids, and *Amphipora*. The Southesk, and particularly the Peechee Member, is less extensive than the Cairn, locally forming pinnacle reefs (Pl. VIII-6), reflecting general, rapid, upward reef growth. It is partly organic, but mainly consists of lime sand. The Arcs is somewhat more extensive, the reef detritus and the laminated silty carbonates of the overlying Ronde Member overlapping the basinal lime muds of the Mount Hawk and reflecting a regression that locally left the carbonates emergent and subject to erosion at the end of the Frasnian.

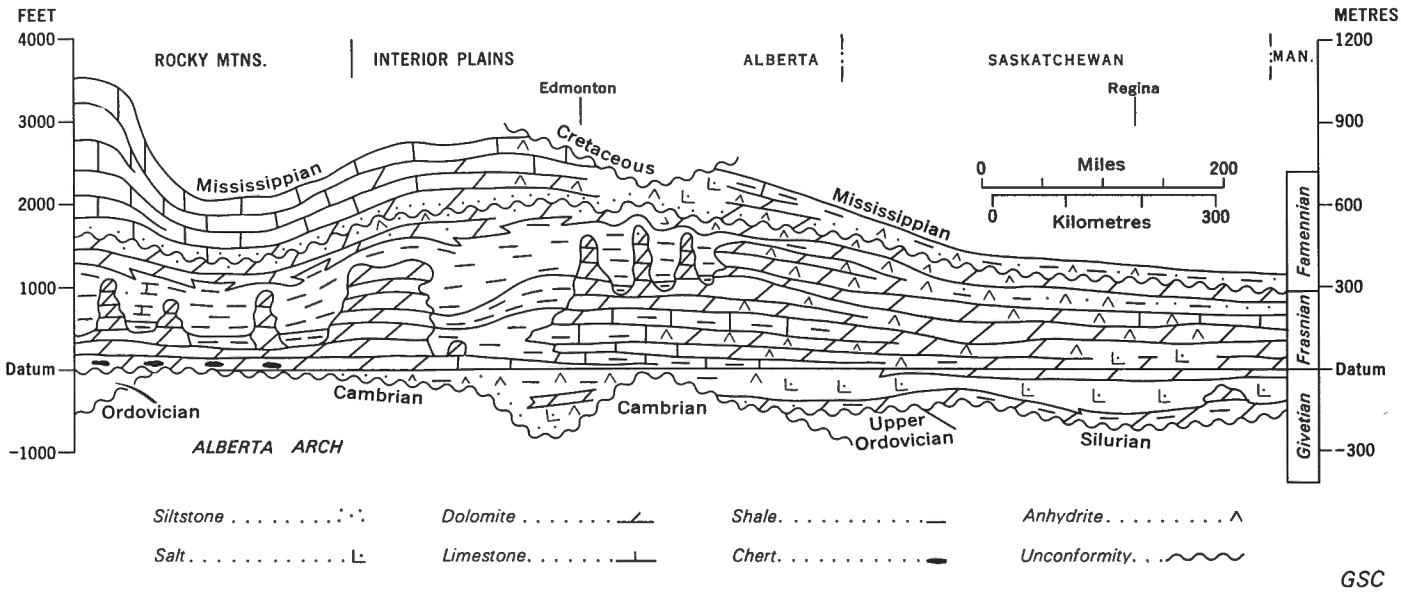
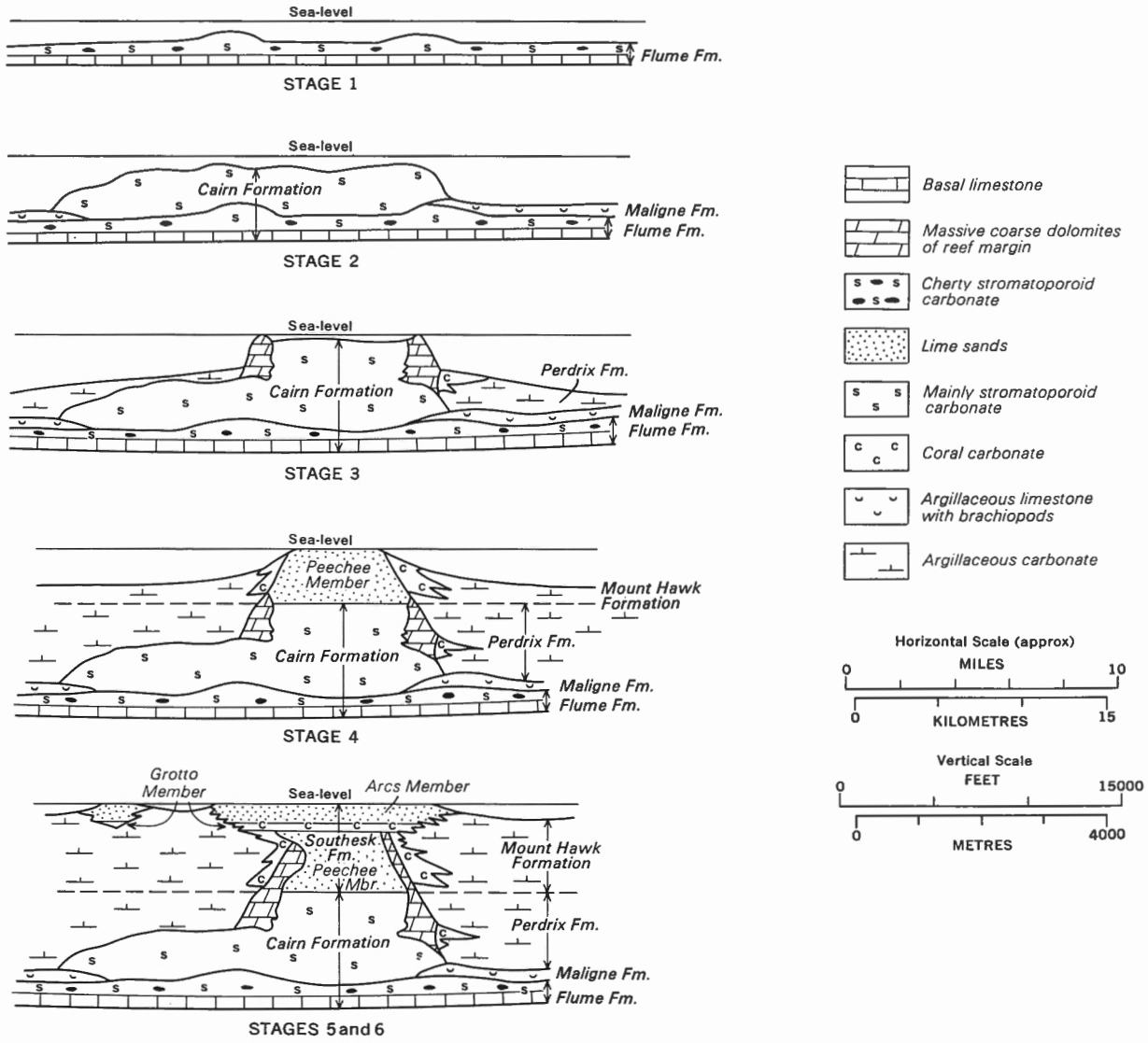


FIGURE VIII-20. Stratigraphic section of Devonian rocks between the southern Rocky Mountains and Manitoba (by H. R. Belyea).



TRANSGRESSIVE PHASES *(progressively increasing water depth in basin)*
Stage 1: Main transgression and development of stromatoporoid carbonate platform (Flume Formation)
Stage 2: Growth of stromatoporoid biostromes (lower half of upper Cairn Formation)
Stage 3: Growth of marginal bioherms (upper half of upper Cairn Formation)

STABLE PHASES *(filling of basins surrounding reef complex)*
Stage 4: Development of lime sand bank (Peechee Member, Southesk Formation)
Stage 5: Growth of coral and Amphipora biostrome (Grotto Member of Southesk Formation)
Stage 6: Extensive development of lime sand bank (Arcs Member of Southesk Formation)

REGRESSIVE PHASE *(not illustrated)*
Stage 7: Influx of quartz silts and spreading of carbonate bank (Ronde Member of Southesk Formation)
After an interval of non-deposition the Famennian transgression occurred

GSC

FIGURE VIII-21. Main depositional stages in the formation of the Upper Devonian, Frasnian, Miette reef complex (Mountjoy, 1965).

The Famennian Sassenach Formation is variable in thickness, the thicker sandstones and silty dolomites, as thick as 600 feet, filling irregularities in the erosion surface. The Morro Member of the overlying Palliser Formation is a massive, cliff-forming, grey aphanitic limestone mottled with buff dolomite, which in the central Foothills grades to fine-grained brown dolomite. The uppermost 100 feet are the Costigan Member, an argillaceous limestone that is abruptly overlain by the Exshaw shale of probable Mississippian age.

Strata representing the early Frasnian, the Beaverhill Lake Group of the subsurface, are only locally present in the southern Rocky Mountains as the thin basal beds of the Flume and Cairn Formations, or as the Hollebeke



PLATE VIII-6. Upper Devonian succession and remnants of protuberances along Peechee reef front, Flathead Range, southern Rocky Mountains, Alberta and British Columbia. Looking west. P—Palliser Formation; s—Sassenach Formation; Sa—Arcs Member, Southesk Formation; Sg—Grotto Member; Sp—Peechee Member; M—Mount Hawk Formation; B—Borsato Formation; H—Hollebeke Formation; E—Cambrian Elko Formation.

Formation of Flathead Range. The latter consists of 500 feet of dark grey, very finely crystalline limestone and laminated argillaceous and silty dolomite with beds of carbonate breccia near the base. It overlies the Cambrian Elko Formation either disconformably or with indications of channelling. The succeeding Borsato Formation of the Flathead region is lithologically similar to the upper Cairn Formation.

In *southwestern Rocky Mountains,* in Stanford and Hughes Ranges, the Middle Devonian Cedared Formation unconformably overlies the Silurian and grades laterally into the Burnais gypsum (Belyea and Norford, 1967). It consists of as much as 700 feet of light grey and brown, sandy aphanitic dolomite, agrillaceous dolomite, breccia, sandstone, and red shale. The conformably overlying Harrogate Formation consists of as much as 300 feet of dark brown limestone, shale, and dolomite with a fauna similar to that of the Pine Point Formation near Great Slave Lake. In Lussier syncline east of Hughes Range, these beds are overlain by shaly limestones equivalent to the basal Fairholme Group, which are succeeded by dark grey shale and siltstone, presumably of Upper Devonian age, as they are overlain by cherty limestones of the Mississippian Banff Formation.

On the west flank and floor of *Rocky Mountain Trench* south of Moyie fault, the Devonian sequence consists of a basal fanglomerate, with slabs of Purcell strata, that rests unconformably on the Purcell lava and the Gateway Formation (Leech, 1962b). It is succeeded by a sandstone containing fish remains and thence by thick gypsum, dolomite, and solution breccia. The overlying beds are fossiliferous limestones equivalent to the Maligne Formation of the Fairholme Group and strata similar to the other formations of the group. Similar, but unfossiliferous fanglomerates occur along the trace of Moyie fault. In northern Purcell Mountains thin, incomplete sequences suggest that the east side of Purcell Arch was periodically emergent during the Devonian. Unfossiliferous red and green shale of the Mount Forster Formation probably represents a marginal facies of the Burnais gypsum. The Mount Forster disconformably overlies the Beaverfoot Formation and is disconformably

overlain by 200 feet of limestone and gypsum of the early Upper Devonian Starbird Formation (Reesor, 1957).

The age of the Broadview and Jowett Formations at the top of the Lardeau Groups in *Selkirk Mountains* and of the Snowshoe Formation in *Cariboo Mountains* is not known. The Broadview and Snowshoe Formations are characterized by poorly sorted, quartz-rich, slightly feldspathic arenites and represent a profound change from the preceding argillaceous régime. If these units are Upper Devonian or even early Paleozoic then it is difficult to contemplate an eastern source. As such clastics are lacking in eastern Purcell Mountains and the Snowshoe Formation is fine grained and more calcareous in the east (Sutherland Brown, 1957), the source for these formations must have been west of Lardeau Trough. An alternative is that the Broadview Formation is equivalent to the Horsethief Creek Group. The arenites from the two units are remarkably similar and the greenstone in the former would then be correlative with the Irene Formation. Structural relations lend support, particularly in the western part of Kootenay Arc where the Broadview is separated from the underlying Index Formation by a slide (P. B. Read). Broadview rocks were deformed prior to the deposition of the Mississippian Milford Group (Wheeler, 1966b). It is possible, therefore, that the Broadview Formation is part of the Horsethief Creek Group thrust over the lower formations of the Lardeau Group during the Caribooan Orogeny prior to the deposition of the Milford Group. In Cariboo Mountains, Sutherland Brown (1963) has commented on the similarities of arenites in the Snowshoe Formation and Kaza Group but maintains that they are separate stratigraphic units.

Middle Devonian of Interior Platform

Eifelian (Fig. VIII-14). The formations exposed in Franklin and southern Mackenzie Mountains are present in the Interior Platform in southwestern District of Mackenzie and northeastern British Columbia, but are much thinner. Dolomites, apparently coeval with the Arnica Formation, are grey and light brown, banded and mottled, cryptocrystalline to finely crystalline, and locally brecciated. Floating sand grains are common; some form sandstone beds near the base. The dolomites grade eastward into the Cold Lake salt and associated red beds and carbonates that border Tathlina Upland on the south and into about 1,200 feet of anhydrite, halite, and dolomite to the north. The latter thin eastward to 300 feet of breccia and shaly anhydritic dolomites unconformably overlying the Upper Ordovician northwest of Great Slave Lake. The overlying Manetoe dolomite, southwest of Tathlina Arch, consists of medium to coarsely crystalline, vuggy dolomite that grades upwards and laterally to light brown, fine-grained, pelletoid limestone. It is missing or grades eastward in northern Alberta, into argillaceous, anhydritic dolomites and anhydrite and southward into sandy anhydrite and shales of the lower Chinchaga. The upper part and top of this unit commonly contains green or blue-grey, waxy, pyritic, non-calcareous shales as irregular layers enclosing blebs or pebbles of limestone and as fracture fillings. A sandstone occurs at this horizon on the flank of Tathlina Upland. The shale suggests an hiatus in deposition accompanied by fracturing, leaching, and soil formation. It is considered to mark the Eifelian–Givetian boundary in the subsurface but no faunal control is available.

In Elk Point Basin, the Eifelian is probably represented by the lower part of the Elk Point Group. The lowest formation, the Lotsberg, occurs in central Alberta and may be represented in basal sandstones and red beds north of Peace River Arch. It may be Eifelian or older. It consists of a maximum of 500 feet of sandstones overlain by orange-red, sandy, silty dolomitic and anhydritic claystones with anhydrite and one, or locally two, halite beds as thick as 400 feet in central Alberta. The halite grades laterally westward through interbedded anhydritic dolomite and mudstone, to the sandstones of the basin margin. It is succeeded by a red claystone with high gamma radiation, followed by the light grey to light brown, very fine grained, slightly argillaceous, thin-bedded limestones of the Ernestina Lake Formation. This formation, although not present over Peace River Arch, is present between the arch and Tathlina Upland as a finely crystalline, locally porous dolomite with intergranular porosity carrying the same fauna as the limestones south of the arch. These beds are transitional upward to light coloured, massive anhydrite succeeded by red claystones and the Cold Lake salt. The salt occurs in two basins, one flanking Tathlina Upland to the south, east, and north and the other between Meadowlake Escarpment and Peace River Arch (Fig. VIII-14). Red beds were deposited over the eastern part of the Peace River Arch. The Cold Lake salt is thought to have been deposited in lagoons, as much as 1,000 miles long, or in lakes of a playa type into which streams drained from the bordering ridges. The Cold Lake salt seems to fill Elk Point Basin to the level of the Upper Ordovician beds forming Meadowlake Escarpment; the Givetian upper part of the Elk Point Group overlaps it.

Givetian (Fig VIII-15). On the craton the thick carbonates of the Nahanni and Hume Formations converge southwards and eastwards. In the subsurface, coeval beds consist of fine-grained and pelletoid limestones, locally fossiliferous, that grade eastward to cryptocrystalline, light brown, brittle limestones, and thence into anhydrite and dolomite of the upper part of the Chinchaga Formation. Equivalent beds farther southeast are argillaceous dolomites with shale laminae of the Contact Rapids Formation. This formation is probably equivalent to Members 1 and 2 of the Methy Formation of eastern Alberta. In Saskatchewan the Contact Rapids grades into 50 feet of red and green siltstone, shale, and grey argillaceous carbonate of the Ashern Formation which un-

conformably overlies the Silurian and Ordovician (Fig. VIII-20).

South and east of the carbonate front, the base of the carbonate facies rises progressively higher stratigraphically and forms a platform carbonate that covers Tathlina Arch and Elk Point Basin as far south as the Manitoba outcrop. This carbonate, the lower member of the Keg River Formation, consists of 100 to 170 feet of brown, fine-grained, argillaceous dolomite and limestone, with brown, argillaceous limestone and bituminous shale in the upper part. It thins southeast to less than 50 feet where it overlaps the Meadowlake Escarpment in Saskatchewan. Above the platform carbonate the sediments are differentiated into laterally contiguous shale, carbonate, and evaporitic facies. The shale facies, the Horn River Formation, consists of about 100 to 300 feet of brownish grey bituminous shales, interbedded with dark brown, fine-grained limestones and greenish grey calcareous shales. It thickens into northeastern British Columbia where it consists of 50 to 100 feet of basal, dark brown to black, bituminous shale and brown, bituminous, fine-grained limestone, the Evie Member, overlain by about 550 feet of grey, calcareous shales of the Otter Park Member succeeded by 20 to 150 feet of black, bituminous, locally pyritic and siliceous shales of the Muskwa Member. The base of the Upper Devonian lies within the Horn River Formation, possibly at the base of the Muskwa Member.

The contact between the shale and carbonate facies is abrupt and follows a sinuous pattern that outlines platform and basin areas. Margins of platform areas are characterized by what appears to be a persistent carbonate barrier, composed in part at least of reef and reef detritus and constituting the Pine Point and Sulphur Point Formations. Small carbonate banks or "island reefs" occur on the basin side of the barrier. Locally, along the front, the limestones are replaced by as much as 1,200 feet of white, coarsely crystalline, vuggy dolomite that completely destroys the original textures and may preclude recognition of the stratigraphical equivalence of adjacent formations within it. Behind the barrier, the equivalent beds comprise the Keg River carbonates and overlying Muskeg evaporite–carbonate sequence. The upper member of the Keg River Formation is divided into the Rainbow (reef) Member, as thick as 800 feet, which may in part be continuous with the Muskeg Formation, and a non-reefal, fine-grained, brown limestone, locally crinoidal, about 150 feet thick. In the Rainbow Member the wide variety of carbonates includes organic reefs and reef detritus on the flanks of atolls or banks and back-reef to lagoonal non-skeletal calcarenites and calcilutites. Locally, they are altered to dolomite. The upper part, locally stromatoporoidal, is predominantly laminated dolomite, possibly of shallow-water algal origin. Eastward in Saskatchewan the Keg River Formation is termed the Winnipegosis Formation, which includes a lower unit of fine-grained, argillaceous dolomite unit less than 30 feet thick and, equivalent to the upper Keg River, light yellow-brown, cryptocrystalline to finely crystalline dolomite. The Winnipegosis locally contains stromatoporoids, bryozoans, and corals, but it generally is barren and exhibits relict pelletoid, oölitic, and pisolitic structures. Tight beds are interspersed with beds characterized by intercrystalline or vuggy porosity. On outcrop in Manitoba the stromatoporoid reefs are draped with bedded dolomites. They presumably also occur in the subsurface where thicknesses vary from 50 to at least 345 feet. The reefs have been interpreted as atolls by Edie (1959), as patch reefs on a fringing carbonate bank by Grayston, Sherwin, and Allan (1964), and as banks partly supported by an organic framework by Jones (1965). The Winnipegosis grades southward to light grey, finely crystalline dolomite with a thin, greyish red, silty dolomite at the base, before reaching the depositional edge in Montana.

Throughout Elk Point Basin, between the topographically high reefs and atolls, salt was deposited. In the northwest, the Black Creek salt at the base of the Muskeg Formation does not cover Tathlina Arch indicating that it either was not deposited there or was eroded later. Southeastward, the Muskeg grades into the Prairie Formation, 250 to 600 feet thick in Saskatchewan and a maximum of 1,000 feet in northeastern Alberta. In part, this variation is caused by infilling of the irregular surface of the Keg River and the Winnipegosis carbonates but it may also be due to subsidence or faulting during deposition. Sylvite and carnellite, economically important, occur in the upper 100 to 200 feet south of Saskatoon. Along the eroded northeast margin the Prairie evaporites have been dissolved and lie back of the Winnipegosis reefs and bedded dolomites. Towards the south and west margins of Elk Point Basin, extensive channelling, brecciation, and slumping have been caused by solution of halite. Both the Winnipegosis and Prairie Formations grade through a thin sequence of brown argillaceous limestone, dolomite, and shale, to red shale and sandstones near the shoreline. Marine conditions must have extended farther northeast than the present subcrop.

Overlying the Pine Point, Muskeg, and Prairie Formations is a thin green shale in the northwest that grades to red claystones in the southeast; these beds represent a depositional break or disconformity. The overlying Sulphur Point Formation to the north and west contains limestones interpreted as having been deposited in reefs, lagoons, or intermittently exposed shallows. Stromatoporoids and corals occur in situ or as detritus in a fine- to coarse-grained, largely skeletal, light grey to buff matrix. The formation is 300 feet thick in the barrier and fringing reefs, but thins on the crest of Tathlina Arch to 25 feet of brecciated limestone and green and blue-grey, non-calcareous pyritic shale suggesting subaerial exposure. Where altered to coarsely crystalline dolomite the equivalent beds are termed the Presqu'ile Formation. The Sulphur Point grades southward to aphanitic lime-

stone, evaporitic dolomite, and anhydrite thinning to the southeast. It is represented by the Gilwood sandstones in the vicinity of Peace River Arch. It thins to about 10 feet of anhydrite in east-central Alberta, possibly the equivalent of the Dawson Bay Formation of Saskatchewan as both contain *Stringocephalus*. The Dawson Bay Formation in Saskatchewan comprises a basal unit 30 feet thick of mottled red and green dolomitic claystone, that gradationally overlies the Prairie evaporite and is interpreted as a soil. It grades upwards into shaly limestone and dolomite with a coral–brachiopod fauna, and overlying, buff and brown, fine- to coarse-grained skeletal limestones containing stromatoporoids and corals. The latter are, in part, altered to euhedral, porous dolomite largely filled with halite. Succeeding, buff, brown, bituminous, cryptocrystalline and pelletoid limestones are interbedded with anhydrite or halite. The Dawson Bay Formation is overlain by a thin red and green shale apparently coeval with the green shale of the Watt Mountain overlying the Sulphur Point. It is thickest, as much as 230 feet, in southeastern Saskatchewan, oversteps the Winnipegosis depositional edge in North Dakota, and thins to about 10 feet of dolomite and anhydrite in southwestern Saskatchewan.

The Slave Point Formation thins southeastward from a maximum of 500 feet at the carbonate front to zero against the Peace River Arch and to less than 50 feet in east-central Alberta. It has not been identified in Saskatchewan. Near the front it contains globular stromatoporoids set in a coarse skeletal calcarenite matrix or present as fragments in calcilutites and pelletoid limestones recrystallized locally to coarse, white, vuggy dolomite. Southeast of the front the stromatoporoid-rich beds decrease in abundance, the rock becomes bedded, largely cryptograined to fine grained, and contains chert nodules and scattered ostracods and brachiopods. Evaporitic dolomite and anhydrite constitute the Fort Vermilion Member at the base which increases in thickness eastward; the Slave Point Formation is represented by 60 feet of anhydrite and a few feet of overlying limestone in the vicinity of Fort McMurray.

Upper Devonian of the Interior Platform

Frasnian (Fig. VIII-16). The Frasnian rock sequence is the result of oscillatory but essentially continuous deposition. The basal beds are still undated locally but are included herein with the Frasnian. They consist predominantly of black, pyritic, locally siliceous shales of the Muskwa Formation, which fill irregularities resulting from post-Givetian warping, faulting, and erosion and may represent westward convergence of the lower part of the overlying Fort Simpson Formation or may include one or more disconformities that cut down westward. In Great Slave Lake area, where Frasnian carbonates occur at a lower stratigraphic horizon than they do to the west, the tongue of shale is named the Hay River Formation. It is about 1,200 feet thick, and like the Fort Simpson, which ranges from 2,000 to 2,500 feet in thickness, consists of greenish grey to greyish green shales with thin siltstone and silty carbonate layers and locally, dark grey and black shales. Both formations are transitional into the overlying carbonates by interbedding of shales with skeletal calcarenites and coral- and stromatoporoid-rich lenses. The Hay River is overlain by the Twin Falls Formation, about 500 feet of algal–stromatoporoid–coral limestone bioherms, biostromes, and associated calcarenites, pelletoid and cryptograined, clotted lagoonal limestones which grade south to the Grosmont dolomite and upper part of Leduc Formation. The Twin Falls is overlain by 450 feet of silty limestones, calcareous siltstones, and shales of the Tathlina Formation and the Redknife limestones and siltstones that thicken westward and include a basal, coral-bearing, bioclastic to reefoid limestone, 50 feet thick, the Jean Marie Member. The Jean Marie thickens southwestward at the expense of the Fort Simpson shales before disappearing as a separate carbonate rock unit. The uppermost Frasnian, Kakisa Formation, 200 feet thick, contains coral–stromatoporoid reefs and skeletal calcarenites as well as pelletoid and cryptograined limestones, and grades west to greenish grey shales. It was subjected to erosion before deposition of the Famennian Trout River Formation. The Redknife and Kakisa converge southward and change to dolomite over Peace River Arch; more southerly equivalents are the uppermost Woodbend and Winterburn Groups.

The Beaverhill Lake Formation and the almost equivalent Waterways Formation of east-central Alberta are 500 to 700 feet thick. The basal, skeletal, calcarenitic limestone of the Beaverhill Lake, which is not included in the Waterways, is abruptly overlain by the Firebag Member which consists of dark brown calcareous and greenish grey shales that are succeeded by light yellow, brown, and buff, very fine to medium grained limestones of the Calumet Member. Greenish grey shales of the Christina Member grade up to the complex of limestone facies of the Moberly Member, which includes stromatoporoid-rich biostromes, skeletal calcarenites, clotted limestone, bedded anhydrite, and greenish grey shale. The Moberly is overlain by interbedded shale and fine-grained limestone of the Mildred Member.

The interbedded shale and limestone facies of the Beaverhill Lake changes along an irregular front to a shelf or platform carbonate facies. The carbonates flank and overlap the Peace River and Alberta Arches and the broad shelves in Montana, North Dakota, and southern Manitoba. The platform front at that time lay mostly in the subsurface of west-central Alberta. A widespread, dark brown, bedded, fine-grained limestone carrying dispersed stromatoporoids, corals, and *Amphipora* with a basal brachiopod–crinoid layer forms a widespread platform on which developed isolated reef-fringed banks as much as 500 feet thick. These are predominantly limestone, although grading southwest to dolomite and possibly forming a continuous but irregular, vuggy, reefoid, dolo-

mite barrier from north of Athabasca River in Jasper National Park to eastern Saskatchewan. The reef-fringed banks, the Swan Hills Member, contain a great variety of carbonate rock types (Murray, 1965). A framework of massive and tabular stromatoporoids, solenaporid algae, and some corals forms a wave-resistant barrier that became progressively restricted to mound-shaped structures. Skeletal debris flanks the reefs as talus both seaward and lagoonward of the reefs, and forms bedded deposits that may separate reefs low in the sequence from those above. The back-reef calcarenites grade to a variety of lagoonal deposits, mostly calcilutites and non-skeletal calcarenites carrying *Amphipora* and dendroid stromatoporoids. Green, waxy, pyritic shale breaks occur locally.

The reef and bank and vuggy dolomite facies in Alberta grade eastward into Saskatchewan to bedded dolomites and evaporites of the Souris River Formation, which encompasses approximately the same stratigraphic interval as the Beaverhill Lake. The Souris River Formation, about 400 to 600 feet thick in central Saskatchewan, comprises (Lane, 1965) from the base up: A unit—red and green claystone, locally brecciated (First Red Beds of the Elk Point Group); B unit—cryptograined, argillaceous or dolomite limestone; C unit—brown, dolomitic, fine- to medium-grained, skeletal and pelletoid limestone, containing stromatoporoids and corals overlain by and laterally equivalent to D unit (the Davidson halite) and forms the barrier that permitted the latter to be precipitated. The Davidson halite grades upwards and westwards to anhydrite. The upper part of the Souris River consists of cyclical alternations of grey, green, and red shale, fine-grained, fossiliferous limestone and anhydrite, and locally, halite, succeeded by fine-grained, argillaceous limestone, and grey shales.

The succeeding, middle Upper Frasnian rock sequence is represented by the greater part of the Fort Simpson shale facies north of the Peace River Arch, by the platform carbonate and reef-fringed bank facies of the Woodbend and Lower Winterburn Groups in central Alberta, and by the evaporite facies of the Duperow and Birdbear Formations, Saskatchewan. The shale facies covers the Beaverhill Lake Platform of western Alberta and embays the platform facies in west-central Alberta (Fig. VIII-16). The basal platform carbonate, the Cooking Lake Formation, consists of interbedded fine-grained limestone with varying amounts of skeletal debris, fine- to coarse-grained skeletal calcarenites, locally with abundant stromatoporoids, *Amphipora*-rich beds, and pelletoid beds. Stromatoporoids increase in abundance towards the front and form biostromes and patch reefs over broad areas. Away from the front, pelletoid and very fine grained limestones become more numerous, and are interbedded with anhydrite, as the latter increases southward. A prominent brachiopod–crinoid shaly limestone and shale zone in the middle of Cooking Lake extends across Saskatchewan. A similar zone occurs at the top of this formation. The shale equivalent to the Cooking Lake Formation, the Majeau Lake Formation, is present between the Windfall platform and Rimbey–Meadowbrook reef chain. It consists of a black shale that grades up to greenish grey shales that filled the basin and spread over the Cooking Lake Platform. Subsequently carbonate deposition was restricted to areas controlled by organically bound banks and reefs that could present a rigid framework to the seas. These post-Cooking Lake carbonates, the Upper Cairn Formation and upper part of the Wymark Member of the Duperow Formation, form a continuous, though irregularly arcuate front from the Rocky Mountains to southeastern Saskatchewan and, as the lower part of the Leduc Formation, form the base on which the grey dolomites of the reef chains accumulated. These strata consist predominantly of brown stromatoporoid biostromes, reefs, and bars that are interbedded with *Amphipora*-rich beds, skeletal calcarenites, and fine-grained laminated limestones. They are commonly altered to fine- to coarse-crystalline brown dolomites. The stromatoporoid beds grade away from the front to bahamites or pelletoid limestones and fine-grained limestones, commonly dolomitized. South and east of the Drumheller front these facies intertongue with cyclically alternating dolomites and evaporites chiefly anhydrite but locally salt. Tongues of black and dark grey shales penetrate the carbonate front at the top of the Cairn Formation. The Cairn and its equivalents extended farther south than any other part of the Frasnian and probably represent the maximum Upper Devonian transgression. In northeastern Alberta and southwest of Great Slave Lake, stromatoporoid–coral lenses, reefs, and calcarenites of the Escarpment Member and lower part of the Twin Falls Formation intertongue with shales, the base of the carbonate facies rising stratigraphically westward as the carbonate tongues grade to shale. Carbonate deposition on the southeast flank of the Peace River Arch also began at this time.

Distribution of the overlying carbonates is largely related to these carbonates. They similarly form an almost continuous front from the Rocky Mountains across southern Alberta (Peechee Member of Southesk Formation) into Saskatchewan (Seward Member of Duperow Formation) and form the Upper Leduc of the reef chains, the fringing reef of the Peace River Arch, and the Grosmont dolomite of northeastern Alberta which grades northward into the Twin Falls limestone of the Great Slave Lake area. Contact with the adjacent shale facies may be abrupt or may interdigitate. These strata most commonly consist of 200 to 500 feet of light grey, medium to coarsely crystalline, vuggy dolomite in which original structures have been almost completely destroyed. Where preserved as limestone, reefs, organic debris, non-skeletal and pelletoidal calcarenites, and laminated and clotted lime mudstones have been recognized. A typical, isolated carbonate mass (Klovan, 1964) is made up of discontinuous organic reef facies on the flanks of the structure, but about 20 to 40 feet below the surface. The reefs are built of massive stromatoporoids, and are separated and

flanked seaward by coarse reef detritus with tabular stromatoporoids, and lagoonward by skeletal calcarenites including reef detritus with massive stromatoporoids. In the protected back reef, or lagoon, non-skeletal calcarenites, locally bored, algal-laminated limestones, and *Amphipora*-rich beds are the most common facies and discontinuous green waxy shales occur as laminae or are disseminated through the rock.

The massive-bedded, locally crossbedded, dolomites forming the front of platform areas also include reefs. They grade south and east of the front in southern Alberta and Saskatchewan to cyclically alternating shaly limestone and shale carrying a brachiopod–coral–crinoid fauna, dolomite, and anhydrite. The shaly fossiliferous beds are absent over the Swift Current Platform and in Montana and North Dakota. The uppermost cycle changes to red shale across southern Saskatchewan. The equivalent Grosmont of northeastern Alberta similarly acts as a barrier behind which the Hondo evaporites and siltstones accumulated.

The shale facies of the Woodbend Group east of the Rimbey–Meadowbrook chain rests on the Cooking Lake Platform carbonate. At the base, the Duvernay Formation, 180 feet thick, consists of dark brown limestone, black, pyritic, in part silty shale, thin sandstone laminae with corrosion surfaces commonly overlain by pyrite and fragments of underlying rock. Northward it is interbedded with greenish grey, calcareous shale and grey limestones that pass northward to become part of the greenish grey shales of the Hay River Formation. South and west of the Rimbey–Meadowbrook chain it and the Majeau Lake Formation condense to about 50 feet of black shale that possibly includes a disconformity. The Duvernay grades up to the greenish grey, poorly fossiliferous shales of the lower part of the Ireton Formation, 500–900 feet thick. The upper part of the clastic sequence represented by the Ireton, Mount Hawk, and Fort Simpson Formations shows the influence of stabilization of the basin and gradual marine regression. Tongues of carbonate extend from the banks that almost surround the basin into the clastics of the upper Woodbend and lower Winterburn Groups, and the Mount Hawk Formation of Rocky Mountains. The lower, more restricted tongue, the Camrose of the subsurface and Grotto Member of the Southesk Formation, contains abundant corals and stromatoporoids, locally forming reefs, in a dark agrillaceous bituminous matrix. The succeeding Nisku Formation of the Winterburn Group of Alberta and the essentially coeval Birdbear Formation of Saskatchewan and the Arcs Member of the Southesk Formation consist of light grey to light brown, crystalline dolomites that cover the shelf carbonates and extend over the Ireton shales of eastern Alberta. These carbonates grade to anhydrite and siltstone in eastern Alberta and Saskatchewan.

In the western Alberta Ireton basin and north of the Peace River Arch the upper Woodbend, lower Winterburn unit thickens as it grades to silty dolomites, calcareous siltstones, and shales, as much as 450 feet thick, known as the Tathlina Formation west of Great Slave Lake and northeastern Alberta. It condenses to a carbonate siltstone unit of 0 to 300 feet thick over the Peace River Arch and thickens westward in the southern District of Mackenzie and northern Alberta to comprise the Redknife and Kakisa Formations. The Redknife greenish grey shales contain at the base the Jean Marie Member, a stromatoporoid–coral bioclastic limestone, locally reefy, 50 to 200 feet thick. The uppermost Frasnian Kakisa Formation, a maximum of 200 feet thick, contains coral–stromatoporoid reefs and skeletal calcarenites as well as pelletoid and cryptograined limestones and grades west to greenish grey shales. It was subjected to erosion before deposition of the Famennian Trout River Formation. Equivalents of the silty units of the north are probably to be found in the Calmar siltstone and shale and the Graminia siltstone, dolomite, and anhydrite of central Alberta.

Famennian (Fig. VIII-17). The Famennian part of the Fort Simpson and Besa River shales of southern District of Mackenzie and northeastern British Columbia grades eastward into carbonates of the Wabamun Group. An intermediate sequence of intertonguing carbonates and clastics contains, at the base, the Trout River sandstones, siltstones, silty limestones and shale, as thick as 200 feet, that thin eastward to less than 20 feet and rest disconformably on the underlying Winterburn Group. The gradationally overlying Tetcho Formation is limestone and the succeeding Kotcho Formation comprises dark grey and black shales. The carbonate facies, the Wabamun Group of Alberta, consists, in northern and western Alberta, of 700 feet of brown, fine-grained and pelletoid limestones including bahamites and sparse oölites locally dolomitized. Locally it contains mounds of vuggy dolomite. It extends over the whole of Alberta, the uppermost part overlapping the Precambrian of the Peace River Arch. Dolomitization increases around the Peace River Arch and into south-central Alberta where interbedded dolomite and anhydrite with local halite form the 500-foot-thick Stettler Member. This member thins towards Saskatchewan as it grades to the cyclically alternating variegated shale, dolomite, and anhydrite of the Torquay Formation which is 60 to 130 feet thick. The Torquay changes towards Manitoba to the red sandy facies of the Lyleton Formation. The Big Valley Formation overlies the Stettler and Torquay Formations but is not readily separable from the Wabamun limestone facies to the northwest. It consists of greyish green shale and argillaceous limestone carrying a crinoid–brachiopod fauna. The Wabamun is overlain by the black shale of the Exshaw Formation in Alberta, and in Saskatchewan by the basal black shale of the Bakken Formation of Devonian and/or Mississippian age.

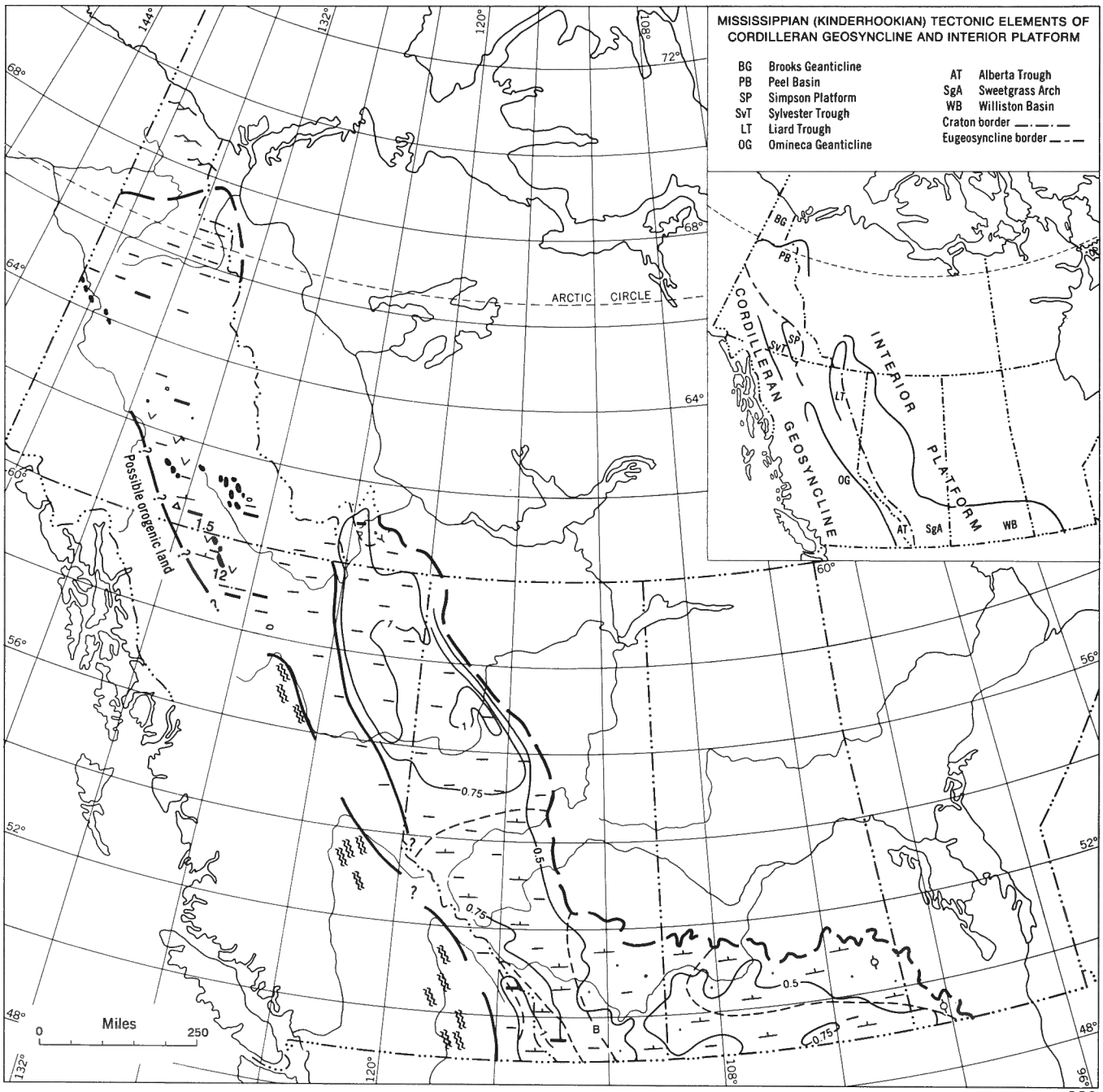
Carboniferous and Permian

Tectonic Summary

Mississippian. The early Mississippian and earlier Caribooan Orogeny (Fig. VIII-22) is represented in the Omineca Geanticline by pre-Osagian metamorphism and deformation of lower Paleozoic and Proterozoic rocks, by the presence of boulders of low-grade metamorphic rocks and felsite in basal conglomerates of the unconformably overlying Osagian and later successions, and by a generally lower degree of metamorphism in the younger rocks. Boulders of granitic and high-grade metamorphic rocks in conglomerate associated with Mississippian rocks in northern Omineca Geanticline indicate that the orogeny may have been accompanied by plutonism. Deposition of black clastic and mafic volcanic rocks in Sylvester Trough probably continued into the early Mississippian, the clastic rocks being derived from uplifted regions to the west, first manifest in the Upper Devonian. In northern Omineca Geanticline, quartzose sediments in an otherwise predominantly volcanic assemblage, probably Mississippian in the Pelly and Omineca Mountains, were derived from the erosion of Lower Cambrian and earlier quartzose sediments and possibly also metamorphic rocks of the eastern part of the geanticline. In Cariboo Mountains the basal clastics were derived from uplifts within the western part of the geanticline. Folding occurred prior to the deposition of Osagian carbonates which are locally associated with clastic and mafic volcanic rocks and lie with strong regional unconformity on strata as old as the Cambro-Ordovician of Kechika Trough to the east. Limestone, fine-grained clastics, and volcanic rocks were deposited in the subsequent site of Stikine Arch east of the northern Coast Geanticline. In British Mountains, Meramecian shales overlap the peneplaned, low-grade, metamorphic rocks of the Ellesmerian Orogen (Fig. VIII-23). Mississippian rocks of the Interior Platform succeed Upper Devonian strata with slight disconformity and, in the more easterly elements of the Cordilleran Geosyncline, may be conformable, particularly where the two systems are represented by very thin black shales that were deposited under starved basin or leptogeosynclinal conditions of deposition. Easterly and northerly derived sands and silts occur low in the successions in Williston, Liard, and Peel Basins and recur at several higher levels, one aspect of the regionally progressive but cyclically interrupted, westward regression of the sea from the craton that occurred during the Carboniferous. A great variety of facies occurs in several types of cyclical alternations. In Liard and Alberta Troughs and adjacent parts of Interior Platform the dominant facies are extremely widespread sheets of crinoidal limestone that grade westward into cherty limestone and shale. Linking of Liard and Alberta Troughs and the establishment of a continuous belt of depression along the margin of the craton east of MacDonald Platform is evident for the first time, the latter element persisting as an arch. In weakly autogeosynclinal Williston Basin the environments of deposition and facies gradually changed from those of euxenic black shale through normal marine shales and limestones with nearshore oölites, into those of a restricted evaporite–red bed basin.

Pennsylvanian. The presence of Pennsylvanian rocks in the southern Cordilleran Geosyncline is established at only a few places but may be represented by greenstone, limestone, and clastic rocks in the upper part of the Mississippian sequences or in the lower part of other sequences established as Permian. In places on northern Coast Geanticline, Pennsylvanian strata are definitely absent, whereas in Fraser Trough, Lower and Upper Pennsylvanian rocks are known and may be fairly abundant to the southeast, in the region between Coast and Omenica Geanticlines. A generally positive behaviour of the geanticlinal belts is indicated during or immediately after Pennsylvanian deposition. Volcanic activity in the southern Coast Geanticline formed islands, probably surrounded by aprons of clastic volcanic material. In northern Yukon the granites of Old Crow batholith have yielded a K–Ar date of 265 m.y. suggestive of an orogenic event perhaps equivalent to the Melvillian Disturbance of the Arctic Islands in latest Pennsylvanian time (Fig. VIII-23). It is not known, however, whether some pre-Permian, northeasterly trending structures of that region are attributable to this disturbance or to the earlier Ellesmerian Orogeny. In northern Liard Trough thick easterly and northerly derived clastics, in part non-marine, reflect uplift of the northern part of the craton and in the south, on the site of the former Peace River Arch, following or during the deposition of red beds and evaporites, warping took place that produced northeast-trending Peace River Embayment. The warping was accompanied by displacement along northwest- and northeast-trending faults which produced horsts and graben, and by peneplanation prior to the deposition of the Middle Permian.

Permian. In the eastern Cordilleran Geosyncline the Permian (Fig. VIII-24) everywhere lies disconformably or unconformably on Carboniferous to Ordovician strata, locally with angular unconformity, as a result of the Ellesmerian and Caribooan Orogenies. In several places in western Cordilleran Geosyncline the Permian lies unconformably or disconformably on Mississippian or Pennsylvanian strata but nowhere are the relationships clearly demonstrable. In general, quiescent conditions prevailed in the southern part of the geosyncline, deposition of carbonate, chert and fine clastic sediments being common. Coarse clastic and pyroclastic rocks are lacking except in the southern Insular and St. Elias Troughs. In the north-central part of the geosyncline the Permian is represented by carbonates and volcanics lying between assemblages of chert, greenstone, volcanic clastic, ultramafic, and fine-grained clastic rocks. Volcanic rocks were extruded mainly as flows and were accompanied by the emplace-



DEPOSITIONAL FEATURES

Conglomerate
Sandstone
Siltstone
Shale
Volcanics
Ultramafics
Greywacke
Chert
Limestone
Oolite

Depositional limit
Truncated limit
Outcrop limit
Facies boundary
Bakken sandstone — B
Yohin sandstone — Y
Isopach (thousands of feet) — 0.5
Thickness (thousands of feet) — 1.5

CARIBOOAN OROGEN

Trend of metamorphic rocks

PHYSIOGRAPHIC FEATURES

Low mountainous land

Blank areas are lowlands and plains developed on crystalline and sedimentary rocks of the craton, and where the Kinderhookian is now absent through erosion, or unknown. In the geosyncline the blank areas may be sea.

FIGURE VIII-22. Early Mississippian (Kinderhookian) sedimentation and tectonism in Western Canada, and some features of the Ellesmerian and Caribooan Orogens.



DEPOSITIONAL FEATURES

Conglomerate
Igneous and metamorphic clasts
Greywacke
Sandstone
Siltstone
Shale
Volcanics
mafic
rhyolitic
Ultramafics
Chert
Limestone
Dolomite
Anhydrite

Depositional limit
Truncated limit
Pekisko PK
Shunda S
Turner Valley TV
Meramecian. M
Chesteran and Pennsylvanian CP
Charles C
Kibbey K
Facies boundary
Outcrop limit
Isopach(thousands of feet)
Thickness (thousands of feet) 0.4

STRUCTURAL FEATURES
(possibly Carboniferous; may be younger and older)

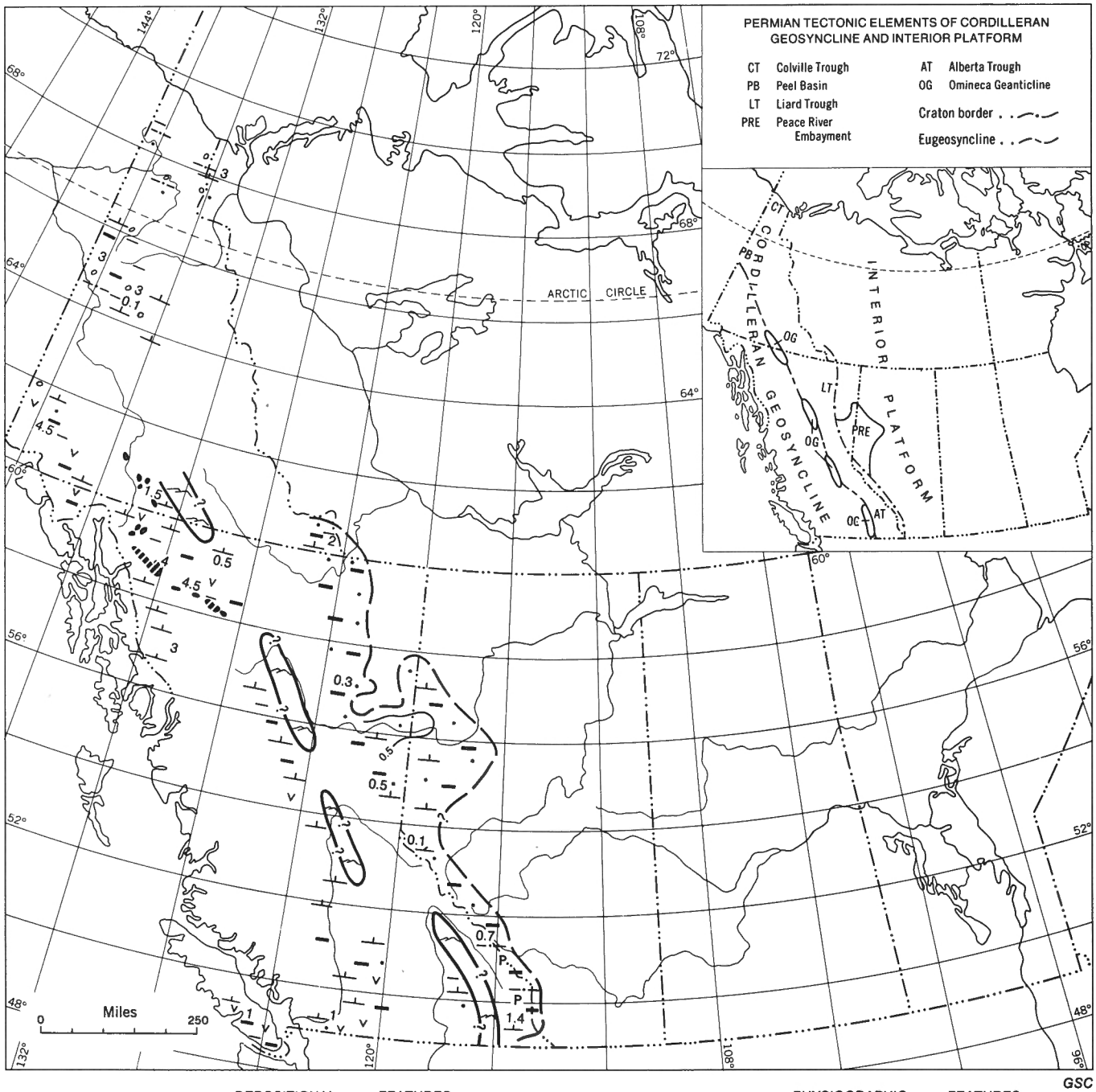
Normal fault
Structural trend
Granitic pluton (age in millions of years) 265

PHYSIOGRAPHIC FEATURES

Mountainous land

Blank areas are lowlands and plains developed on crystalline and sedimentary rocks of the craton, and where the Carboniferous (post-Kinderhookian) is now absent through erosion, or unknown. In the geosyncline the blank areas may be sea.

FIGURE VIII-23. Carboniferous (post-Kinderhookian) sedimentation and tectonism in Western Canada.



DEPOSITIONAL FEATURES

- Conglomerate
- Sandstone
- Siltstone
- Shale
- Volcanics
- Ultramafics
- Chert
- Limestone
- Dolomite
- Phosphate
- Depositional limit
- Truncated limit
- Isopach (thousands of feet)
- Thickness (thousands of feet)

PHYSIOGRAPHIC FEATURES

- Low mountainous land

Blank areas are lowlands and plains developed on crystalline and sedimentary rocks of the craton, and where the Permian is now absent through erosion, or unknown. In the geosyncline the blank areas may be sea.

FIGURE VIII-24. Permian sedimentation and tectonism in Western Canada.

ment of ultramafic rocks near major fracture zones. Volcanoes, built up above wave base, were eroded and local thick sequences of volcanic arenite, mud, carbonate, and chert were formed.

Caribooan Orogeny

In early Mississippian time, the region of Omineca Geanticline was deformed and uplifted. The orogeny is not closely dated except in the Cassiar region where the Osagian and younger Nizi Formation overlies with angular unconformity strata as young as Upper Devonian (Gabrielse, 1963). In Cariboo Mountains the Slide Mountain Group, which overlies the Lower Paleozoic Cariboo Group, is dated as Mississippian, possibly Osagian on the basis of the provisional identification of *Spirifer minnewankensis* Shimer (Sutherland Brown, 1963). Coarse conglomerates of the basal Guyet Formation contain clasts of quartz, quartzite, and chert with subordinate phyllite, argillite, limestone, felsite, and volcanic rocks. The conglomerate was evidently derived from the west, from a terrain comprising rocks of the Cariboo Group previously subjected to low-grade regional metamorphism and intrusion of the Proserpine felsite dykes.

Less certain evidence for the Caribooan Orogeny comes from Aiken Lake map-area (Roots, 1954). A predominantly volcanic unit has yielded poorly preserved fossils thought to be of Mississippian age, but some may be Permian. The upper part contains conglomerate and sandstone derived from a siliceous metamorphic terrain that probably lay to the east. Roots has also included in the unit a bed of coarse conglomerate with clasts of massive and foliated granite he regarded as identical with rocks of the Wolverine Complex. As this complex contains Lower Cambrian rocks it must have been metamorphosed later, most likely during the Caribooan Orogeny. However, the conglomerate containing the granitic clasts lies along a fault, and according to Souther and Armstrong (1966) it may represent a sliver of Mesozoic rocks and be the product of a later orogeny.

Further evidence for the Caribooan Orogeny comes from southern Selkirk Mountains where the Upper Mississippian and younger Milford Group contains a basal conglomerate with clasts of quartz, chert, argillite, and phyllitic grit. One lens contains clasts as large as 12 inches of phyllitic grit and greenstone of the Broadview Formation (Wheeler, 1966b). The foliation within the clasts is randomly oriented in contrast to the regional foliation which is restricted to the matrix of the conglomerate. Thus, the Broadview Formation must have been deformed, metamorphosed to low greenschist facies, and, as previously mentioned, possibly thrust over Lower Paleozoic rocks prior to the deposition of Milford sediments.

Granitic plutons may have been emplaced locally into the Omineca Geanticline as post-tectonic plutons following the Caribooan Orogeny. The Adamant batholith in northern Selkirk Mountains and the Toby stock in Purcell Mountains, both hypersthene monzonite plutons, have yielded K–Ar dates only as old as 281 m.y. and 232 m.y., respectively, but these plutons may have been modified during periods of Mesozoic regional metamorphism.

It is probable that in early Mississippian time orogeny and some uplift took place in the region of the Omineca Geanticline. Uplift apparently was greatest in the north as a large amount of clastic rock occurs in the Upper Devonian to Lower Mississippian sequences. Uplift was probably less in the south, as there are no indications of coarse clastics derived from the west having been deposited in adjacent Liard and Alberta Troughs.

Mississippian of Western Cordilleran Geosyncline

Along the Omineca Geanticline where evidence of Caribooan Orogeny is best, formations containing sparse Mississippian fossils overlie older units. These are: the Milford Group near Kootenay Lake, the Slide Mountain Group in Cariboo Mountains, and an unnamed group in Omineca Mountains. All are characterized by clastic sediments, commonly carrying recognizable detritus of older formations.

In *southern Selkirk Mountains,* the Milford Group is either unconformable on or faulted against the Broadview Formation of unknown age. The lower part of the Milford Group contains late Meramecian and early Chesteran corals whereas the upper part contains poorly preserved Pennsylvanian or Permian fossils.

The Slide Mountain Group in *Cariboo Mountains* is relatively unmetamorphosed and overlies the characteristically low-grade metamorphic rocks of the Cariboo Group with great unconformity. The basal Guyet Formation is not in contact with the Snowshoe Formation, the youngest of the Cariboo Group, but successively overlies other formations from the Cunningham to Midas. The conglomerate contains clasts of quartz, quartzite, and chert and, subordinately, of phyllite, argillite, and limestone derived from the Cariboo Group and of felsite from the Proserpine dykes. The conglomerate was apparently derived from a nearby westerly source (Sutherland Brown, 1963) where the Cariboo Group had been deformed during the Caribooan Orogeny and subsequently intruded by the Proserpine dykes. The Greenberry limestone at the top of the Guyet Formation yielded poorly preserved fossils of probable Osagian age. It is overlain by pillow basalts, chert, and argillite of the Antler Formation into which are intruded Mount Murray diabase sills. The Slide Mountain Group is deformed into broad north-trending folds in contrast to the tightly folded and cleaved Cariboo rocks beneath.

In *Omineca Mountains* possible equivalents of the Slide Mountain Group include pillowed greenstone, ribbon chert, slate, and limestone yielding Permian or Carboniferous fossils mapped with the Cache Creek Group southwest of Wolverine Complex (Armstrong, 1949) and volcanic and sedimentary rocks that are separated from older and younger formations by faults (Roots, 1954).

The latter unit has yielded poorly preserved fossils ranging in age from Mississippian to Permian. It is composed principally of green bedded tuff and andesite and basalt flows with subordinate limestone, argillite, greywacke, chert, sandstone, and conglomerate. Although occurring within a predominantly volcanic succession, the sandstone is unusual as it contains no volcanic material but is composed mainly of quartz, quartzite, and mica. This detritus is the product of erosion of quartzose metamorphic rocks, presumably the Lower Cambrian and older formations lying to the northeast. Conglomerate is rare and most beds contain clasts of chert, quartz, quartzite, sandstone, andesite, tuff, jasper, and cherty argillite set in a limonitic gritty matrix. One bed included in the sequence by Roots occurs along a red-weathering fault zone. The boulders, as much as 14 inches across, are foliated granitic rocks like those in the Wolverine Complex enclosed by a hematitic gritty matrix. Roots considered the conglomerate to be part of the late Paleozoic sequence and consequently inferred that an episode of plutonism and deformation took place which produced the Wolverine Complex subsequent to the Lower Cambrian and prior to the late Paleozoic. The conglomerate resembles that of the fault-bounded Aptian Uslika Formation and if it is indeed part of the Uslika then the plutonic episode represented by the clasts may be dated only as pre-late Lower Cretaceous.

In eastern *Cassiar Mountains* Osagian and younger strata comprising about 1,200 feet of cherty limestone with basal siltstone, sandstone, and shale, unconformably overlie strata as old as Cambrian (Gabrielse, 1963). Farther west, the Chesteran Nizi Formation, comprising 1,600 feet of limestone with basal greywacke and conglomerate, overlies the Sylvester Group unconformably. In northern Cassiar Mountains an assemblage of andesitic and basaltic volcanic rocks, chert, slate, quartzite, argillite, limestone, and conglomerate probably correlative with the Sylvester Group is more than 8,000 feet thick and probably includes Mississippian rocks. The overlying sedimentary assemblage of chert, slate, quartzite, argillite, limestone, and conglomerate, more than 10,000 feet thick, probably includes rocks younger than Mississippian. East of Teslin Lake the Englishmans Group contains a lower unit of limestone, as much as 1,500 feet thick, and an upper unit dominantly of slate, quartzite, chert, and conglomerate (Mulligan, 1955). Mississippian rocks may also be included in the underlying or partly correlative Big Salmon Complex, an assemblage of schist, gneiss, greenstone, amphibolite, and limestone that probably also includes equivalents of the Sylvester Group to the east.

On *MacMillan Plateau,* chert-pebble conglomerate of the Crystal Peak Formation is at least 4,000 feet thick and may be entirely Osagian as the unit appears to be underlain and overlain by fossiliferous limestone of that age (Campbell, 1967). Shattered syenite stocks in Pelly Mountains and numerous ultramafic bodies in Cassiar and Pelly Mountains are considered to be of Mississippian age. In Simpson Range northeast of Tintina Trench are thick eugeosynclinal rocks of Mississippian or later age. Similar rocks are believed to form part of the metamorphic terrain of the Yukon Group southwest of Tintina Trench.

Between Stikine and Iskut Rivers in *Stikine Plateau* a limestone, 250 to 300 feet thick with Mississippian fossils, is separated from another limestone unit more than 1,800 feet thick by 2,000 feet of volcaniclastic rocks (J. K. Rigby). The strata structurally overlie phyllitic and volcanic rocks of unknown age and are either overlapped by Triassic rocks or faulted against them.

No Mississippian rocks have been positively identified in *St. Elias Mountains* but may be present in a complex of volcanic rocks, chlorite schist, greywacke, slate, phyllite, conglomerate, and a limestone that contains Upper Devonian fossils (Wheeler, 1963; Muller, 1967).

Pennsylvanian and Permian of Western Cordilleran Geosyncline

The Cache Creek Group of Pennsylvanian and Permian age is exposed in *eastern Interior Plateau* along a northwesterly trending belt, the Pinchi Geanticline, and in a belt *en echelon* to the southeast that extends southeastward into the Shuswap Metamorphic Complex. The group consists of a variable assemblage of limestone, ribbon chert, slate, and volcanic rocks and relatively minor arenite and conglomerate. The volcanic rocks are mainly massive greenstone, originally andesitic and basaltic lavas, pyroclastics, and volcanic arenite, within which ultramafic rocks occur locally. The stratigraphy of the group is not yet understood because of the intense deformation of the rocks, the poor exposure, and the lack of distinctive marker beds which can be recognized from place to place. Fossils occur most commonly in the limestones, the fusulinids being the most diagnostic types and representative of the Early and Late Pennsylvanian and all stages of the Permian (Danner, 1964). Shelly fauna range in age from Late Pennsylvanian to Early Permian although some northeast of Kamloops have been referred to the late Mississippian (Cockfield, 1948). Estimates have been made that as much as 20,000 feet of strata are present but wherever detailed work has been done repetition has been recognized and thicknesses vastly reduced. The Lower Permian and earlier Asitka Group at the northernmost end of the Pinchi Geanticline is unusual in that it contains abundant rhyolite that is locally spherulitic (Lord, 1948), although it contains andesitic lavas and associated pyroclastics also. The Cache Creek Group is overlain with erosional unconformity by the Upper Triassic or Lower Jurassic on the flanks of the Pinchi Geanticline (Armstrong, 1949; Tipper, 1959), separated along the east side by the Pinchi and related faults.

Near Vernon in the *Monashee Mountains,* unfossiliferous rocks correlated with the Cache Creek Group appear to lie with angular unconformity on the Monashee Group of the Shuswap Metamorphic Complex (Jones, 1959), but the contacts may be faults according to V. A. Preto. Other unfossiliferous rocks west of Vernon similar

to the Cache Creek Group lie with angular unconformity on the Chapperon Group and the Old Dave ultramafic dykes that intrude it. In the Vernon area the rocks beneath the unconformities, and also the Mount Ida Group, are isoclinally folded whereas the rocks above are not. It is uncertain, however, when this deformation took place because of the uncertainty of the ages of the beds above and below the unconformities. Jones considered the rocks beneath to be Precambrian but recent work (Hyndman, 1968) suggests that the Monashee Group contains rocks as young as late Paleozoic and perhaps early Mesozoic as well. Furthermore, fossils presumed to be of Carboniferous or Permian age occur in a northwest extension of the Mount Ida Group (Campbell, 1963). The Chapperon Group contains rocks similar to the fossiliferous strata and on this tenuous basis may be considered partly Permian or Carboniferous. Accordingly, if the overlying strata are Cache Creek Group, which are late Pennsylvanian to early Permian nearby, then the Chapperon Group must be Carboniferous and the deformation and unconformity intra-Carboniferous. Furthermore if the Mount Ida and Chapperon Groups are Permian then the unconformably overlying strata may be Lower or Middle Triassic, and the deformation would be earliest Mesozoic, possibly the Tahltanian Orogeny which is also evident near Greenwood.

On *Stikine Plateau* the upper part of the Cache Creek Group includes a lower limestone member, locally about 300 feet thick of Middle Permian age (Monger, 1968). The succeeding middle member comprises jasper-bearing, maroon to green, basic to intermediate volcanic flows, locally pillowed, and foliated lithic tuff and agglomerate, with a maximum apparent thickness of about 2,000 feet. The upper member of Upper Permian limestone is possibly as much as 2,000 feet thick. This succession overlies a thick assemblage of silica-rich phyllites and slates, thin-bedded cherts, and minor limestone, rhyolite, and volcanic greywacke of the lower part of the group that may be Lower Permian and Pennsylvanian in age. A single mass of limestone north of Atlin Lake contains fossils ranging in age from Middle Pennsylvanian to Upper Permian. In the Coast Mountains near Stikine River limestone sequences, as much as 2,000 feet thick, represent all stages of the Permian except the Ochoan. Drift covers the contact between Permian and Mississippian beds and although the beds are structurally discordant, their stratigraphic relations are not known.

About 150 feet of sandy tuffaceous limestone in northern *Cassiar Mountains* may be correlative with the upper limestone member of the Cache Creek group. The limestones overlie the Sylvester Group with marked discordance but whether the contact is a fault or unconformity is not known. In northwestern *Cassiar Mountains* about 1,500 feet of early Pennsylvanian limestone is overlain by about 5,000 feet of chert, argillite, and limestone, with a basal conglomerate of well-rounded cobbles and boulders of chert and quartzite. On *MacMillan Plateau* and to the southeast, the Pennsylvanian may be represented by a belt of volcanic rocks, limestone, chert, and fine-grained clastic rocks that extend along the northeast side of Tintina Trench (Tempelman-Kluit, 1968).

Permian rocks in *St. Elias Mountains* are as much as 3,000 feet thick and include a lower unit of volcanic breccia and conglomerate and an upper unit of greywacke, sandstone, argillite, and limestone (Muller, 1967). They apparently unconformably overlie Devono-Mississippian metamorphic rocks and are overlain by Middle Triassic strata.

Limestone containing Lower Permian fusulinids and shelly fauna is associated with argillite and greenstone in the eastern *Coast Mountains* near Skeena River (Duffell and Souther, 1964). Permian limestone also occurs farther southeast, west of Whitesail Lake (P. B. Read) and much of the metasediments and metavolcanics in roof pendants within the Coast Plutonic Complex may be of the same age. Generally unfossiliferous groups such as the Fergusson in southeastern Coast Mountains and the Hozameen consist of assemblages of greenstone, ribbon chert, argillite, and minor limestone that strongly resemble parts of the Cache Creek Group.

The Chilliwack Group, in western *Cascade Mountains,* contains much volcanic arenite and tuff in contrast with effusive greenstones and non-clastic sediments characteristic of the Cache Creek Group. Furthermore, the Chilliwack Group includes a conglomerate with volcanic clasts that lies not far above Lower Pennsylvanian (Morrowan) limestone. The conglomerate is associated with beds containing fossil plants of Pennsylvanian age. Overlying volcanic arenite and tuff are succeeded, apparently conformably, by andesitic lavas. Equivalent limestone beds and lenses contain Lower Permian (Leonardian) fusulinids (according to J. W. H. Monger).

On *Vancouver Island,* near Buttle Lake, the Sicker Group consists of altered basalt flows and pyroclastics underlying Lower Permian limestone (Yole, 1963). In the southeast, however, greywacke and chert more commonly underlie the limestone. Although younger Permian strata are not recognized on Vancouver Island, late Permian ribbon chert, argillite, greywacke, and volcanic rocks containing lenses of limestone are present on nearby San Juan Island (Danner, 1964).

Melvillian Disturbance

During the late Pennsylvanian and early Permian, tectonic disturbances, possibly orogenic, took place (Fig. VIII-23). In northern Yukon the structures have not been differentiated from earlier ones. Evidence of plutonic activity are the K–Ar dates of 225 and 265 m.y. from the granites of Old Crow Range. General instability is indicated by the widespread pre-Permian unconformity and by the presence of much chert-pebble conglomerate in the Permian and some in the Pennsylvanian. Post-Mississippian displacement occurred on the southwest extension of the McDonald fault system of the Precambrian basement.

As a result of late Pennsylvanian to early Permian movements, probably normal, on the faults bordering Peace River Arch, this tectonic element ceased to exist as a positive feature and subsequently the region was generally relatively more negative than adjacent regions (Figs. VIII-23, -25). It is possible that some structures in northern Yukon Territory also reflect normal or transcurrent fault movements. The anomalous north trends of Barn Mountains, relative to the southeast trends of British Mountains and the northeast trends of Keele Range may indicate the presence of a northeast-trending dextral transcurrent fault system passing south of Barn Mountains. North-trending pre-Albian, post-Middle Devonian faults in Colville Hills west of Great Bear Lake may have had displacement at this time.

In the White River region of southern Rocky Mountains, diorite sills and dykes are intrusive into the Cambro-Ordovician McKay Group (Leech, 1964). Associated diatreme-like breccias with fragments of diorite and the host and other formations cut strata as young as the Rocky Mountain Group.

Carboniferous and Permian of Eastern Cordilleran Geosyncline and Interior Platform

In *British Mountains* the Mississippian lies with regional angular unconformity on low-grade metamorphic rocks of the Neruokpuk Formation, the deformation of which has been treated as part of the Late Devonian Ellesmerian Orogeny. According to E. W. Bamber, the basal 600 feet of calcareous black shale and limestone carry Meramecian fossils near the top, and is provisionally termed the Kyak Formation. In the Barn Mountains it is represented by 300 feet of non-marine sandstone and shale with basal conglomerate. The Kyak is conformably overlain by 2,000 feet of finely crystalline limestone and coarsely crystalline dolomite of the Lisburn Formation which is Chesteran and Pennsylvanian in age. Disconformably overlying it is at least 600 feet of light grey shale, siltstone, and sandstone constituting the Permian Echooka Member of the Saddlerochit Formation, the upper member of which is Triassic. In southern *Porcupine Plateau* the Mississippian probably conformably succeeds the Upper Devonian, the boundary between the systems lying within thin black shales. The oldest known Mississippian is thin Meramecian limestone on upper Peel River, and shales and siltstones of possible Kinderhookian age east of Richardson Mountains, conformably overlying the Imperial Formation. The Pennsylvanian is incompletely represented by a sequence of massive-bedded bioclastic limestone alternating with black silty shale that grades eastward and southward into shale, siltstone, and chert-pebble conglomerate. The Permian Tahkandit Formation contains grey shales, grits, fossiliferous limestones, and grey chert as much as 1,000 feet thick. In northern Yukon, the Permian generally lies disconformably or unconformably on older beds. In *Keele Range*, where crossed by Porcupine River, the conglomerates lie with angular unconformity on the Silurian and Devonian Road River shales, and on the Devonian Imperial Formation in northern *Richardson Mountains*. Along the Alaska Boundary near Yukon River in *Ogilvie Mountains* all or part of the Mississippian may be represented in a post-Middle Devonian to pre-Permian sequence of black shale, siltstone, and argillaceous limestone generally about 2,000 feet thick.

No Mississippian rocks have been identified in Selwyn and Mackenzie Mountains although they may be present locally in the fine clastic sequences generally referred to the Upper Devonian. In *Liard Plateau* and southernmost *Franklin Mountains*, nearly 6,000 feet of Carboniferous and Permian strata conformably succeed Upper Devonian black shale in the axial part of Liard Trough. Brown, fine-grained sandstones of the Yohin Formation, 600 feet thick, are considered Mississippian, and are conformably succeeded by 550 feet of non-calcareous, black shale of the Kinderhookian and Osagian Clausen Formation (Harker, 1963). Thin-bedded, dark, argillaceous calcarenite and shale of the Flett Formation, nearly 2,000 feet thick, contain intercalations of sandstone in northern Liard Range and grade westward, together with the Yohin, into dark grey, non-calcareous shale and siltstone of the Etanda Formation. The latter shales are 2,300 feet thick and are inseparable from the Besa River Formation. The Mattson Formation is a thick, northeasterly derived sequence of sandstone, siltstone, and shale, mainly Lower Pennsylvanian with late Mississippian coal seams in the lower part. It is overlain with regional unconformity by sandstone and bedded grey chert of the Permian Fantasque Formation. Most of the Mattson probably grades southwestward to shale of the Besa River Formation and the Fantasque to siltstone and shale of the upper Kindle Formation (E. W. Bamber). South of Liard River, and in the subsurface to the east, massive dark cherty limestones of the Prophet Formation representing the Osagian and Meramecian grade westward into the Besa River shale. The vestigial remnant of the Prophet is in many places a persistent thin bed of black chert and limestone. The Prophet grades eastward into the Shunda, Pekisko, and Debolt Formations, about 1,000 feet of interbedded carbonates, shales, and minor anhydrite. Conformably succeeding the Prophet and Debolt Formations is the Stoddart Group, best known in the subsurface of the Peace River Basin. The basal Golata Formation is composed of dark grey to greenish grey shale transitionally overlain by the Kiskatinaw Formation of grey and black shale, siltstone, light grey sandstone, and sandy cherty limestones and dolomites. The Kiskatinaw grades eastward into more sandy beds and upwards into a predominantly carbonate sequence, possibly partly Pennsylvanian in age, forming the Taylor Flat Formation. It consists of dark grey and brown, cherty, silty and sandy dolomites, siltstones, and sandstones. Uplift and movements along faults trending northeast and northwest, in part possibly contemporaneous with deposition of the Stoddart (Fig. VIII-25), were followed by extensive erosion prior to the deposition of the Lower

Permian Belloy Formation. The latter consists of thin glauconitic sandstone, white, silty limestone, and chert with minor conglomerate.

Throughout most of Alberta Trough in the southern Rocky Mountains and beneath the southern Interior Platform the Mississippian has been truncated by pre-Mesozoic erosion, and its present distribution is only the southwesterly remnant of its original extent. The boundary between the Devonian and Mississippian systems is generally considered to be marked by the Exshaw Formation, although its assignment to either system is still unsettled. The Exshaw is a thin, black, bituminous shale, spore-bearing in the subsurface, and with cherty layers in outcrop. It is missing locally in the mountains near Athabasca River. The Exshaw seems to be continuous with the basal black shale of the Bakken Formation of Williston Basin which is overlain by the Coleville sandstone, a light grey to brown, very fine grained, calcareous sandstone about 60 feet thick (Fuller, 1956). The upper Bakken is black shale, about 25 feet thick, fissile, and slightly calcareous. The base of the Kinderhookian Banff Formation is commonly marked by a thin dolomitic siltstone. Succeeding beds are alternating dark grey fissile shales and bands of argillaceous limestone, locally cherty, that grade upwards into dark grey, calcareous or dolomitic shale and limestone in Bow Valley and to the north, or into dark grey, massive, finely crystalline, cherty limestone and dolomite in the south. Near the transitional upper contact with the Rundle Group is a thin persistent siltstone which may be the eastern part of some 400 feet of sandstone, siltstone, and silty limestone near Connor Lake and which apparently has a western source.

On Sweetgrass Arch of southern Alberta, the Banff Formation, 500 feet thick, grades into green shales and shaly limestone, and thence to cream, buff, or pink, cryptograined, cherty limestone, with echinoid and bryozoa debris, interbedded with thin silty argillaceous layers and at the top of variably persistent, red and green, glauconitic shale and siltstone (Penner, 1959). The essentially coeval Souris Valley Formation of the Williston Basin in eastern Saskatchewan and Manitoba consists of a lower shaly unit 20 feet thick, lying with sharp contact on the Bakken and overlain by 400 feet of grey and dark grey, argillaceous, fine-grained, cherty limestone, and an upper argillaceous limestone about 100 feet thick. Along the eastern margin of the basin the middle carbonate thins and becomes lighter and more granular, and interfingers with a complex of calcareous shales and thin beds of coarse oölitic and crinoidal limestone of considerable purity (Stanton, 1958).

The Rundle Group in southern Alberta Trough is Osagian to Chesteran in age. It is dominated by extensive sheets of crinoidal debris, intercalated with beds of dark grey argillaceous limestone and cherty limestone, presumably tongues of a more westerly facies, or with anhydrite and dolomite with shale and sandstone derived from the east. The Rundle Group embraces three formations, the Livingstone, Mount Head, and Etherington (Douglas, 1959). The thick, massive-bedded, crinoidal limestones and dolomites of the Livingstone transitionally overlie the Banff Formation in southern Alberta Trough and thin eastward on southern Sweetgrass Arch. The formation thins and changes facies northward and eastward, with rock units of member status (the Pekisko, Shunda, and

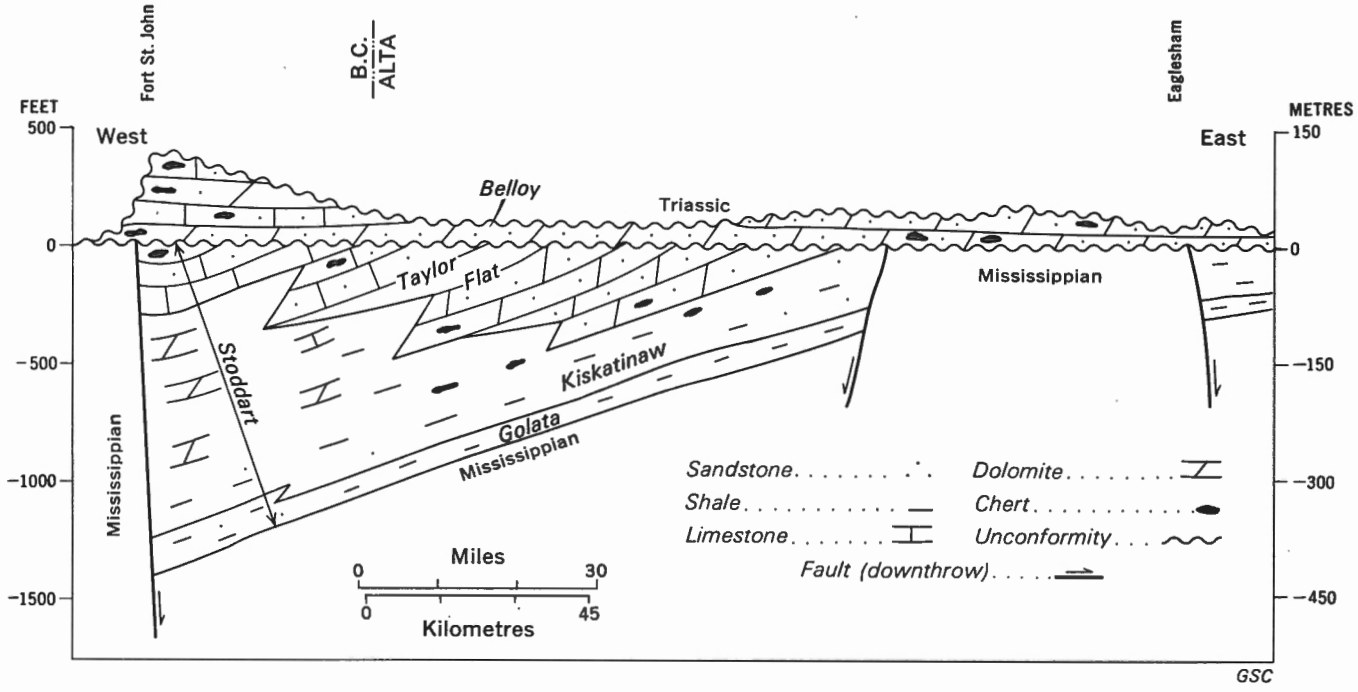


FIGURE VIII-25. Cross-section of Carboniferous and Permian rocks of Peace River Embayment, British Columbia and Alberta, showing pre-Permian (Belloy Formation) normal faults (by H. R. Belyea).

Turner Valley) attaining formation status. The lowest unit, the Pekisko Formation, consists of medium to coarsely crystalline crinoidal limestone in beds as much as 50 feet thick interbedded with finely crystalline, dark grey to brownish grey, argillaceous dolomitic limestone with dark green chert nodules. It grades eastward to about 200 feet of light yellowish brown, coarse-grained, bioclastic chalky limestone, with dark shale and siltstone, and northward into dark grey shales. The overlying Shunda Formation is a very finely crystalline limestone and limestone breccia that grades northward into shale. To the east it consists of argillaceous silty dolomite, breccia, and anhydrite. The Turner Valley Formation has a lower or Elkton Member and an upper member formed of coarsely crystalline crinoidal limestone, locally altered to porous, coarsely crystalline dolomite. They are separated by a middle member of very fine grained silty, laminated, cherty dolomite. The Meramecian Mount Head Formation is divisible into six members, of which the Wileman, Salter, and Marston Members are mainly sandstone and silty dolomite, and grade westward into argillaceous and crinoidal limestone or eastward near their present truncated margins into anhydrite, evaporitic dolomite, and red sandstone. Intervening members, the Baril and Loomis, are mainly crinoidal limestone that grades eastward into porous dolomite. The uppermost member, the Carnarvon, is a persistent, dark argillaceous cryptocrystalline limestone, the lower part of which grades westward into shale and limestone inseparable from the Marston. The Mount Head and Turner Valley Formations pass northward into the Debolt Formation of southern Liard Trough and adjacent Peace River Embayment (Halbertsma, 1959). This unit, about 800 feet thick, consists of light yellowish brown, cherty, massive, bioclastic limestone and dolomite interbedded with shale in the lower part, and an upper part of light brown, very finely crystalline to dense, locally cherty dolomite with pink and white anhydrite. The Chesteran Etherington Formation consists of thin, cyclically alternating beds of green and red shale, and dense, silty, cherty, finely crystalline limestones and dolomites that grade westerly into thick, coarsely crystalline, bioclastic and dark argillaceous limestone.

East of Sweetgrass Arch on the northern flank of Williston Basin in Saskatchewan and Manitoba, the Mississippian sequence (Fig. VIII-26) has been truncated by pre-Mesozoic erosion so that progressively younger members appear downdip to the south towards the centre of the basin in North Dakota. The Osagian and Meramecian (Thomas, 1954) consist predominantly of carbonates in the west that grade into cyclical alternations of carbonates and evaporites, which, farther east, become progressively more evaporitic and contain more sandstone and siltstone lenses (Fig. VIII-26). They comprise the Mission Canyon and overlying Charles Formations of which a maximum of 1,000 feet has been preserved in the south. All Mission Canyon cycles resemble one another although varying somewhat in detail. The lowest cycle, the Tilston Beds, is typical. Cherty, crinoidal limestones at the base grade up to fine-grained, chalky limestones and finely crystalline, porous dolomites with oölite, pisolite, and algal zones prominent in the upper part (Member MC1 of Thomas). These are overlain by argillaceous silty dolomitic limestones with streaks of shale that grade eastward to anhydrite, dolomitic anhydrite, and mottled, red and grey, cryptocrystalline algal dolomite (Member MC2). From east to west and northeast to southwest the evaporites grade to argillaceous dolomites, and the oölitic, algal carbonates grade to fine-grained cherty limestones. These facies changes rise to successively higher levels westward, each succeeding cycle developing into a complete carbonate sequence near the central part of Williston Basin. The carbonate–evaporite cycles are interpreted by Thomas and Glaister (1960) as the deposits of a shallow-water shelf with marginal evaporites, from which the sea was retreating gradually but with a rhythmic inflow of marine waters. The Charles Formation as defined by Thomas and Glaister is restricted by erosion to southern Saskatchewan. It also comprises cycles of evaporites and dolomite. The evaporites are anhydrite, anhydritic dolomite, and finely crystalline dolomite. Salt occurs in the uppermost and lowermost units in the central part of the basin. The carbonates are similar to those of the Mission Canyon Formation.

The Chesteran Big Snowy Group, occurs mainly in the United States and as an erosion remnant only in southernmost Saskatchewan. The Kibbey Formation consists of sandstone with anhydrite and carbonate near the top, and the Otter Formation consists of variegated red and green shales.

In the southern Rocky Mountains, disconformably overlying the Mississippian carbonates of the Rundle Group and disconformably overlain by the Triassic Spray River Formation, are sandstone and dolomite of the Rocky Mountain Group of Early and Middle Pennsylvanian age and phosphatic beds and chert of the Ishbel Group, late Early and Late Permian in age (McGugan and Rapson, 1962). The succession is thickest and most continuous in southern Alberta Trough, the formations thinning rapidly eastward and northward through convergence and truncation at several disconformities. The sandstones of the Tyrwhitt, Storelk, Tobermory, and Tunnel Mountain Formations are massive or thinly bedded, dominantly fine- and even-grained, crossbedded and quartzose, and are probably easterly derived. They are overlain by the Kananaskis Formation, a maximum of 180 feet of silty and cherty dolomite. The disconformably overlying Johnson Canyon Formation of the Ishbel Group consists of thinly bedded, phosphatic quartzose siltstone conformably overlain by the sandy limestones of the Telford. The Ross Creek contains alternating thin and thick beds of silty cherty carbonate, siltstone, and chert. The uppermost unit, the Ranger Canyon chert, is

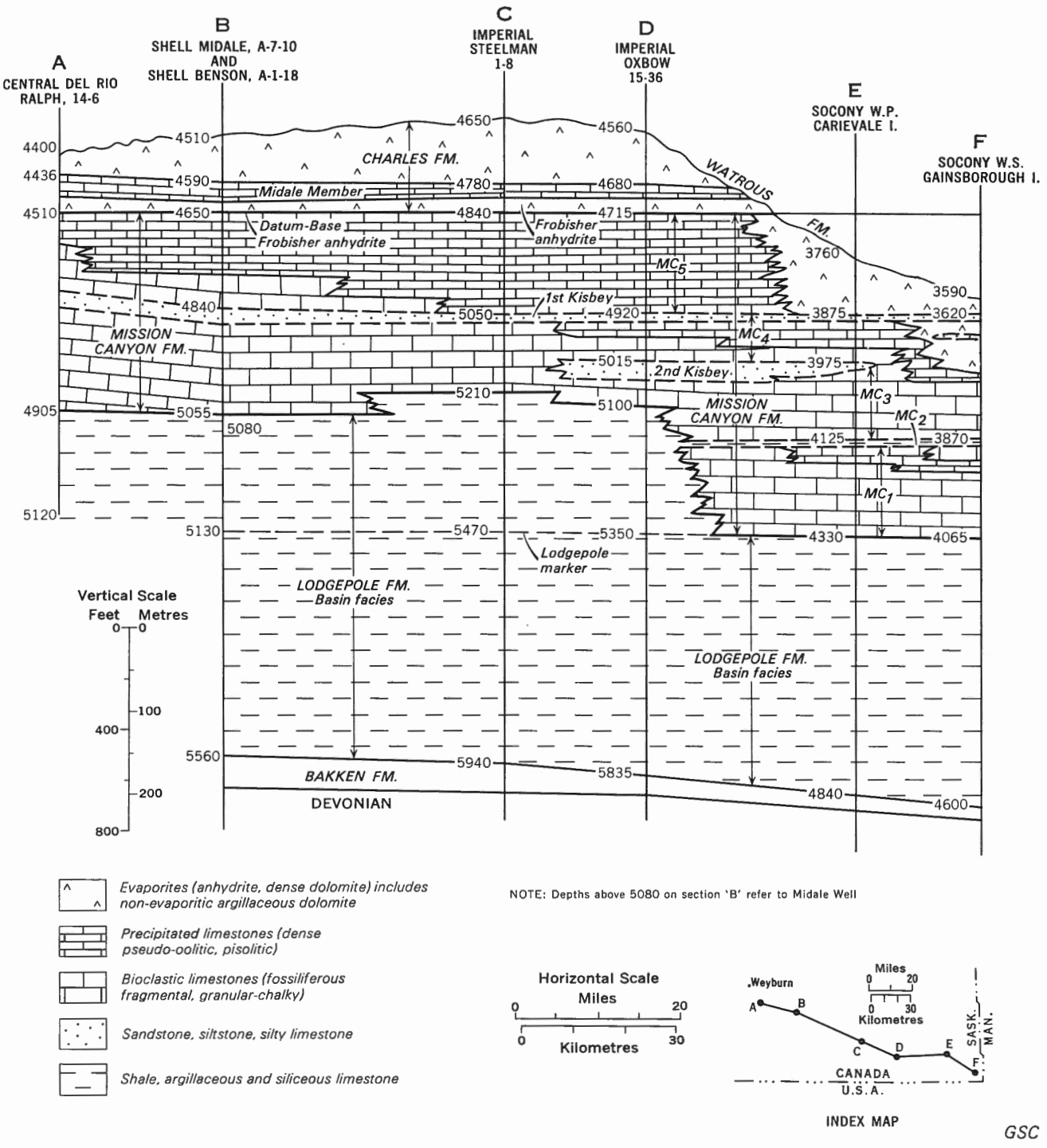
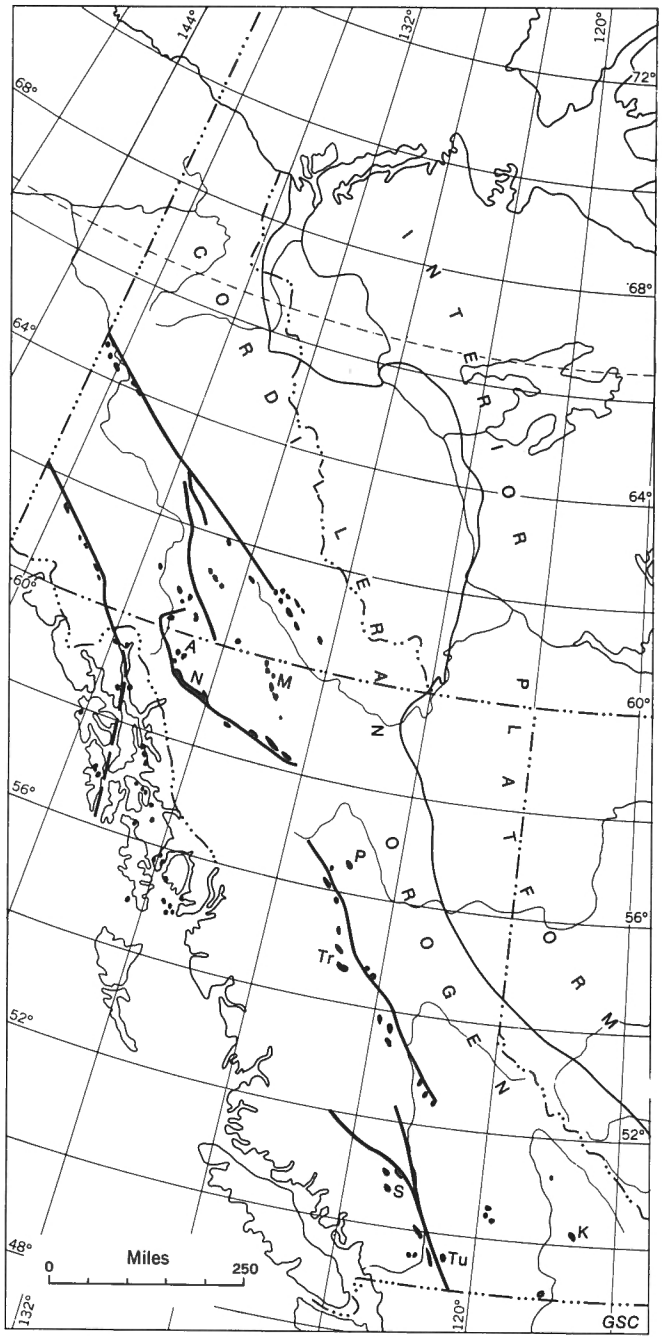


FIGURE VIII-26. Stratigraphic section through Mississippian rocks of Williston Basin in Saskatchewan (Edie, 1958).

a massive, persistent but thin bed that unconformably overlaps eastward, resting on the Mount Head Formation in the eastern Foothills. Locally, near Jasper, the uppermost Permian is represented by the Mowitch Formation, a coarse-grained sandstone that is glauconitic, phosphatic, and fish-bearing.



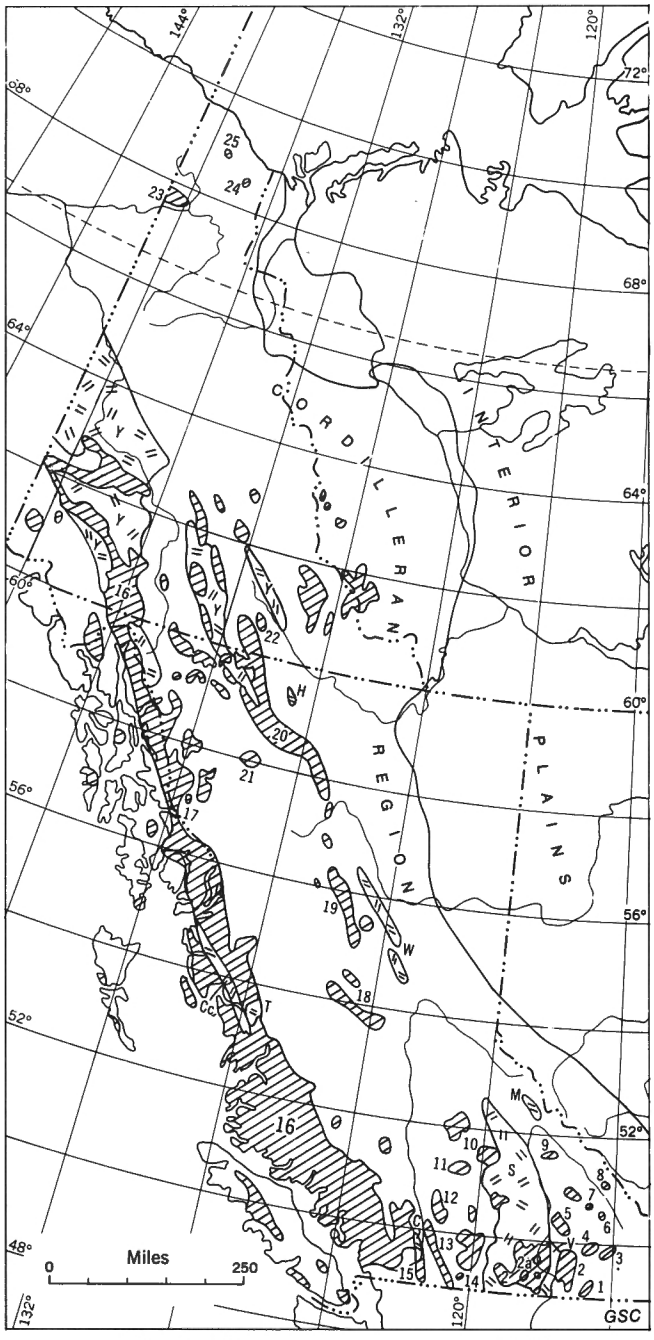
ULTRAMAFIC INTRUSIONS

Atlin . . . . A	*Polaris* . . . . P
Kaslo . . . . . K	*Shulaps* . . . S
McDame . . . M	*Trembleur* . . Tr
Nahlin . . . . N	*Tulameen* . . Tu

Fault . . . . . ———

Fault (underwater) . . . . — —

FIGURE VIII-27. Distribution of ultramafic intrusions in the Cordilleran Orogen, British Columbia, Yukon Territory, and southeastern Alaska.



GRANITIC PLUTONS

1 Bayonne Batholith	13 Mt. Lytton Batholith and Eagle granodiorite
2 Nelson Batholith	14 Copper Mountain Stock
2a Coryell Intrusions	15 Chilliwack Batholith
3 White Creek Batholith	16 Coast Plutonic Complex
4 Fry Creek Batholith	17 Galore Creek Pluton
5 Kuskanax Batholith	18 Topley Intrusions
6 Horsethief Batholith	19 Hogem Batholith
7 Bugaboo Batholith	20 Cassiar Batholith
8 Ice River Complex	21 Hotailuh Batholith
9 Adamant Batholith	22 Marker Lake Batholith
10 Takomkane Batholith	23 Old Crow Batholith
11 Thuya Batholith	24 Mt. Fitton Stock
12 Guichon Batholith	25 Mt. Sedgwick Stock

METAMORPHIC COMPLEXES

C Custer Gneiss	
Cc Coast Complex	
H Horseranch Complex	
M Malton Gneiss	
S Shuswap Complex	
T Tahtsa Complex	
V Valhalla Complex	
W Wolverine Complex	
Y Yukon Complex	

Granitic pluton . . . ///

Metamorphic complex . . . = ||

FIGURE VIII-28. Distribution of granitic plutons and metamorphic complexes of the Cordilleran Orogen.

ULTRAMAFIC ROCKS

Most of the known Cordilleran ultramafic intrusions (Fig. VIII-27) are alpine-type peridotites originally composed mainly of olivine and orthopyroxene (harzburgite). They characteristically occur as elongate bodies, commonly faulted, sheared, and highly serpentinized. The ultramafic rocks were emplaced at various times from the late Devonian to the early Jurassic, but mostly during the late Devonian to early Mississippian interval and the Pennsylvanian to late Middle Triassic interval. Several of the bodies, now largely serpentinite, have been re-intruded and squeezed tectonically into younger rocks during one or more of the orogenies that occurred after their initial emplacement as peridotites.

Some of the ultramafic intrusions are restricted largely to belts of thick greenstone, originally basalt and pyroxene andesite, and appear to be coeval with them. These are the McDame Intrusions within the Sylvester Group of Late Devonian and Mississippian age (Gabrielse, 1963), the Atlin Intrusions within the Pennsylvanian and Permian Cache Creek of Atlin Horst (Aitken, 1959), unnamed intrusions in the Triassic Kaslo Group (Cairnes, 1934), and perhaps also those in the lavas of the Lower Jurassic part of the Rossland Group (Little, 1960). Ultramafic rocks are not common in the pyroclastic-rich, younger Mesozoic volcanic sequences. The intimate association between the greenstone and ultramafic rocks suggests a community of origin similar to occurrences elsewhere in the world. Others, however, are intrusions into mixed assemblages of volcanic and sedimentary strata, for instance, the Trembleur Intrusions in the Pinchi Geanticline which cut Permian strata and supplied detritus to Upper Triassic formations (Armstrong, 1949), and the Shulaps body (Leech, 1953) which was emplaced into strata tentatively correlated with Upper Triassic formations and contributed detritus to sediments east of the Yalakom fault, that are possibly of late Early Jurassic age.

Some of the major faults of the western Cordillera, in particular the Pinchi, Shakwak, and Nahlin faults and the Yalakom–Fraser system, may have provided channelways for emplacement of a large amount of ultramafic and associated volcanic material in the Permian and Triassic. These faults, which were generated mainly in the zone between the crystalline rocks of the Coast and Omineca Geanticlines, penetrated to the mantle in the Permian and Triassic and permitted emplacement into the crust of the greatest amount of ultramafic rock in the history of the Cordillera. Subsequent movement along the faults has caused the ultramafic bodies to be displaced so that they now may be in fault contact with adjacent strata.

An exception to the alpine-type are the Tulameen and Polaris bodies. The former consists of a dunite core fringed by clinopyroxenitic rocks, in turn partly bordered by slightly older, undersaturated, alkalic gabbroic rocks similar to Copper Mountain stock, which is dated at 195 m.y. The Tulameen body has intruded the Upper Triassic Nicola Group and has yielded a K–Ar date of 186 m.y., that is, an early Jurassic age. The Polaris body is mainly dunite and olivine-rich peridotite (wehrlite) but it also includes clinopyroxenitic and hornblendic units characteristic of undersaturated ultramafic rocks. K–Ar dates of 164 and 152 m.y. on biotite and hornblende indicate that its age is Middle Jurassic. More than thirty-five similar intrusions in southeastern Alaska are aligned on a northwesterly trend. K–Ar dates from these range from 85 m.y. to 109 m.y. (Lanphere and Eberlein, 1966). Similar bodies may exist in St. Elias Mountains.

PLUTONIC ROCKS

The greatest concentration of plutonic rocks is in the Coast Plutonic Complex and the Omineca Crystalline Belt, especially in the former (Fig. VIII-28). Even though granitic intrusions are common elsewhere in the Cordilleran Orogen their abundance in no way compares to the great volume present in two crystalline belts. Throughout the history of the Cordilleran Geosyncline these geanticlinal belts have been essentially positive and mobile, particularly since the Triassic. They were the sites of repeated uplift and plutonism evidenced not only in the debris shed from time to time and deposited in the adjoining troughs, but also from radiometric data. Clearly, the plutonic rocks, particularly the moderate to high-grade metamorphic rocks, evolved for the most part not in deeply depressed troughs but in the intermittently persistent, uplifted belts marginal to them. Within these belts, then, there were greater concentrations of heat than elsewhere, and from time to time plutonic fronts arose.

Gabrielse and Reesor (1964) indicate that the composition of the plutons in the Omineca Geanticline changes with age from predominantly quartz diorite and monzonite in late Paleozoic, through intermediate types during the Mesozoic, to syenites and quartz monzonites in the Tertiary; furthermore, they conclude that the idea of a granitic series ranging from autochthonous granitized rocks through mobilized material and finally to intrusive granite does not hold. Instead, they consider the plutons to have evolved in successive pulses over a very long period of time. Roddick (1965) has also concluded that the southern part of the Coast Plutonic Complex progressively evolved with continually changing composition over a long period. It appears that many of the plutonic pulses

are broadly synchronous with orogenic spasms. Plutonism accompanied the following orogenies: the Hadrynian East Kootenay; the Triassic Tahltanian; the Jurassic Nassian and Inklinian; the Jura-Cretaceous Columbian; and the late Cretaceous to early Tertiary Laramide. Evidence for plutonism accompanying the Devono-Mississippian Caribooan Orogeny is indecisive. Metamorphic rocks occur either as regionally metamorphosed terrane in which the stratigraphy can generally be recognized or as metamorphic complexes in which the stratigraphy is commonly in doubt. The less metamorphosed rocks occur throughout the Omineca Crystalline Belt, commonly exhibiting the Barrovian sequence of metamorphism (kyanite–sillimanite type) and attaining the upper grades of the almandine–amphibolite facies without the development of much granitic material. The metamorphic complexes, widespread in the Coast Plutonic Complex and underlying several large areas in the Omineca Crystalline Belt, include abundant granitic material, particularly in the cores of the gneiss domes. The regionally metamorphosed rocks commonly yield the youngest K–Ar dates, suggesting that the complexes were not only sites of more intense plutonic activity, but also sites of greater heat flow until relatively late in the history of the region.

Yukon Group

Regionally metamorphosed rocks comprising metasediments, metavolcanics, and paragneisses known as the Yukon Group, underlie large areas of Yukon Plateau. Along the boundary with Alaska these rocks are continuous with the Birch Creek Schist of Alaska. Where fossiliferous, the rocks have previously been described, but where the metamorphic grade is high local names are applied, such as Harvey Group in the Glenlyon area, Big Salmon Complex in the Teslin area, and Oblique Creek Formation in the Jennings area.

In the Dawson and McQuesten areas the grade of metamorphism is significantly higher southwest of Tintina Trench than to the northeast. Northeast of the Trench, in Mayo and McQueston areas, great thicknesses of quartzite and schist with some limestone are partly correlative with the Hadrynian. Southwest of Tintina Trench, in Dawson, Ogilvie, and McQuesten areas, a thick section of quartzites, micaceous quartzites, and quartz–mica schists form the Nasina Group. Sericitized and chloritized intrusive rocks of the Klondike Schist and granitic gneiss termed the Pelly Gneiss are probably in part equivalent to the less metamorphosed Hadrynian rocks to the northeast. Limestones associated with metasediments just south of Yukon River at the Alaska Boundary have yielded crinoid ossicles that indicate a Paleozoic age. In the Dawson area it seems probable that two main sequences of rocks are involved in the regional metamorphism: a lower one of Hadrynian age, and an upper one, including the ultramafic bodies, greenstone, and basic intrusions of Paleozoic age, and possibly in large part Devonian and Mississippian.

Along the Coast Geanticline in southwest Yukon Territory, the Yukon Group consists largely of quartz–mica schists, quartz–biotite–feldspar gneisses, crystalline limestones, slate, and quartzite. These metasedimentary rocks of possibly Precambrian age and later include minor hornblende and chlorite schist presumably derived from volcanic rocks.

Metamorphic rocks in southern Cassiar Mountains include strata as young as Mississippian and probably as old as Cambrian. The Oblique Creek Formation consists of thick sequences of quartz–albite–mica gneiss, albite–actinolite schist, meta-chert, conglomerate, greenstone, and meta-tuff in Jennings and Wolf Lake areas. To the northwest in Teslin area, the Big Salmon Complex includes intercalated crystalline limestones with Mississippian fossils. Farther north, quartz–mica schist and gneiss, granitic gneiss, quartzite, and crystalline limestone are the equivalent of Cambrian and younger strata.

Northeast of Tintina Trench in Finlayson Lake, Frances Lake, and Watson Lake areas, granitic gneisses, characterized by opalescent quartz grains, and associated schists and phyllites were evidently derived from Hadrynian clastic rocks.

K–Ar ages ranging from 178 to 222 m.y. suggest that the Yukon Group was regionally metamorphosed during the Tahltanian Orogeny, and at least locally during the Inklinian Orogeny. The general lack of metamorphism in Upper Triassic rocks suggests that much of the metamorphism took place in the earlier orogeny.

Horseranch Group

The Horseranch Group, in Cassiar Mountains, comprises regionally metamorphosed and locally granitized metasedimentary rocks similar to those of the Wolverine Complex but considerably less granitic (Gabrielse, 1963). It contains recrystallized and locally feldspathized quartzite, quartz–mica schist, granitic gneiss, augen gneiss, hornfels, skarn, and crystalline limestone. In places granitic pegmatite sills and dykes are common. The group is exposed in a fault-bounded, doubly plunging anticline about 30 miles long; stratigraphic relationships are unknown. In gross lithology the Horseranch Group most closely resembles the Lower Cambrian Atan Group. A K–Ar age of 57 m.y. was obtained on muscovite from granitic rocks at the north end of Horseranch Range but whether this dates the regional metamorphism is not known.

Wolverine Complex

The Wolverine Metamorphic Complex in Omineca Mountains is composed of highly metamorphosed and granitized rocks of Proterozoic and early Paleozoic age (Armstrong, 1949; Roots, 1954). It forms the eastern part of central Omineca Geanticline. Granitic material, zones of feldspathization, and pegmatite are concentrated along northeast-trending belts and domes. The domes and accompanying depressions result, in part, from the intersections of north-northwesterly and west-northwesterly



PLATE VIII-7
Veined gneiss of Shuswap Metamorphic Complex, intensely folded, in core of Thor-Odin Dome, central Monashee Mountains, British Columbia.

folds. The former trend, characterized by crenulations and wrinkle lineations of nearly constant orientation, are considered by Roots to have been deformed by the moderately open folds with the latter trend. Furthermore, because the isograds are apparently parallel with the stratigraphic succession and are folded by west-northwest-trending folds Roots argued that the metamorphism was developed during the earlier folding. Schistosity in the complex and adjoining less metamorphosed rocks of the Tenakihi and Ingenika Groups is also parallel with the bedding.

Roots considered that the plutonism took place in early Paleozoic as Wolverine debris occurs in a conglomerate he mapped as Mississippian. However, the conglomerate could be late Lower Cretaceous in which case the plutonism could be mid-Mesozoic and contemporaneous with that in the Shuswap Metamorphic Complex. Upper Jurassic dates on a muscovite–biotite pair of 139 m.y. and 143 m.y. have been obtained from the Misinchinka schist in the Rocky Mountain Trench adjacent to the Wolverine Complex, but the K–Ar dates from the Wolverine Complex fall into two groups, one ranging from 38 m.y. to 43 m.y. and the other 22 m.y. to 29 m.y.

Shuswap Complex

The Shuswap Metamorphic Complex, which underlies the southern culmination in the Omineca Geanticline, comprises rocks ranging in age from Proterozoic to early Mesozoic which are metamorphosed to upper almandine–amphibolite facies. Although many domes are present in the complex, it is dominated by three, paired gneiss domes occurring at 50 mile intervals along the eastern margin. These domes [Valhalla (Reesor, 1965; *Frontispiece*), Thor-Odin (Reesor, 1966), and Frenchman's Cap (Wheeler, 1965)] have cores of veined augen gneiss and granite gneiss (Pl. VIII-7) intertongued with and mantled by hybrid metasedimentary gneiss and schist, the outer part of which is riddled with pegmatite and leucogranite and surrounded by a fringing zone of low-grade rocks. Nowhere in the cores of the domes has an ancient crystalline basement been identified. On the contrary, Reesor (1966) concluded that the core of Thor-Odin Dome is composed of Windermere and lower Paleozoic strata. The pegmatite-rich zone is particularly extensive west and northwest of Frenchman's Cap Dome and west and south of Thor-Odin Dome. The fringe of low-grade rocks is commonly featured either by a zone of crushing and



PLATE VIII-8. Fold-mullion structures parallel with axes of recumbent folds in hybrid gneiss of Shuswap Metamorphic Complex mantling Valhalla Dome, southern Selkirk Mountains, British Columbia.

mylonitization, or by a rapid gradation from sillimanite–almandine–muscovite and staurolite-bearing rocks to greenschist. The distinctive structural style of the complex, apart from the development of domes, is an easterly trending lineation marked by mineral orientation and the axes of isoclinal and recumbent folds (Pl. VIII-8).

In the salient northeast of Revelstoke, the foliation within wedges of gneiss emplaced into Shuswap gneisses antedates the formation of the Shuswap gneissosity according to J. V. Ross. The earliest recognized structures belonging to the development of the Shuswap gneisses are large isoclinal recumbent folds. These, and the local, smaller isoclinal folds refolding them, probably result partly from flowage off rising and expanding easterly trending warps on the present sites of the domes and partly synchronously with subsequent diapirism. Following the movement that developed an easterly trending penetrative mineral lineation, the complex was redeformed resulting in superimposed northerly and northwesterly trending folds and lineations. This later deformation produced domes and depressions that formed at shallow levels and included the upward diapiric movement of the cores of the domes and possibly also upthrusting of gneiss wedges. As the metamorphism waned, movement along the eastern edge of the complex induced cataclasis, mylonitization, alteration, and late faulting with brecciation.

The petrogenic evolution within the Shuswap Complex has been established only in Valhalla Dome (Reesor, 1965) and is summarized in Table VIII-1 together with

TABLE VIII-1 *Summary of events in evolution of Valhalla Complex (Reesor, 1965)*

		Structural Events	Petrogenic Events
↑ Second Phase Late Stage	Locally confined	8. Late faulting in 'crushed zone'	8. Pyritization
		7. Minor faulting and jointing in complex	7. Basalt dyke intrusion in western Valhalla. Sericitization and epidotization in 'crushed zone' east of Valhalla Complex
		6. Igneous emplacement under regional tensional conditions	6. Intrusion of hornblende and/or augite monzonite both east and west of Valhalla Complex
	Destructive	5. Diapiric upward movement of cores of Valhalla and Passmore Domes	5. Possibly some synchronous emplacement of massive leucogranite in zone west of Valhalla Complex
		4. Strong, penetrative, cataclastic flowage in all layers of all rock types of eastern part of Valhalla Complex. Easiest relief of stress vertically upward	4. Minor pegmatite and pegmatitic granite. Metamorphism in garnet-sillimanite subfacies of amphibolite facies
		Termination of principal constructional phase of evolution	
↑ First Phase	Constructive	3. Minor structural impulses, with gradual growth of antiformal masses. Shape and location of Passmore Dome dependent on formation of Valhalla Dome and thus slightly after	3. Leucogranite and early pegmatite in all layers of the complex. Grey granodiorite–gneiss in all layers
		2. Initial uplift of Valhalla Dome with original form determined in east-west elongate antiformal complex. Coincident flowage of 'hybrid gneiss' to the flanks of Valhalla Dome governed by shape of the confining space. Initial syntectonic movement confined so that easiest relief of stress horizontal	2. Metamorphism to perhaps hornblende granulite facies, with partial melting of pelitic and semi-pelitic rocks. Horizontal emplacement of quartz diorite and granodiorite of 'veined gneiss' in Valhalla Dome and upper layer ('mixed gneiss') of Passmore Dome
		1. Possible earlier structural episodes not recognized	1. Deposition of thick succession of pelites, semi-pelites, quartzite, minor limestone. Minor, conformable, basic igneous rocks

associated structural events. Reesor noted that " . . . no single process, and no single associated structural episode is responsible for the formations of the complex. It is, rather, the sum of many processes, structural and petrogenic, evolving successively through a long period." He noted that the early structural episode was characterized by horizontal movement of lenticular migmatitic lenses of augen granodiorite gneiss in the core. This was followed by accumulation of granitic and pegmatitic material. There had been accumulation and growth rather than compression and shortening. After stagnation, a second phase followed that was essentially deformational and expressed as upward diapiric movement. The least migmatized, though highly metamorphosed metasediments, are lowest in the structural succession and therefore neither simple magmatic emplacement from below nor simple granitization in situ is possible for the formation of the augen gneiss. Furthermore, at some levels, and at some localities, the amount of quartzo-feldspathic material is so low that even the melting or transforming of all the rocks would be insufficient to produce the amount of granite present. Reesor therefore concluded that some of the granitic material was derived from beyond the complex in a succession of impulses, although some may have originated from melting in situ.

Coast Plutonic Complex

The Coast Plutonic Complex extends the length of Coast Mountains for 1,100 miles, merging northward with the Yukon Crystalline Platform and terminating against the Cascade Fold Belt in the south. The complex is composed mainly of foliated and non-foliated granitoid plutonic rocks but also contains large areas of metamorphic rocks (Roddick, *et al.*, 1966) (Pl. VIII-9). An irregular belt of narrow, elongate, steep-walled roof pendants of metasedimentary rocks stretches southeastward from within the Wrangell–Revillagigedo gneiss belt on the west side of the complex in southeastern Alaska to near the eastern side, near Bella Coola. The metasedimentary rocks range from greenschist to almandine–amphibolite facies and comprise schist, quartzite, limestone, and conglomerate, locally carrying granitic clasts. East of the metasediments and apparently underlying them is a belt of gneiss and migmatite that has a gently dipping foliation over large areas and locally exhibits recumbent folds with easterly and northerly trending axes. The gneiss may be overlain and intruded by granitic rocks, or grade into them. East of Kemano the gneisses appear to underlie Permian metasediments and rocks forming part of the pre-Hazelton Tahtsa Complex (Stuart, 1960).

More or less homogeneous plutonic rocks form a large part of the complex between latitudes 52° and 55°. Quartz diorite and diorite are abundant west of the metasedimentary belt whereas granodiorite and quartz monzonite bodies occur throughout the complex. Granodiorite is characteristic of the composite plutons such as the northeast-trending apophyses near Skeena River, regarded by Duffell and Souther (1964) as forceful intrusions. The quartz monzonite bodies are distinct, young plutons, some forming the core of the granodiorite bodies. Syenite and granite are rare and occur only in small plutons in an east-northeast-trending belt west of Bella Coola. Roddick, *et al.* (1966) stated that a quartz diorite line cannot be drawn through the complex, but remarked that the potassium feldspar content east of the metasedimentary belt is greater than to the west. In southeastern Alaska, south of Stikine River, quartz diorite is dominant on both sides of the Wrangell–Revillagigedo gneiss belt but granodiorite and quartz monzonite are more common to the east.

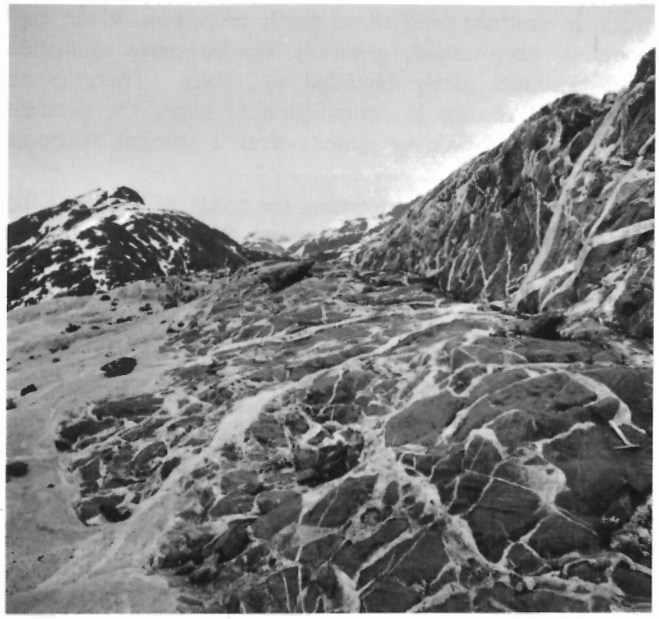


PLATE VIII-9. Agmatite, Coast Plutonic Complex, Douglas Channel west of Kiltsuit River, British Columbia. Aplitic leucogranite veining and surrounding angular blocks of granular amphibolite and granodiorite.

Northwest of Stikine River the Coast Plutonic Complex contains narrow, steep-walled belts of metasediments and amphibolite but is composed mainly of an interlocking system of discrete plutons. Near the British Columbia–Yukon boundary leucogranite and granodiorite plutons intrude older foliated quartz diorite that now appears as a matrix between the younger plutons (Christie, 1957). Farther south, near Stikine River, the eastern part of the complex contains pre-Late Triassic foliated diorite plutons and metamorphic rocks and younger, steep-walled plutons of quartz monzonite, whereas the axial part is characterized by granodiorite.

The southernmost part of the complex has been studied in the greatest detail (Roddick, 1965). It is a heterogeneous complex of plutons ranging in composition from diorite and quartz diorite to granodiorite. Gabbro and granite are rare. The average composition is that of quartz diorite with hornblende exceeding biotite, and potassic feldspar totalling about 5 per cent. This part of the complex is characterized by diffuse interplutonic con-

tacts, in contrast with those north of Stikine River, and contains steep-walled, generally northwesterly elongated roof pendants partly bounded by faults. There is no appreciable change in composition of either the pendant rocks or the adjoining plutons over a vertical range of 7,000 feet.

Opinions differ concerning the mode of origin of the plutonic rocks in the southernmost part of the complex. Bacon (1957) and others emphasize that the invaded rocks, which had previously been metamorphosed, have not been greatly disturbed by the emplacement of granitic rock and invoked the passive intrusion of andesitic magma. Roddick (1965) concluded that the granitic rocks formed from 'wet' pre-existing metamorphic rocks by evolution through a plutonic series, mainly from hornblende diorite towards biotite granite in the course of a long period of retrograde metamorphism during the gradual uplift of the Coast Geanticline. The evolution was accompanied by mass movement of plutonic rock partly bounded by synplutonic faults and partly as plastic, saltdome-like intrusions. Roddick has based this conclusion on numerous microscopic and megascopic data in which the occurrence of relict dykes is particularly striking (Roddick and Armstrong, 1959). The 'synplutonic' dykes intruded earlier consolidated plutons and were eventually partly transformed into granitic rock by continuing metasomatism. Locally this happened after the dykes were fractured, faulted, or deformed within remobilized plutons.

The Coast Plutonic Complex exhibits several periods of plutonism. It is not known when these began or whether the complex experienced the late Ordovician period of granitic emplacement known in southeastern Alaska or the pre-Devonian metamorphism and diorite intrusion evident in Cascade Mountains and on San Juan Islands. The earliest phase of plutonism for which there is evidence took place during the mid-Triassic Tahltanian Orogeny in the Stikine River region, followed there by the early Jurassic Inklinian Orogeny. Episodes of plutonism elsewhere, perhaps coeval in part with the Tahltanian or Inklinian phases, took place mainly in the southeastern part of the complex. K–Ar dates, ranging from 139 m.y. to 95 m.y., indicate episodes of plutonism in the western third of the complex during the early and middle phases of Columbian Orogeny. The K–Ar dates ranging from 77 m.y. to 42 m.y. come mainly from quartz monzonite and granodiorite plutons in the central and eastern parts of the complex. Some plutons intrude Upper Cretaceous strata and support radiometric evidence for granitic emplacement during latest Cretaceous and early Tertiary time. Many of the young plutons have been intruded along the eastern margin of the complex into regions where granitic rocks were emplaced in the early Mesozoic. Thus the Coast Geanticline, an uplifted zone that was emergent since the mid-Triassic, was for the same period a zone of repetitive plutonism.

Cascade Complex

In Cascade Fold Belt, clastic sediments as young as Albian are folded, faulted, and intruded by a granitic stock with a K–Ar date of 84 m.y. The deformation is referred, accordingly, to the late phase of the Columbian Orogeny. However, in the northwest steep reverse faults, dipping southwest or locally northeast, involve Upper Cretaceous volcanics, and some deformation also took place later. In western Cascade Mountains early Lower Cretaceous and older beds are involved in westerly and northwesterly directed recumbent folds and thrusts transported over an autochthone featured by a northeasterly trending, faulted anticline with pre-Devonian basement in its core (J. W. H. Monger). The recumbent structures are redeformed by northwesterly trending warps, faults, kink bands, and minor conjugate folds, and in Washington State are overlain unconformably by late Upper Cretaceous clastic sediments.

Cascade Fold Belt underwent various types of regional metamorphism at different times. In the core of the belt the migmatitic Custer gneiss developed from the late Paleozoic Hozameen Group (McTaggart and Thompson, 1967) during regional metamorphism of Barrovian type in latest Paleozoic or earliest Mesozoic time, contemporaneously with the Skagit metamorphism in Washington (Misch, 1966). Later, parts of the Hozameen Group were subjected to low pressure, regional or contact metamorphism featured by andalusite. Finally, in latest Cretaceous time, coeval with the emplacement of the tonalite and diorite of the Spuzzum Intrusions, dated at 76 m.y., the metamorphic rocks were raised to sillimanite grade.

The western part of the Cascade Fold Belt is characterized by regional metamorphism of glaucophane schist type which in Washington is considered by Misch to be of Permo-Triassic age. In the Cascade Mountains of Canada J. W. H. Monger has discovered lawsonite in Mesozoic clastic sediments but is uncertain whether it is of detrital or metamorphic origin.

MESOZOIC

Triassic

Tectonic Summary

Lower Triassic rocks are not known in the western parts of the Cordilleran Geosyncline in Canada although they are present a few miles south in Washington. Middle Triassic ribbon chert, argillite, greenstone, and possibly coeval ultramafic rocks, and minor limestone occur locally in the northwestern and central parts, apparently lying conformably on the Permian. In southernmost parts of the geosyncline, Middle Triassic sediments featured by sharpstone conglomerate, unconformably overlie late Paleozoic

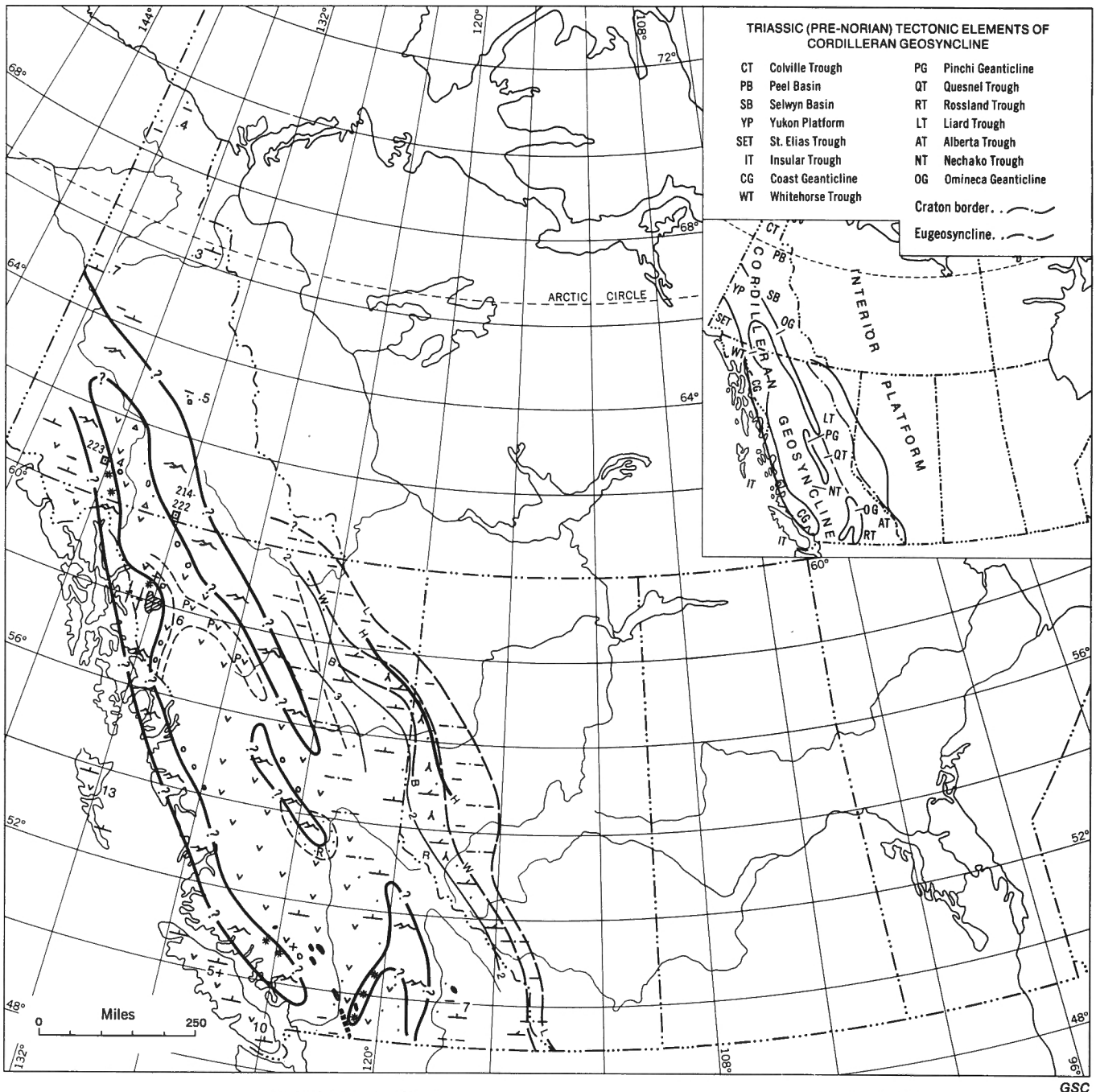
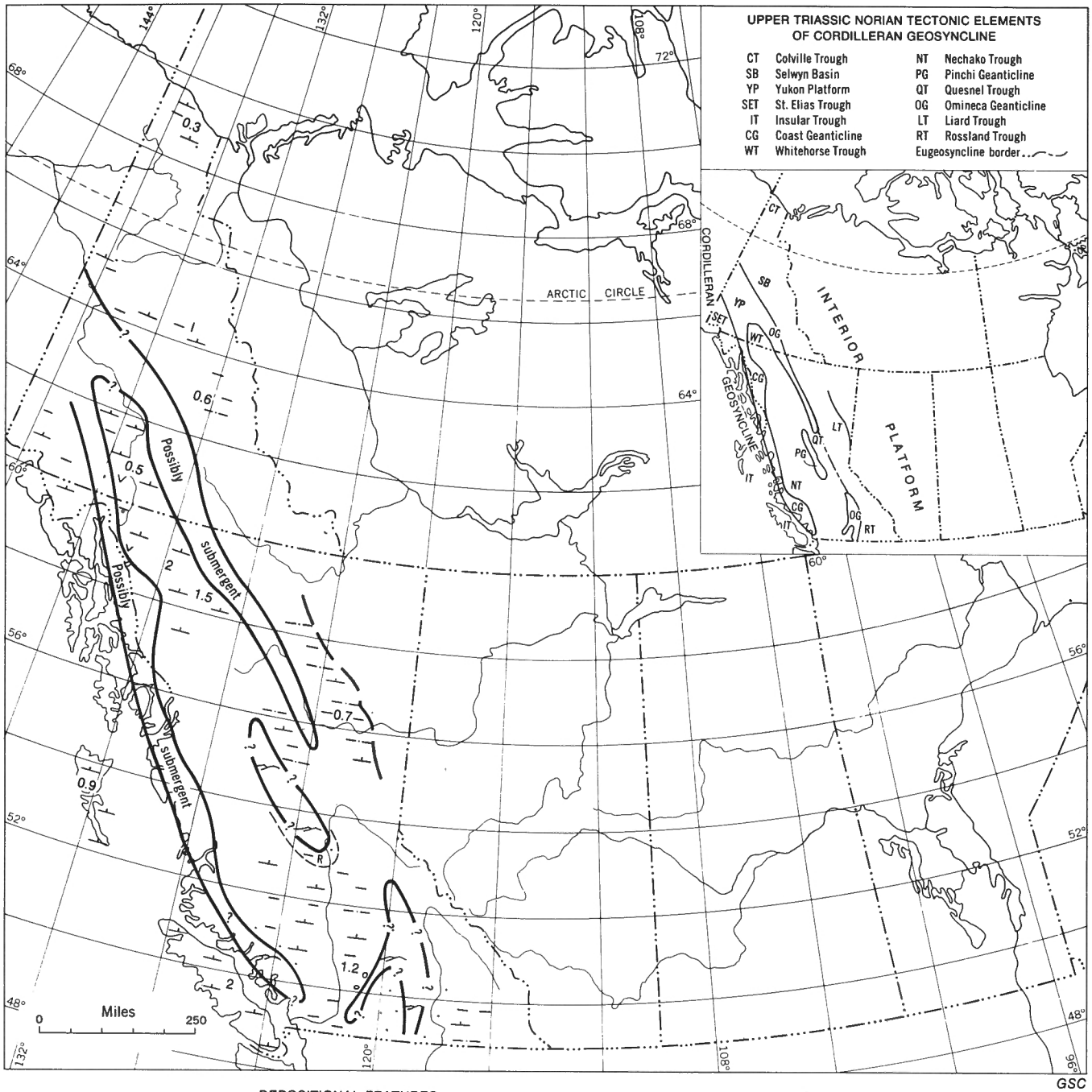


FIGURE VIII-29. Upper Triassic, Karnian, sedimentation and tectonism in Western Canada. Includes Lower and Middle Triassic sediments of Liard and Alberta Troughs and some features of the early and mid-Triassic Tahltanian Orogeny.



DEPOSITIONAL FEATURES

Conglomerate . . . . . o o

Volcanics V

Siltstone . . . . . —·—

Shale . . . . . —

Limestone . . . . . —

Red beds . . . . . R

Depositional limit . . . . . ~~

Truncated limit . . . . . ~~

Facies boundary . . . . . ~~

Thickness

(thousands of feet) 1.2

PHYSIOGRAPHIC FEATURES

Low mountainous land . . . . . ~~

Blank areas are lowlands and plains developed on crystalline and sedimentary rocks of the craton, and where the Norian is now absent through erosion, or unknown. In the geosyncline the blank areas may be sea.

FIGURE VIII-30. Upper Triassic, Norian, sedimentation and tectonism in Western Canada.

rocks. The western sequences represent generally quiescent eugeosynclinal conditions possibly prevailing from the Permian. The Omineca Geanticline, an intermittently emergent arch during most of the period, separates the eugeosyncline from the miogeosynclinal Alberta and Liard Troughs (Fig. VIII-29). Lower Triassic siltstone and shale in the troughs disconformably overlie the early Upper Permian, the hiatus increasing eastward on the craton through overlap and bevelling of the Permian and Carboniferous. The entire western margin of the craton was probably emergent and stable, these conditions prevailing throughout the period. Triassic red beds are known in Williston Basin in the northern United States, but if formerly they extended into Canada, they have been removed by pre-Jurassic erosion.

In the late Middle Triassic, parts of the Cordilleran Geosyncline underwent deformation and plutonic activity constituting the Tahltanian Orogeny. Uplift at the end of the orogeny established the main tectonic elements that prevailed until mid-Jurassic time. The early Upper Triassic, Karnian, is represented in Insular Trough mainly by a thick succession of submarine basaltic flows and in Whitehorse Trough principally by andesitic and basaltic flows, pyroclastics, and clastics containing volcanic and mid-Triassic and earlier debris. The latter for the most part were derived from islands and volcanoes forming Coast Geanticline and Stikine Arch. Similar volcanics and volcanic clastics accumulated during the Karnian in most of Nechako and Quesnel Troughs although some red beds were deposited around the fringes of Pinchi Geanticline. Quartzose clastics derived from southernmost Omineca Geanticline were laid down in eastern Quesnel and Rossland Troughs. In Liard and Alberta Troughs siltstone and sandstone accumulated, grading eastward, along the margin of the craton, into an evaporitic and red bed facies with basal shoreline and offshore sandbars.

In the late Upper Triassic, Norian (Fig. VIII-30), volcanism persisted in Whitehorse Trough adjacent to the Coast Geanticline and Stikine Arch, and in Nechako and Quesnel Troughs. Clastic sediments accumulated as red beds around the uplifted Pinchi Geanticline and were also derived from the southern part of the Coast Geanticline and parts of the Omineca Geanticline. The latter was locally covered by thin limestone. Clastics were deposited in Insular Trough. During the latest Norian carbonate deposition prevailed throughout much of the geosyncline, as the sections were remarkably thin but complete in the northern and eastern elements, and indicated quiescent conditions of deposition to the end of the Triassic.

Lower and Middle Triassic of Western Cordilleran Geosyncline

An assemblage of argillite, chert, greenstone, and minor limestone similar to that in the Cache Creek Group overlies apparently conformably the youngest Permian at several localities. This assemblage includes unfossiliferous rocks in the Atlin Horst, regionally metamorphosed beds yielding Ladinian *Daonella* in the Stikine Arch (Souther, 1959), and unmetamorphosed strata containing *Daonella* in St. Elias Trough. The Pavilion Group, containing hexacorals, occurs near Lillooet close to the southwestern end of the Pinchi Geanticline. Trettin (1961) considered the Pavilion Group to overlie the Upper Permian Marble Canyon Formation but Campbell and Tipper (1966) regard some of the Pavilion Group to be an older part of the Cache Creek Group that has been thrust over the Marble Canyon Formation. Near Greenwood an assemblage of sharpstone conglomerate, siltstone, and limestone with *Daonella* lies with angular unconformity on upper Paleozoic rocks of the southern Omineca Geanticline. The Fennell Formation of pillowed greenstone at the southeastern end of Quesnel Trough underlies fossiliferous Karnian beds. It and beds containing *Daonella* or *Halobia* along the eastern Coast Mountains near Skeena River and near Chilko Lake may be of Middle Triassic age. Near Skeena River the Triassic beds contain limestone boulders and chert-pebble conglomerate lying unconformably on Permian beds.

Tahltanian Orogeny

In the Middle Triassic, during the Tahltanian Orogeny, parts of the Cordilleran eugeosyncline were deformed, intruded by diorite and granodiorite, and uplifted (Fig. VIII-29). An early phase is evident in the southernmost Cordillera near Greenwood, where Middle and possibly Lower Triassic sharpstone conglomerate unconformably overlies upper Paleozoic rocks. A late phase in late Ladinian time affected ancestral Stikine Arch where Middle Triassic and earlier strata were folded along northerly trends and were more severely deformed and metamorphosed than the succeeding Upper Triassic and Lower Jurassic beds that were later folded along northwesterly trends. Plutonism during the Triassic in the Coast and Omineca Geanticlines is indicated by the following K–Ar dates: 206 and 227 m.y. from dioritic boulders in Lower Jurassic conglomerates on Stikine Arch, presumably derived from a nearby pluton (Fig. VIII-28); 223 m.y. from a hornblende–biotite granodiorite in the Coast Geanticline in southern Yukon Territory; 214 to 222 m.y. from muscovite–biotite–albite–epidote gneiss in the northern Omineca Geanticline; and 205 m.y. from greenschist and a granitic pluton in the southern Omineca Geanticline. Widespread uplift is inferred from the erosional unconformity between Upper and Middle Triassic beds in the northern Coast Geanticline and the southern Omineca Geanticline, and between the Upper Triassic and upper Paleozoic along the flanks of the Pinchi Geanticline and in Cascade Mountains.

It is uncertain how much deformation of upper Paleozoic rocks in the Pinchi Geanticline took place during the Tahltanian Orogeny. The structures in these rocks are roughly parallel with the long axes of the geanticline and are characterized mainly by moderately tight, generally upright folds in relatively competent lime-

stone beds and by more irregular and severe deformation in the argillaceous and cherty rocks. On the other hand the Mesozoic rocks of adjoining Nechako and Quesnel Troughs are gently folded, warped, and faulted and this contrast in tectonic style may indicate an orogeny in the Pinchi Geanticline. It may reflect, however, an earlier pulse in Late Pennsylvanian or Early Permian time or merely differences in structural behaviour as both successions were probably deformed during later Mesozoic orogenies.

Upper Triassic of Western Cordilleran Geosyncline

The stratigraphy of the Upper Triassic in the western Cordilleran Geosyncline is best known along the flanks of Whitehorse Trough and in Insular Trough. In these regions the Karnian and Norian rocks are fossiliferous and easily recognized, but in Nechako and Quesnel Troughs volcanic rocks are widespread and fossils scarce.

Whitehorse Trough. Karnian deposits along the western side of the Whitehorse Trough (Figs. VIII-28, -29, -31a) are included in the lower part of the Lewes River Group of southern Yukon (Tozer, 1958; Wheeler, 1961) and in the Stuhini Group and its eastern clastic facies, the King Salmon Formation, exposed along the eastern part of northern Coast Mountains (Fig. VIII-32a, b) (Souther and Armstrong, 1966). Along this belt Karnian beds are predominantly volcanic. The more westerly comprise andesitic flows, pyroclastics, and abundant volcanic clastics derived partly from the deposition of pyroclastics in water and partly from the erosion of volcanic rocks. Some of the volcanic rocks are apparently subaerial and the sequences contain disconformities resulting from local uplift and the building of volcanic cones. Gneiss and granodiorite clasts in the lower parts of the Stuhini Group suggest an early Karnian uplift in the western part of Stikine Arch and that the volcanic rocks were deposited on a terrain deformed and metamorphosed during the Tahltanian Orogeny. The Lewes River Group grades eastward into greenstone and graded-bedded volcanic arenite,

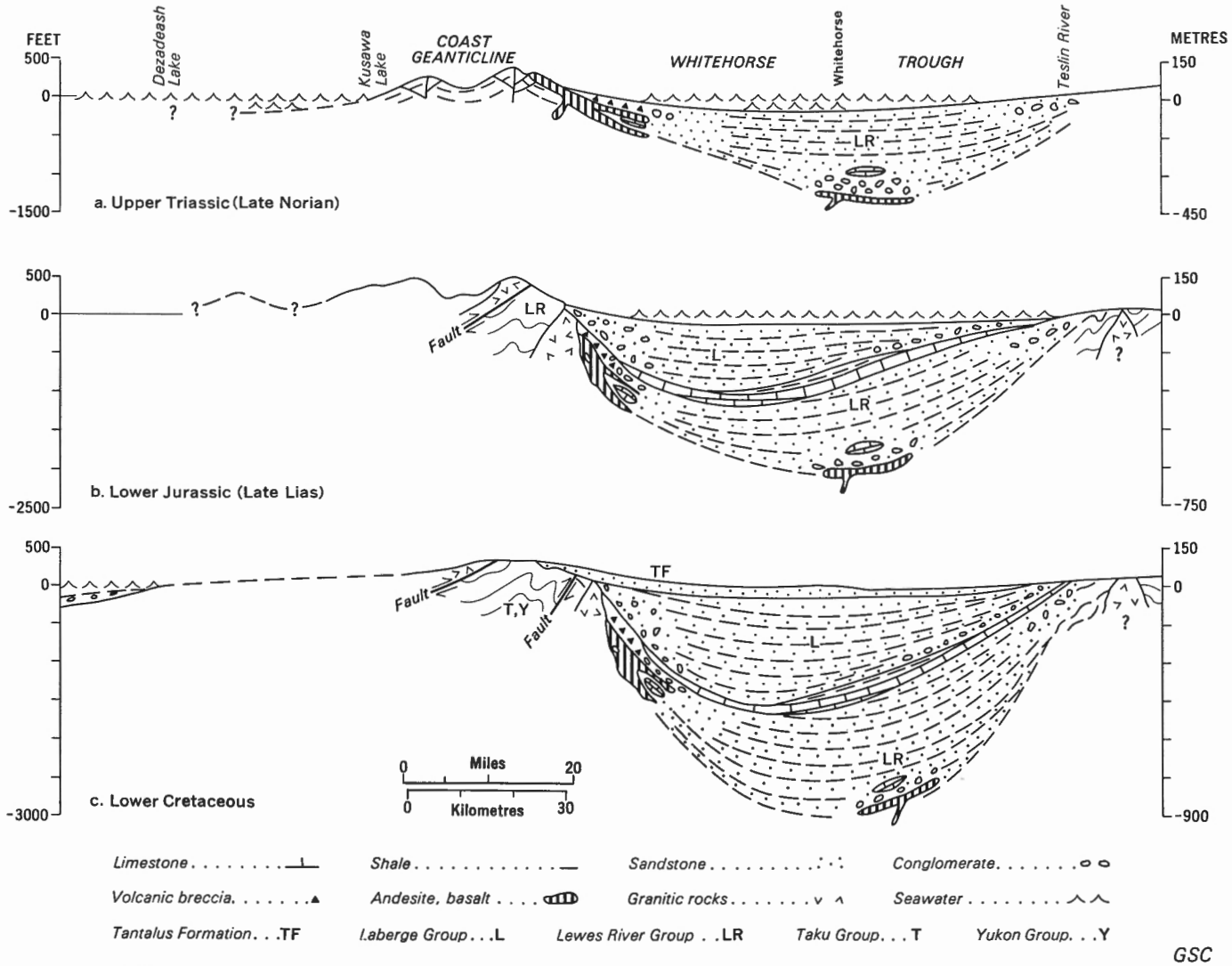


FIGURE VIII-31. Diagrammatic restored sections across Whitehorse Trough, Yukon Territory, during the Mesozoic era (Wheeler, 1961).

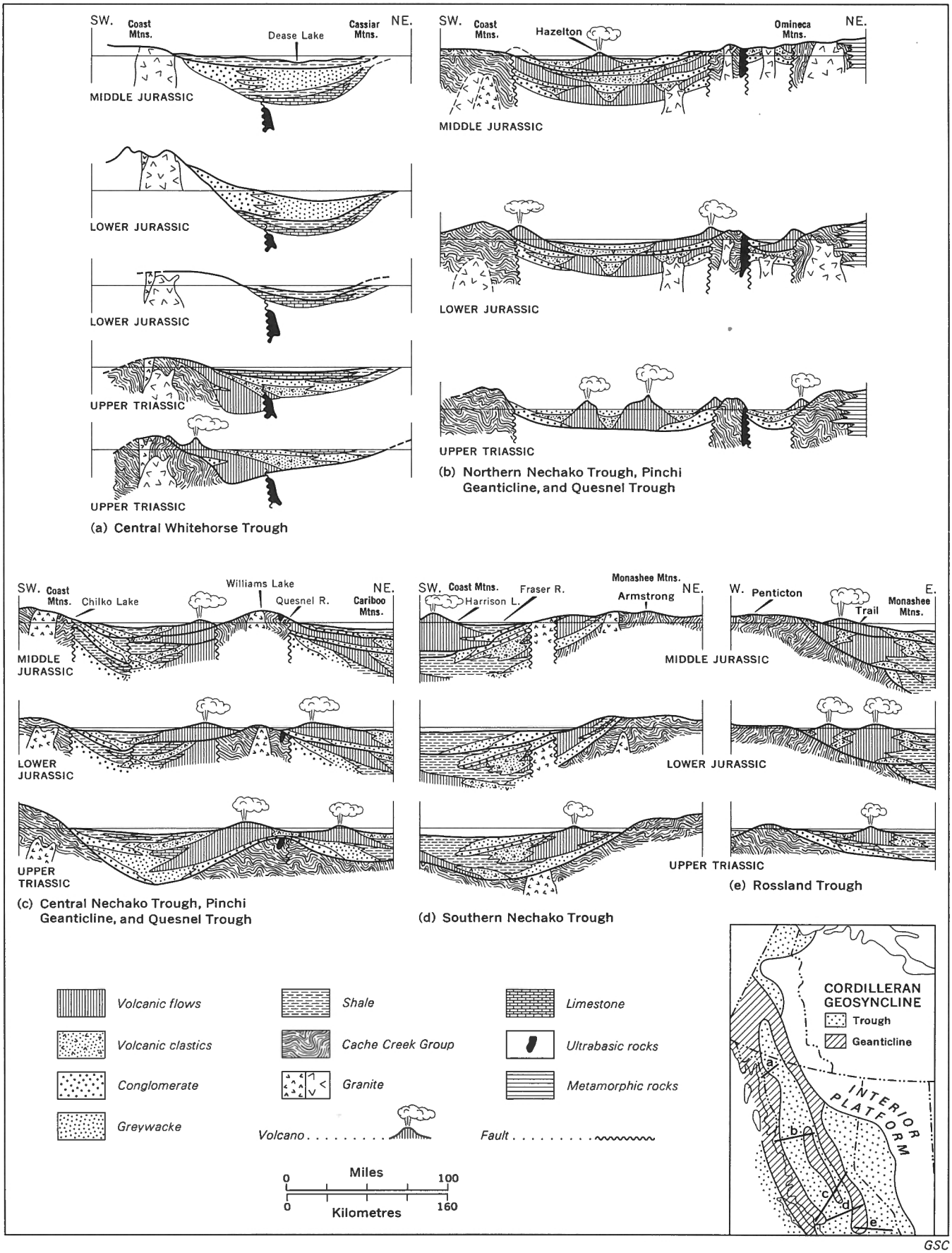


FIGURE VIII-32. Diagrammatic restored sections across Whitehorse, Nechako, Quesnel, and Rossland Troughs of the Cordilleran Geosyncline from Upper Triassic to Middle Jurassic times (Souther and Armstrong, 1966; Campbell, 1966).

whereas the more easterly outcrops of the Stuhini Group include spilitic pillowed basalt more than 6,000 feet thick that forms an arcuate belt. Still farther east volcanic clastic and fine-grained sedimentary rocks are predominant.

Volcanic rocks are much less extensive in the Norian than in the Karnian as they are restricted principally to the western exposures of the Lewes River Group where they overlie a mid-Norian *Halobia* fauna and underlie limestone carrying latest Norian *Spondylospira lewesensis.* The volcanic rocks comprise about 2,500 feet of andesitic and basaltic flows and pyroclastics. The latter carry fragments as long as 8 inches and were probably derived from sources no more than 5 or 10 miles to the west. The pyroclastics become finer eastward and, east of Yukon River, pass into graded-bedded volcanic arenite deposited in the central part of Whitehorse Trough. The arenite is overlain by late Norian argillite containing *Monotis subcircularis* succeeded by limestone yielding *Spondylospira.* These late Norian sediments grade farther northeastward, near Teslin River, into an assemblage of greywacke, limestone, and some pebble-conglomerate with greenstone and chert clasts deposited along the northeastern margin of Whitehorse Trough.

In Stikine Arch the upper part of the volcanic Stuhini Group may be partly Norian. The overlying sediments, principally limestones of the Sinwa Formation, are definitely Norian. This limestone extends discontinuously across the Whitehorse Trough to the Atlin Horst and indicates that the latest Norian was a period of tectonic quiescence within Whitehorse Trough.

Nechako and Quesnel Troughs. Much of the Interior Plateau of British Columbia and the southern Omineca Mountains is underlain by the poorly exposed and sparsely fossiliferous Upper Triassic and Lower Jurassic Takla and Nicola Groups and related strata that were deposited in the Nechako and Quesnel Troughs (Figs. VIII-29, 32c, d) (Cockfield, 1948; Tipper, 1959). The Upper Triassic rocks are predominantly dark grey to dark green basic andesite and basalt flows with augite phenocrysts, pyroclastics, and volcanic conglomerate and arenite. Intercalated within the volcanics are minor sedimentary rocks. Dark grey to black shaly limestones typically carry Karnian fauna in the south-central part of the region whereas grey limestone and dark shales commonly yield Norian fossils. The Takla Group fringing the Pinchi Geanticline near Prince George is composed of red and brown shales, black limestone, and coarse conglomerate. The latter contain clasts as large as 18 inches in diameter of andesite, porphyritic andesite, quartz, and chert, apparently derived from the Cache Creek Group as a result of uplift of the Pinchi Geanticline. Near Fort St. James the Takla Group contains clasts of serpentine in conglomerate and grains of chromite in the matrix that were derived from ultramafic intrusions emplaced into the Cache Creek Group of the Pinchi Geanticline (Armstrong, 1949).

Volcanic rocks of the Nicola Group are most abundant and the pyroclastics are coarsest along a north- to northeast-trending belt south of Kamloops where they probably represent an axis of late Triassic volcanism (Fig. VIII-29). To the east near Hedley, equivalent strata contain considerable quartzose clastic sediments (Rice, 1947) that must have had a siliceous source farther east, in the Omineca Geanticline. Furthermore, Nicola and equivalent strata apparently overlie older rocks with erosional unconformity west of Okanagan Valley (Cockfield, 1948). To the west the volcanic rocks of the Nicola Group pass into clastic rocks, as in western Cascade Mountains the Cultus shales disconformably overlie the Chilliwack Group. Farther north, near Bridge River, the Noel, Hurley, and Tyaughton Formations are principally clastics intercalated with some andesitic volcanics constituting the Pioneer Formation. The Tyaughton Formation ranges in age from Norian to Hettangian and contains Rhaetian fossils. It is featured by near-shore coquina-bearing limestone interbedded with red and green conglomerate, the clasts of which are mainly volcanic rocks, chert, some argillite, and rare granitic rocks (H. W. Tipper). This facies is indicative of the southwestern margin of Nechako Trough, as the sediments are derived from the uplifted Coast Geanticline.

Omineca Geanticline. Late Karnian or Norian limestone, 30 feet thick, lies on Permian (?) volcanic and sedimentary rocks on the east side of Cassiar Mountains (Gabrielse, 1963), the remnant indicating that during the late Triassic in this region the Omineca Geanticline was slightly depressed. To the southeast, near Prince George and in Rocky Mountain Trench fossiliferous Upper Triassic clastic rocks occur, along with volcanic rocks similar to the Takla Group. The regions east of Okanagan Valley now underlain by the Shuswap Metamorphic Complex and granitic plutons were probably a source not only for the quartzose clastics near Hedley but also for part of the Slocan Group that was deposited in Rossland Trough to the east.

Rossland Trough. Typical argillaceous flysch of the Slocan and Ymir Groups underlies the inner or western part of Kootenay Arc. Fossils dated only as Triassic have been obtained from the Slocan but none occurs in the Ymir. The thickness of the Slocan Group is controversial, estimates ranging from 6,800 feet to 37,000 feet. Nevertheless the thickness is considerable and it would appear that the sediments accumulated in a deeply subsiding trough that lay east of the Omineca Geanticline, roughly on the site of Lardeau Trough, first evident in the early Paleozoic (Fig. VIII-32e). The base of the Slocan Group contains debris of the underlying Kaslo Group. The Kaslo consists of pyroclastics and andesitic and dacitic lava within which lies a belt of serpentine. The Kaslo Group is regarded by Cairnes as lying unconformably on strata mapped as Milford Group but which contain Mesozoic fossils.

Insular Trough. The Upper Triassic strata of Queen Charlotte, Vancouver, and some of the smaller islands off the southern mainland were deposited in Insular Trough west of the Coast Geanticline. The Lower Karnian is represented by 10,000 to 14,000 feet of volcanic rocks of the Karmutsen Group and related formations that lie disconformably on the Permian Sicker Group. The volcanics are sodic basalt comprising massive submarine flows, pillow lava, pillow breccia, and aquagene tuff that resulted from the breaking up of glassy, vesicular submarine lava upon being quenched in sea water (Carlisle, 1963). The Upper Karnian and most of the Norian is represented by the limestone of the Kunga Formation in the Queen Charlotte Islands, and by the Quatsino on Vancouver Island. Upper Norian clastic sediments occur at the base of the overlying Bonanza Group, which near Cowichan Lake on Vancouver Island consist of arenite and conglomerate, with fossil logs (Fyles, 1955).

St. Elias Trough. In the St. Elias Mountains late Middle Triassic *Daonella*-bearing calcareous siltstones lie apparently disconformably on Permian strata and are overlain by volcanic rocks of the Mush Lake Group (Muller, 1967). The latter, in Kluane and Donjek Ranges, is probably correlative in part with the Nikolai greenstone of eastern Alaska that lies between the *Daonella* beds and Karnian limestone. Although near the Alaskan border the Upper Triassic consists mainly of volcanic rocks, farther southeast it comprises a Karnian limestone, equivalent to the Chitistone and Nizina limestones in Chitina Valley area of eastern Alaska (Moffitt, 1938), a lower Norian limestone, and upper Norian shales that contain *Monotis subcircularis* and are equivalent to the McCarthy shale in eastern Alaska.

Triassic of Northern Cordilleran Geosyncline

In *British Mountains,* about 1,000 feet of interbedded silty shale, argillaceous siltstone, and fine-grained sandstone with occasional carbonate beds occur near the Alaska–Yukon boundary (Mountjoy, 1967a). They are referred to the Ivishak Member of the Sadlerochit Formation that contains early Triassic (Griesbachian) ammonites in Alaska. Farther east in British Mountains these beds have not been recognized and Upper Triassic strata unconformably rest on Carboniferous and older strata. The Upper Triassic Shublik Formation comprises 270 to 300 feet of brown-weathering, pelecypod limestone coquina and overlying siltstone and fine-grained sandstone between Firth and Malcolm Rivers. The rocks appear to overlie pre-Mississippian strata of the Neruokpuk Formation with angular unconformity and locally a thin basal conglomerate is present. The moderately resistant Upper Triassic sequence is overlain by recessive weathering Jurassic shales.

Isolated remnants of Triassic strata outcrop along the eastern side of *Richardson Mountains* on Rat and Vittrekwa Rivers and near Caribou and Peel Rivers. On Rat River between 50 and 65 feet of carbonaceous shales with lenses of impure coal unconformably overlie undated chert and sandstone pebble-conglomerates that occur above Permian clastics. They are overlain conformably by Middle Jurassic sandstone. Spores and pollen from the carbonaceous strata indicate a Middle and/or Upper Triassic age. On Vittrekwa River, 90 feet of fine-grained calcareous sandstone and limestone of probable Upper Triassic age unconformably overlie the upper Devonian Imperial Formation and disconformably underlie Upper Jurassic sandstones. Farther south between Caribou Lake and Peel River about 300 feet of calcarenite with a few beds of silty limestone also contain a fauna of probable Upper Triassic age (Mountjoy, 1967a).

In the *Ogilvie Mountains* about 1,000 feet of recessive shales and platy, argillaceous limestones of the Shublik Formation occur in the trough of Monster syncline, disconformably overlying Permian strata and unconformably overlain by Upper Cretaceous rocks. Although this sequence has yielded only an Upper Triassic fauna, similar strata about 45 miles to the west in Alaska contain Middle Triassic Ladinian ammonites. About 1,000 feet of poorly exposed, thin-bedded, black shales and limestones on Rackla River are referred to the Shublik Formation.

Black shale and slate 2,000 feet thick occur in the Tombstone–Tintina area (Tempelman-Kluit, 1966). A basal member, 100 feet thick, of slaty, impure, fetid limestone contains Middle and Upper Triassic fossils. Much of the slate and shale could be of Jurassic age. More than 500 feet of interbedded micaceous sandstone and shale with minor conglomerate and concretionary shale containing an Upper Triassic fauna outcrops along Tay River (Roddick and Green, 1961b).

Triassic of Eastern Cordilleran Geosyncline and Interior Platform

Triassic rocks in the *Rocky Mountains* between the International Boundary and Smoky River are included in the Spray River Group. The group totals about 1,700 feet at North Saskatchewan River (Best, 1958), increasing to 2,800 feet near Smoky River (Gibson, 1968), and is relatively thin in the eastern Rockies and Foothills south of Bow River. The Spray River of southeastern British Columbia, attaining a maximum of about 1,500 feet, is mostly assigned to the Sulphur Mountain Formation which is divided into a lower recessive shaly siltstone, a blocky brown-weathering unit, and an upper dolomitic siltstone unit. Overlying light grey to buff weathering limestones and calcareous sandstone, 20 to 40 feet thick, are included in the Whitehorse Formation, which is restricted mainly to the Elk River and Wigwam Creek areas.

The same formations are recognized north of Bow River. The Sulphur Mountain Formation consists of dark grey-brown to rusty brown weathering siltstones, silty dolomites, shales, and fine-grained sandstone, and the Whitehorse Formation is a light grey to buff weathering assemblage of limestones, sandy limestones, dolostones, intraformational breccias, and evaporites. Between Bow

and Smoky Rivers, the Sulphur Mountain is divisible into four members (Gibson, 1968). The Phroso Siltstone Member comprises 150 to 652 feet of shaly siltstones and silty shales with minor intercalations of dolomitic siltstone and sandstone, and contains *Claraia stachei* of Griesbachian age. Overlying dolomitic siltstone and shale, 118 to 442 feet thick with *Euflemingites* of Smithian age, are included in the Vega Siltstone Member. The Whistler Member of Anisian age is composed of recessive dark grey weathering dolomite with phosphatic pebble-conglomerate, 12 to 140 feet thick. The Llama Member of silty to sandy dolomite and siltstone, 9 to 145 feet thick, is also Anisian. Two members are recognized within the Whitehorse Formation between North Saskatchewan and Athabasca Rivers. The lower Starlight Evaporitic Member with a maximum thickness of 1,320 feet, contains buff to yellow limestone, collapse breccias, and gypsum. A prominent unit of well-indurated quartzitic sandstone, as much as 256 feet thick, occurs in the upper part in that region. The upper Winnifred Member contains sandy dolomite and limestone. Between Athabasca and Smoky Rivers, 11 to 205 feet of pelletoid fossiliferous limestone, lying between the Starlight and Winnifred Members, is termed the Brewster Limestone Member. In the mountain north of Smoky River, the lower Triassic shales and siltstones are overlain by sandstone and dolomite, and the upper part contains massive to shaly limestone. A remarkable fauna of large fossil fish is known from beds of Lower Triassic age at Wapiti Lake and has also been reported north of Peace River.

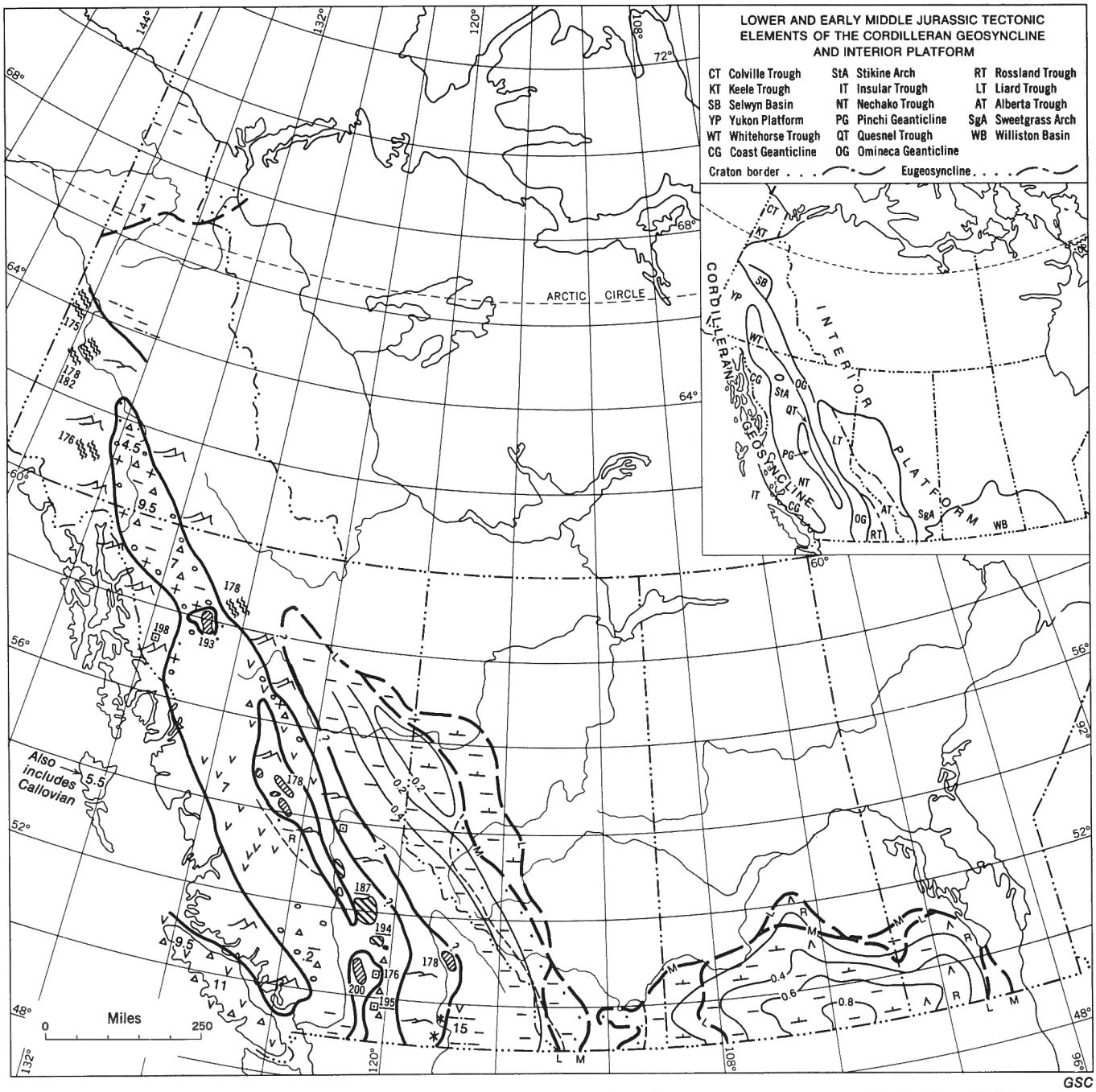
Sediments of the Sulphur Mountain Formation are postulated to have been laid down during minor marine transgressions and regressions in a deltaic-type environment. In contrast, sediments of the Whitehorse Formation were probably deposited in a more restricted shallow marine environment, with the occurrence of gypsum, collapse breccias, and red beds indicating deposition under subaerial and highly saline, evaporitic environments.

In the vicinity of Peace River (McLearn and Kindle, 1950), much of the lower part of the Triassic succession is not exposed, although several hundred feet of interbedded shale and siltstone are known near Halfway River. At Peace River, the lowest exposed beds, representing the Liard Formation, comprise the thinly bedded, calcareous, dark siltstones, 75 to 430 feet thick, with the Ladinian Meginae Zone. The succeeding Grey Beds, consisting of more than 2,000 feet of massive quartzitic sandstones, siltstones, reddish fine clastic sediments, limestones, anhydrite, dolomite, and breccias, represent depositional environments varying from marine to near-shore and evaporitic conditions. The Grey Beds contain a *Nathorstites* fauna (Sutherlandi Zone) near the base, the early Late Triassic Mahaffy Cliffs and Red Rock Spur fauna near the middle, and *Lima? poyana* fauna toward the top. The carbonate rocks of the upper Grey Beds and the Pardonet Formation were deposited during regional subsidence and renewed marine transgression in late Karnian and early Norian time. The Upper Triassic Pardonet beds of dark calcareous siltstones and shales with some dark limestones, although only 200 to 700 feet thick, contain a remarkably complete sequence of Upper Karnian and Norian ammonoid zones. They consist mainly of dark calcareous siltstone and shale with some dark coquinoid limestone.

Between the Liard and Halfway Rivers, the earliest Triassic seas transgressed slightly bevelled Paleozoic rocks. The Grayling Formation, at the base of the succession, consists of 350 to 1,500 feet of dark non-calcareous shales, minor limestones, with some sandstone, and contains *Claraia stachei* (Griesbachian). It is recognized only between Liard and Sikanni Chief Rivers, apparently grading southward into a siltstone facies generally included with the Toad Formation. The latter, 600 to 3,000 feet thick, comprises grey calcareous siltstones, calcareous shales, minor limestone, and sandstone with Dienerian, Smithian, Spathian, and Anisian ammonoid faunas (Pelletier, 1963).

Increased uplift during Middle Triassic resulted in an increased influx of sand and the deposition of the Liard Formation, which comprises as much as 600 feet of grey calcareous, crossbedded quartzitic sandstone and dark grey siltstone. The succeeding beds, with a maximum thickness of more than 2,000 feet, include fine- to coarse-grained sandstones, light grey to yellow limestone and dolomite, intraformational breccia, and bioclastic, bituminous limestones. Recent studies by Pelletier suggest that the dark calcareous siltstones were deposited in the most northwesterly region more or less continuously from Smithian (Lower Triassic) to Norian (latest Triassic) time, representing a coalescence of Toad and Pardonet facies without any intervening sandstone of Liard or Grey Beds type.

Triassic beds beneath the Peace River Plains have been placed (Hunt and Ratcliffe, 1959; Armitage, 1962) in the Montney and Doig Formations of the Daiber Group and in the Halfway, Charlie Lake, Baldonnel, and Pardonet Formations of the Schooler Creek Group. The Montney, similar to the Toad of the Foothills, grades upward from dark shale into argillaceous siltstones. The Doig, consisting of bituminous dolomitic siltstone with nodular phosphate at the base, may be more than 400 feet thick in the west. Its lower beds have been correlated with phosphatic beds of Anisian age in the Sulphur Mountain and Toad Formations. The Halfway Formation, containing well-sorted, fine- to medium-grained sandstone has been interpreted as near-shore sandbar deposits (Fig. IX-27). Its maximum thickness is more than 400 feet, decreasing eastward to a zero depositional edge. An interval of stability is indicated by the near-shore and evaporitic deposits of the overlying Charlie Lake Formation. The latter, ranging from 0 to more than 1,400 feet, consists of massive anhydrite, red dolomitic siltstone, dolomite, and minor salt, and contains the important oil-producing Boundary Lake Member of dolomitized skeletal limestone. The Charlie Lake beds grade laterally westward into sandstones and limestones similar to those of



DEPOSITIONAL FEATURES

- Conglomerate
- Granitic clasts
- Greywacke
- Sandstone
- Siltstone
- Shale
- Limestone
- Anhydrite
- Volcanic rocks
- Red beds
- Depositional limit
- Truncated limit
- Lower Jurassic
- Middle Jurassic
- Facies boundary
- Isopach (thousands of feet)
- Thickness (thousands of feet)
- Volcanic center

INKLINIAN OROGEN

- Granitic plutons (age in millions of years)
- Granitic plutons, under Jurassic deposits (age in millions of years)
- Metamorphic trends (age in millions of years)

PHYSIOGRAPHIC FEATURES

- Mountainous land (high, low)

Blank areas are lowlands and plains developed on crystalline and sedimentary rocks of the craton, and where the Jurassic is now absent through erosion, or unknown. In the geosyncline the blank areas may be sea.

FIGURE VIII-33. Lower and early Middle Jurassic sedimentation and tectonism in Western Canada, and some features of the Lower Jurassic Inklinian Orogeny.

the Halfway and are probably represented in the Foothills by part of the Grey Beds and by evaporites of the Whitehorse Formation farther south. The Baldonnel Formation, as much as 300 feet thick and composed of dolomitized skeletal limestone, is the major gas reservoir. The Baldonnel is generally correlated with microcrystalline and fragmental limestone of the upper unit of the Grey Beds, although it is considered by some to be equivalent in part to the Pardonet Formation. The Pardonet is best known in the Foothills, as it was largely removed from more easterly areas by post-Triassic erosion.

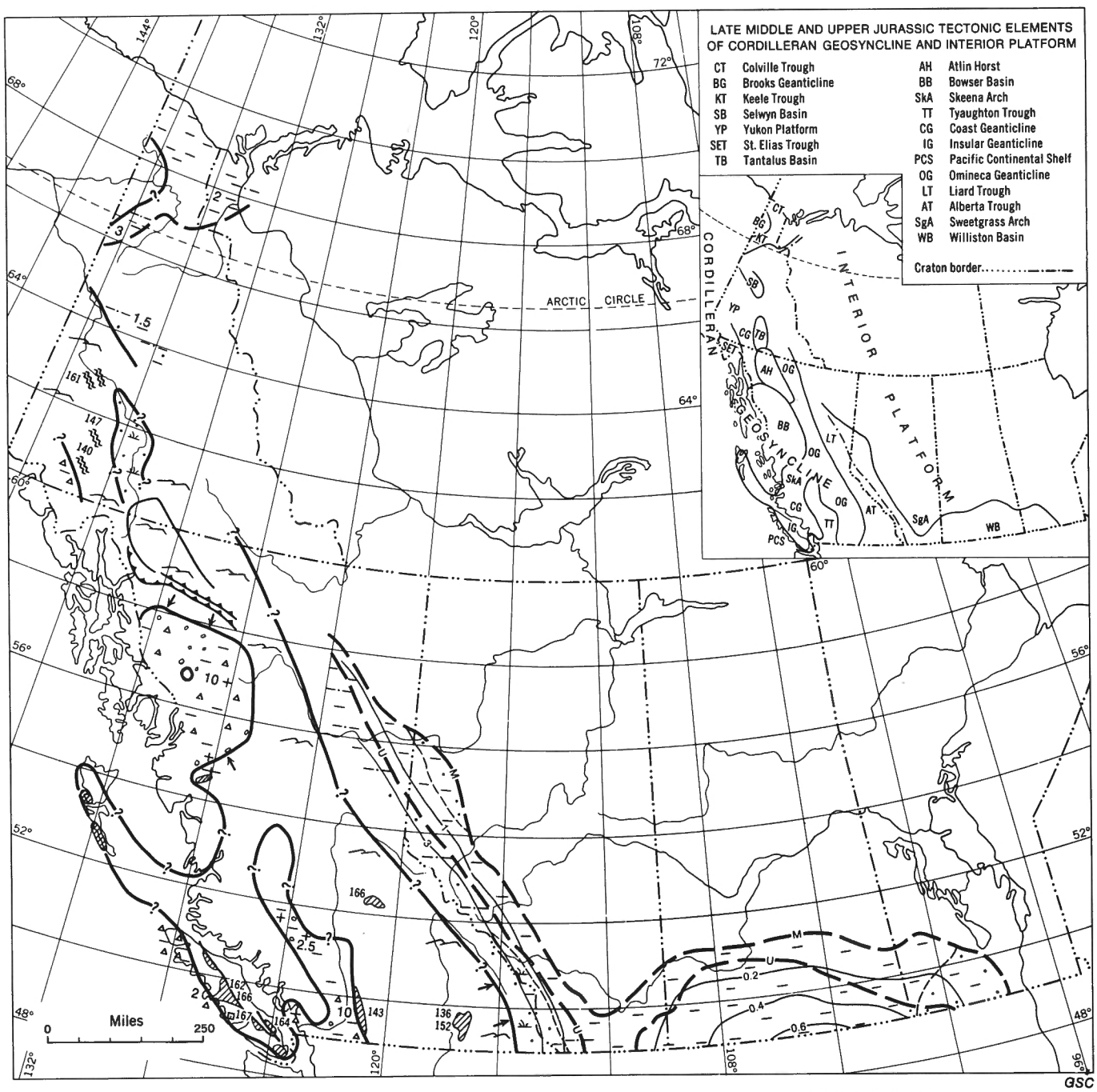
The origin and correlation of the Halfway Formation is controversial. Although the Halfway, Liard, and lower Grey Beds contain similar near-shore sand deposits, neither their precise relationships nor their depositional history has been clearly established. Some attribute the Halfway to regression; others maintain that it resulted from widespread transgression and that it lies unconformably on earlier Triassic beds. The development of the Liard and Halfway sandbars and their gradation westward into a silty to shaly facies also suggest shorelines in the east (Pelletier, 1963). The northward increase in sand ratios and the direction of sediment transport deduced from primary sedimentary structures point to a northeasterly or northerly positive cratonic element, probably in southeast Yukon Territory or District of Mackenzie, as a source for some, at least, of the Triassic sediments in Liard Trough.

Jurassic

Tectonic Summary

The Jurassic System lies conformably on the Triassic in parts of Insular, Nechako, and Whitehorse Troughs and unconformably and disconformably on Triassic and older strata, or plutonic rocks, around the margins of the Coast, Pinchi, and Omineca Geanticlines as a result of uplift during the Inklinian Orogeny, and a general epeirogenic uplift of the craton (Fig. VIII-33). The orogeny is represented by emplacement of granitic plutons in Stikine Arch and the Pinchi and Omineca Geanticlines, and by regional metamorphism in Yukon Platform. These elements were progressively reduced in relief and adjoining troughs subsided strongly during the remainder of the Lower and early Middle Jurassic. Coarse conglomerates containing granitic, volcanic, and some metamorphic debris flank the uplifts. In the non-volcanic Whitehorse Trough the conglomerates grade into thick turbidites but in Quesnel and Nechako Troughs they are associated with abundant pyroclastic and volcanic clastic rocks. In western Rossland Trough andesitic and basaltic flows grade eastward into shales and siltstone. Possibly some ultramafic rocks were emplaced into the volcanic rocks in the Lower Jurassic. In the early Jurassic thick porphyritic pyroclastic andesite accumulated in southern Insular Trough. The northern part was stable and received limestone and shale, succeeded in early Middle Jurassic time by explosive, partly subaerial volcanic rocks. In Colville Trough of northern Yukon Territory, thin black shale was deposited with sandstone and siltstone flanking Aklavik Arch. These sediments represent an embayment of the Arctic Sea that lacked direct connections with seas in the south. In Alberta Trough, subsidence was slight and sedimentation was interrupted only by minor regressions. Thin euxenic shale, siltstone, and limestone were possibly contiguous across Purcell Arch with similar sediments in Rossland Trough. Along the margin of the craton the shales overlap eroded Triassic and upper Paleozoic strata and contain interbedded sandstone derived from Sweetgrass Arch. Lower Jurassic marginal red sediments were derived from the craton and grade into evaporites in the centre of Williston Basin, which was strongly autogeosynclinal throughout the Jurassic. The basin was connected with the geosyncline by a narrow seaway in the northern United States, more open circulation being established in the Middle Jurassic as southern Sweetgrass Arch was depressed.

Rocks of the Middle Jurassic Bathonian stage are unknown in the western Cordilleran Geosyncline, apparently the result of a general regression accompanying Nassian Orogeny. Granitic rocks were intruded along Skeena Arch transverse to Nechako and Whitehorse Troughs. Southern Whitehorse and northern Nechako Troughs were apparently deformed as the Upper Jurassic lies unconformably on Permian to Lower Jurassic strata. Faults bordering Atlin Horst and associated, southwest-directed thrusts and folds were probably produced at this time. Following Nassian Orogeny three discrete depositional basins developed, the Tantalus and Bowser Basins and Tyaughton Trough, and much of the older orogenic lands were broadened as positive elements, some under active erosion (Fig. VIII-34). Much of Insular Trough was emergent, thick clastic sediments being deposited along the western margin of the Insular Geanticline. Tyaughton Trough received considerable granitic debris from terrain along the western margin of the Pinchi Geanticline, now merged with the Omineca Geanticline. Bowser Basin received detritus, especially from Atlin Horst and Skeena Arch, but the sea was connected across a depressed central part of the Coast Geanticline with that of northern Insular Trough. Tantalus Basin was a landlocked basin in which siliceous clastic sediments and coal-forming plants accumulated. Deep, narrow foredeeps filled with intercalated sandstone and shale developed in Colville Trough bordering the uplifted Brooks Geanticline, which in Alaska was the site of orogenic disturbances and intrusion of granitic rocks. The effects of Nassian Orogeny are not evident in Alberta Trough, the subsidence being slight and intermittent. Minor hiatuses within late Middle Jurassic shales are marked by glauconitic shoreline sands derived from the east. Accompanying a general depression of the craton in the late Jurassic, shale and siltstone derived from the west was spread throughout Alberta Trough, across



DEPOSITIONAL FEATURES

- Conglomerate
- Granitic clasts
- Greywacke
- Sandstone
- Siltstone
- Shale
- Coal
- Depositional limit
- Truncated limit
- Late Middle Jurassic (M)
- Upper Jurassic (U)
- Facies boundary
- Isopach (thousands of feet)
- Thickness (thousands of feet)
- Direction of sediment transport

NASSIAN OROGEN

- Granitic pluton (age in millions of years)
- Metamorphic trends
- Fault
- Thrust fault
- Structural trend

PHYSIOGRAPHIC FEATURES

- Mountainous land

Blank areas are lowlands on crystalline and sedimentary rocks of the craton, and where the Jurassic is now absent through erosion, or unknown. In the geosyncline the blank areas may be sea.

FIGURE VIII-34. Late Middle and Upper Jurassic sedimentation and tectonism in Western Canada, and some features of the Middle Jurassic Nassian Orogeny.

Sweetgrass Arch, and into Williston Basin. In the latest Jurassic, non-marine environments prevailed in southern Alberta Trough as it was filled with sands derived from the uplifted southern Omineca Geanticline during the initial phase of the Columbian Orogeny.

Inklinian Orogeny

A period of regional uplift, granitic intrusion, and locally, regional metamorphism, termed the Inklinian Orogeny, occurred in the Pinchi, Omineca, and northern Coast Geanticlines in latest Triassic and earliest Jurassic time. Lower Jurassic strata, commonly Sinemurian, occur along the eastern flank of the northern Coast Geanticline and around Stikine Arch, resting unconformably on beds of Middle and Upper Triassic age. Along the flanks of the Pinchi and Omineca Geanticlines they lie on Upper Triassic, Permian, Pennsylvanian, and Cambrian formations or on granitic plutons. The Lower Jurassic conglomerates commonly carry granodiorite, diorite, and quartz diorite debris which is particularly abundant in the thick wedges bordering the northern Coast Geanticline and Stikine Arch where uplift was greatest (Fig. VIII-33).

Several oval, biotite–hornblende granodiorite plutons in Stikine Arch and in the Pinchi and Omineca Geanticlines intrude Upper Triassic and older rocks that have undergone little or no metamorphism. They were rapidly unroofed and contributed debris to the Lower Jurassic conglomerates. Some of these plutons (Fig. VIII-28) and their oldest K–Ar dates are the Hotailuh (193 m.y.), Galore Creek (198 m.y.), Guichon (200 m.y.), Thuya (194 m.y.), and Takomkane (187 m.y.) batholiths and Copper Mountain stock (195 m.y.). Other plutons that may be partly of the same age are the granodiorite and pink granite phases of the Topley Intrusions and the aegirine–augite monzonite of the Kuskanax batholith. Both these plutons have yielded K–Ar dates up to 178 m.y. and the Topley pink granites occur as debris in Middle Jurassic sediments. K–Ar ages of about 180 m.y. from schists and gneisses of the Yukon Group near the Alaskan Boundary suggest that regional metamorphism took place in the Yukon Crystalline Platform during the early Jurassic, but no evidence of deformation at this time has been found elsewhere.

Lower and Early Middle Jurassic of Western Cordilleran Geosyncline

Whitehorse Trough. The Laberge Group of Early and early Middle Jurassic age underlies much of south-central Yukon Territory (Wheeler, 1961) and part of northwestern British Columbia northwest of Bowser Basin and south and west of the Atlin Horst (Souther and Armstrong, 1966). It lies upon the Upper Triassic with a disconformity that increases to the southwest. The Laberge Group consists of three facies: a southwestern, coarse clastic facies represented in British Columbia southwest of King Salmon thrust by the Takwahoni Formation, a central turbidite facies represented northeast of the thrust by the Inklin Formation; and a local northeastern conglomeratic facies along Teslin River in Yukon Territory (Figs. VIII-31b, 32a).



PLATE VIII-10. Conglomerate of Lower Jurassic Laberge Group with well-rounded granite cobbles, southwest of Whitehorse, Yukon Territory.

The southwestern facies is interbedded conglomerate, greywacke, siltstone, and shale, more than 11,000 feet thick, and characterized by rapid facies changes, channelling, and local disconformities. The conglomerates are prevalent in the lower part of the group. Those in the Takwahoni Formation are associated with mid-Pliensbachian and early Toarcian ammonites whereas farther northwest conglomerates occur with early Sinemurian and early Toarcian ammonites. The conglomerates, which at several localities carry boulders nearly 4 feet across, must have been derived from sources nearby. They are composed essentially of clasts of volcanic rock, which predominate in the lower parts of the conglomerate sections or granitic rocks, which are most abundant in the upper sections (Pl. VIII-10). Metamorphic detritus is rare. The greywackes contain considerable quartz and potash feldspar (Pl. VIII-11). The conglomerates are most abundant to the southwest and individual beds, which are commonly lenticular, thicken rapidly in that direction suggesting that they were derived from granite-cored highlands in the Coast Geanticline that were uplifted at intervals during the early Jurassic. Known sources of granitic rocks are: the granodiorite and quartz diorite of the Coast Plutonic Complex southwest and northwest of Whitehorse, Yukon Territory; mid-Triassic diorites in Stikine Arch; and the post-Karnian, pink hornblende granodiorite of the Hotailuh batholith (Gabrielse and Reesor, 1964). Siltstones and shales are abundant in the early Middle Jurassic, upper part of the Laberge Group and suggest that by this time the highlands had been reduced to regions of low relief.

In Yukon Territory, the coarse clastic facies of the Laberge Group grades eastward into the turbidite facies

of the Inklin Formation which is more than 9,000 feet thick. It is featured by graded-bedded greywacke, siltstone, and shale displaying slump structures, convoluted bedding, and intraformational breccia with but a few fossils and no ripple-marks. The sediments were deposited in the rapidly subsiding central part of Whitehorse Trough, presumably from turbidity currents that originated as slumps off the rapidly deposited deltas along the southwest margin of the trough.

The northeastern facies is characterized by a tabular sheet of conglomerate at the base of the Laberge Group that is extremely coarse locally, carrying 12-foot blocks of limestone, and therefore must have been derived from the east, from the Omineca Geanticline. The latter was uplifted only in earliest Jurassic time and considerably less than the Coast Geanticline.

The fauna from the Laberge Group (Frebold, 1964) lacks about eight ammonite zones, indicating two gaps possibly caused by non-deposition. One spans the upper Sinemurian and most of the lower Pliensbachian and the other is in the upper part of the Pliensbachian. The gaps occur elsewhere in western Canada and Frebold considered that they may represent periods of general regression. Locally, as in Rossland Trough, they coincide with times of strong volcanic activity, and in other areas with times of uplift.

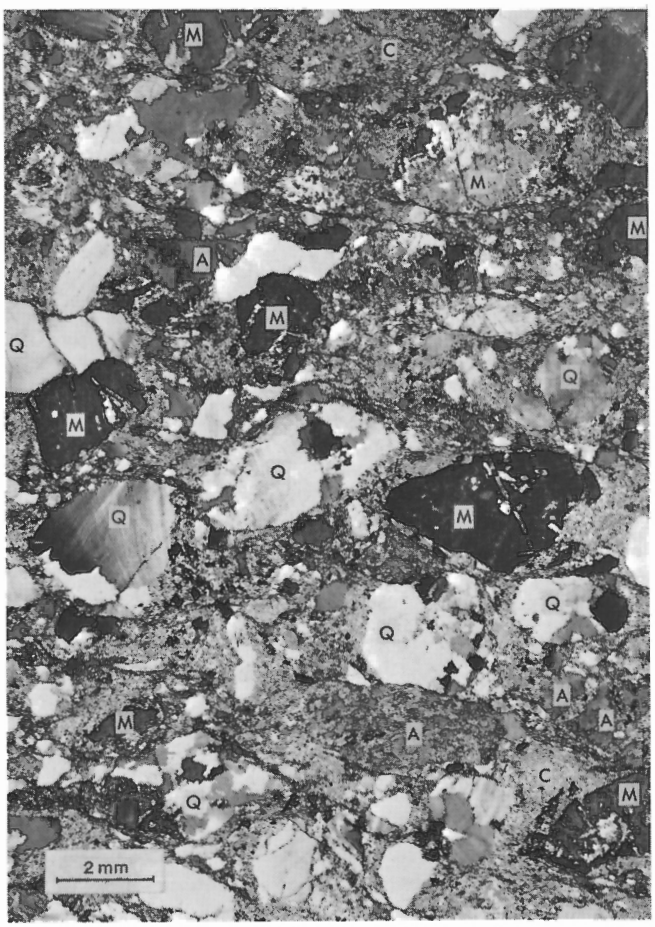


PLATE VIII-11. Photomicrograph of Lower Jurassic microcline-bearing greywacke from northern Cassiar Mountains, British Columbia. Q—quartz; M—microcline (stained); A—altered plagioclase; C—chert.

Nechako Trough. The unfossiliferous red beds of the Takla Group, derived from the erosion of the rising and expanding Pinchi Geanticline to the northeast, are overlain disconformably by the chert-pebble conglomerate unit at the base of the Hazelton Group and may therefore be partly Lower Jurassic (Fig. VIII-32c). The red beds apparently grade southwestward into marine volcanic rocks deposited in the central part of the Nechako Trough (Tipper, 1959). The conglomerate unit, which is both marine and non-marine, contains andesite and basalt flows and tuffs not unlike those in the Takla Group and also quartzite, shale, and siliceous speckled greywacke. The conglomerate clasts, averaging about an inch in diameter but locally as much as 3 feet, are quartz, chert, minor greenstone and argillite, and rare metamorphic and pink granitic rocks. The latter are typical of the nearby Topley Intrusions, which are composed of diorite and pink granitic phases, and cut the lower part of the Takla red beds thus indicating granitic emplacement during uplift and expansion of the Pinchi Geanticline in the early Jurassic. The conglomerate unit is overlain by the main unit of varicoloured andesitic and rhyolitic flows, breccia, and tuff and arkose, carrying early Middle Bajocian fossils.

Near Coast Mountains, the Hazelton Group is more than 11,000 feet thick and comprises two volcanic divisions separated by a mainly marine sedimentary division, all of which were deposited in the deeply subsiding Nechako Trough. The volcanic rocks are fine breccias or coarse tuffs with subordinate flows of red and green andesite and basalt with some rhyolite and dacite. No volcanic centres have been recognized although east of the Coast Mountains near Skeena River breccias contain fragments 2 feet or more in diameter (Duffell and Souther, 1964). Elsewhere along the east side of the Coast Mountains the breccias contain fragments of granitic rocks and metamorphic equivalents of the Hazelton Group lie with angular unconformity on a complex of greenstone and older plutonic rocks (Stuart, 1960). Near Bella Coola greenstone containing Bajocian or older fossils lies unconformably on foliated saussuritized granodiorite of the Coast Plutonic Complex (Baer, 1967). It is apparent, therefore, that the western margin of the volcanic Nechako Trough in late Lower or early Middle Jurassic time lay just west of the eastern edge of the present Coast Plutonic Complex.

Near Taseko Lakes, clastic sediments have yielded Norian, Rhaetian, and Hettangian fossils. This is the only region in the Cordilleran Geosyncline where the Jurassic can be demonstrated to lie conformably on the Triassic. The late Lower and early Middle Jurassic rocks are mainly shales, which implies remote or subdued source lands. North of the lakes a conglomerate, probably Lower Jurassic, contains boulders of Upper Triassic, Upper Paleozoic, and Devonian rocks. In the Cascade Mountains,

the Lower Jurassic is represented by shales of the Cultus Formation and the Middle Jurassic by basaltic pillow lava, breccia, and tuff. In eastern Coast Mountains, the lower part of the Dewdney Creek Group contains Lower and Middle Jurassic volcanic greywacke and a few flows (Coates, 1967).

Quesnel Trough. The predominantly volcanic Takla Group, deposited in the northern part of Quesnel Trough, contains Lower Jurassic intercalated sediments, with green andesitic and basaltic flows, breccias, and tuffs. In the southern part of the trough, near Prince George, the volcanic rocks are similar but contain a purple analcite-bearing unit. The clastic rocks are derived mainly from the erosion of volcanic rocks. Some are metamorphosed to zeolite facies.

Coarse, clastic, Lower Jurassic rocks with granitic debris from recognizable sources occur along the edge of Quesnel Trough and lie unconformably on the Triassic, indicating uplift of the Omineca and Pinchi Geanticlines in the earliest Jurassic. The latter geanticline perhaps was also uplifted in the Middle Jurassic (Fig. VIII-32c, d). In southern Omineca Mountains *Harpoceras* is associated with coal-bearing and conglomeratic strata (Armstrong, 1949), presumably deposited along the margin of the Omineca Geanticline. Southeast of Prince George a conglomerate, probably Sinemurian, contains boulders of granite similar to that cutting Karnian beds (H. W. Tipper). At the north end of the belt of Cache Creek rocks, lying *en echelon* southeast of the Pinchi Geanticline, Sinemurian conglomerate contains boulders of older phases of Thuya batholith that intrudes Karnian rocks around Bonaparte Lake. Lower Jurassic rocks south of Quesnel contain boulders of granite similar to those exposed nearby in the Pinchi Geanticline and other Jurassic conglomerates contain granitic and Cache Creek debris (Campbell, 1961). At the south end of the Pinchi Geanticline, near Ashcroft, conglomeratic beds carrying granitic debris and associated with strata yielding Bajocian fossils (Crickmay, 1930) lie unconformably on the Guichon batholith, which intrudes the Karnian (Duffell and McTaggart, 1952). The upper part, above an upper granite-bearing conglomerate, contains Callovian fossils. Crickmay considered the succession conformable but elsewhere Callovian strata are separated from Bajocian and older beds by an unconformity reflecting the Nassian Orogeny. If an hiatus exists in this section it may be beneath the upper conglomerate.

Rossland Trough. The Rossland Group of andesite and basalt flows southwest of Kootenay Lake was deposited in western Rossland Trough (Fig. VIII-32e). The group is composed almost entirely of volcanic rocks in the west, containing Sinemurian fossils near the base and unconformably overlying the Pennsylvanian Mount Roberts Formation (Little, 1960). Northeastward, the lower part of the Lower Rossland Group intertongues with clastic sediments of the Sinemurian and (?)Hettangian Archibald Formation whereas higher beds become the Elise Formation. Volcanic clastic beds near the top of the Archibald Formation contain granitic pebbles. The Archibald presumably grades northeastward into the unfossiliferous Ymir Group that was probably deposited close to the axis of Rossland Trough. It appears that the Rossland lavas were extruded near the flexure between Rossland Trough and the southernmost Omineca Geanticline, which was uplifted in earliest Jurassic or latest Triassic time. The Lower Rossland Group is overlain by the Hall Formation which carries faunas ranging in age from early Toarcian near the base to Middle Bajocian or younger near the top (Frebold and Little, 1962). The Hall Formation is chiefly argillite, but contains siltstone, greywacke, and conglomerate towards the north and is overlain conformably by volcanics and sediments of the Upper Rossland Group.

Insular Trough. On Vancouver Island the Lower Jurassic is represented by explosive rocks of the Bonanza Formation ranging in age from late Karnian to Toarcian. They consist of porphyritic andesitic agglomerates and tuffs. Sequences 6,000 feet thick accumulated in some places. In northwestern Vancouver Island considerable quantities of volcanic arenite and argillite were deposited. No early Middle Jurassic (Bajocian) beds have been recognized on Vancouver Island. On Queen Charlotte Islands the Lower Jurassic is represented by the Upper Kunga shales and the Maude shale and sandstone, conformably succeeded by the Middle Jurassic Yakoun Formation. The Yakoun is composed of explosive volcanics as much as 6,000 feet thick and contains Bajocian and Callovian fossils. The basal beds are scoriaceous lapilli tuffs whereas higher in the section slightly sodic, andesitic, block agglomerates are interbedded with clastic sediments. Sutherland Brown (1966) considers that a chain of volcanoes formed along the east coast of present Queen Charlotte Islands, flanked on the west by a shallow marine basin in which coal accumulated, intercalated with volcanic arenite and tuff.

Nassian Orogeny

The Nassian Orogeny during mid-Jurassic (Bathonian) time resulted in granitic intrusion on Vancouver Island and regional uplift of the western Cordillera. Uplift of Atlin Horst and Skeena Arch segmented the Whitehorse and Nechako Troughs into Tantalus and Bowser Basins and Tyaughton Trough (Fig. VIII-34). The major southwesterly directed thrust faults and related structures on the southwest side of Atlin Horst were probably produced at this time. Souther and Armstrong (1966) suggest that the Lower Jurassic turbidite facies and associated rocks glided southwesterly off the rising horst as a gravity slide, the King Salmon thrust. In the Skeena Arch the Hazelton and older rocks were folded along northeasterly trends and were probably intruded by granitic plutons prior to the deposition of the Bowser Group in the earliest Upper Jurassic. This is suggested by intrusion of the Hazelton Group by the early, more easterly phases of the Coast

Plutonic Complex and by the presence of debris not unlike these early phases in the southwesternmost exposures of the Bowser Group, itself cut by still younger granitic intrusions.

Some deformation and plutonism also took place in Insular Trough. This is suggested by an unconformity on the west coast of Vancouver Island between Toarcian and Callovian beds, the presence of granitic clasts in the conglomerates, and the association of steep northwest-trending faults with northwesterly elongate, syntectonic and late syntectonic plutons intrusive into Lower Jurassic and older rocks. Several K–Ar dates on the granodiorite plutons range from 151 m.y. to 167 m.y. The deformation was mild, apparently consisting of the development of broad, northwesterly trending folds, in contrast to northerly trends within late Paleozoic rocks, and fracturing into elongate blocks along northwest-trending faults.

Elsewhere in the eugeosynclinal belt Middle Jurassic orogenic movements are more obscure but may be represented by the following features: absence of Bathonian fossils; an unconformity between early and late Middle Jurassic beds along the east side of the southern Coast Geanticline; the late Middle Jurassic granite-bearing conglomerates resting unconformably upon the Inklinian Guichon batholith; and a K–Ar date of 166 m.y. on the granite phase of the Thuya batholith.

Upper Jurassic of Western Cordilleran Geosyncline

Tantalus Basin. The Tantalus Formation in south-central Yukon Territory comprises quartz- and chert-bearing conglomerate and sandstone, shale, and coal apparently derived from siliceous rocks in the Coast and Omineca Geanticlines (Fig. VIII-31c). The sediments were deposited under non-marine conditions if one judges from the presence of coal and fossil plants of Neocomian or possible Portlandian age, and the lack of marine fossils. The Tantalus Formation lies apparently conformably on the Laberge Group in the north-central part of the basin near Carmacks but lies unconformably on siliceous metamorphic rocks in its southwesternmost outcrops and disconformably on mid-Lower Jurassic (Pliensbachian) in its southernmost exposures (Wheeler, 1961).

Bowser Basin. The Bowser Group is a thick succession of marine and non-marine shales, greywacke, and conglomerate that was deposited in the Bowser Basin. The group includes, in the east, the Oxfordian strata of the upper division of the Takla Group (Lord, 1948) and the Groundhog coal measures. Chert-pebble conglomerate is present mainly in the northwestern and northern parts of the basin. It occurs as huge, wedge-shaped channel-fills that apparently represent streams carrying Cache Creek chert debris derived from Atlin Horst and Stikine Arch (Souther and Armstrong, 1966). Conglomerate in the southwest occurs in Oxfordian shales and is composed of clasts, as large as a foot in diameter, of volcanic rocks from the Hazelton Group, or chert and granite. The granitic debris is not unlike some of the older phases of the Coast Plutonic Complex that cut Hazelton strata in Skeena Arch. The Bowser Group lies unconformably upon the Hazelton Group in Skeena Arch and on older Jurassic rocks in Stikine Arch. It appears to lie unconformably upon Permian strata in the Oweegee Mountains in the west-central part of Bowser Basin.

Tyaughton Trough. South of Skeena Arch and east of the Coast Mountains the Relay Mountain Group occurs in a fault-bounded belt along Fraser River and extends into the Cascade Mountains. The group in the northern part of the Tyaughton Trough (Jeletzky and Tipper, 1968) is composed of an essentially conformable sequence of clastic sediments that range in age from late Callovian to Barremian and contains fossils representing all intervening stages. The base of the group is not exposed. Shales and siltstones occupy the central part of the trough but grade laterally to the southwest and northeast into greywacke and locally conglomerate. Granitic detritus, derived from masses exposed to the northeast, is restricted to the northeastern exposures of the group and first appears in Upper Oxfordian beds carrying *Buchia concentrica.* The latter may lie disconformably on Lower Oxfordian shales with *Cardioceras.*

In eastern Cascade Mountains the lower part of the Ladner Group is mainly a slate containing belemnites. Around Harrison Lake, Callovian beds of the Mysterious Creek Formation are locally overlain disconformably by 3,000 feet of early Oxfordian conglomerate. The clasts are argillite, chert, quartzite, tuff, agglomerate, and the Middle Jurassic Harrison Lake porphyries and micropegmatite. Elsewhere early Lower Cretaceous beds lie disconformably on the Billhook Formation which conformably overlies the Mysterious Creek Formation. The southwestern flank of Tyaughton Trough accordingly appears to have been subjected to periodic uplift during the Late Jurassic. In southwest Cascade Mountains, the Nooksack Group is mainly slate (J. W. H. Monger), flysch-like in character, and commonly contains sand- or silt-sized volcanic clastic rocks or locally, in its uppermost part, beds of granite-bearing conglomerate.

Pacific Continental Shelf. Late Middle and Upper Jurassic clastic sediments are found only locally on the west coast of Vancouver Island near Esperanza Inlet. The lower unit, Division A of Jeletzky (1950), ranges in age from mid-Callovian to (?)early Kimmeridgian. It consists mainly of tuffaceous mudstone, siltstone, and greywacke with minor pebble-conglomerate which at one locality contains pebbles of granitic rocks and lies with angular unconformity on the Bonanza Group. The upper part, of Oxfordian age, is mainly a succession of greywacke beds, but the top is not exposed. The upper unit, Division B, not in contact with Division A, comprises Portlandian shale and sandstone. It is overlain disconformably by the early Lower Cretaceous One Tree Formation.

Jurassic of Eastern Cordilleran Geosyncline and Interior Platform

Colville Trough. In Aklavik Range of northern Richardson Mountains (Jeletzky, 1967), the Bug Creek Formation, late Sinemurian to possibly early Oxfordian in age, consists of about 600 to 800 feet of fine-grained quartzose sandstones and siltstones, probably derived from the east. The Upper Jurassic is represented by Husky shale, a succession some 1,300 feet thick of dark to brownish grey concretionary shale and siltstone with minor fine or pebbly sandstone and lenses of *Buchia* coquina. The uppermost 190 feet are lowermost Cretaceous and are transitionally overlain by Valanginian sandstones. Near Vittrekwa River the Husky grades southward into the North Branch conglomerate, 600 feet thick, a shoreline deposit on the northern flank of Aklavik Arch.

The Jurassic of British Mountains is not well known, but concretionary grey and black shales are similar to the Kingak shale of Alaska which is 4,000 feet thick. The shale lies disconformably on the Triassic and Carboniferous but unconformably on the Neruokpuk Formation. In southern Keele Range, 2,000 feet of black shale is succeeded by 2,400 feet of siltstone, shale, and quartzose sandstone, the uppermost part of which is of early Lower Cretaceous age (Frebold, *et al.,* 1967).

Selwyn Basin. At least 1,000 feet of black shale, slate, and siltstone containing Middle Bathonian, Lower Callovian, and Lower Oxfordian ammonites occurs in the hanging-wall of a major thrust fault in the Tombstone–Tintina region (Tempelman-Kluit, 1966).

Alberta and Liard Troughs. In the Foothills and Rocky Mountains of southeastern British Columbia and southwestern Alberta, the Jurassic is represented by the Fernie Formation. It is about 600 to 1,000 feet thick, and lies unconformably on eroded Triassic or Paleozoic rocks. In the Fernie area (Frebold, 1957), basal Jurassic beds containing the *"Arnioceras"* fauna consist of a few feet of black pelletoidal and nodular phosphate rock and shale overlain by at least 100 feet of black shales. Farther east, in the vicinity of Blairmore, the transgressive basal deposits are Toarcian sandstone and shales. The overlying Bajocian Rock Creek Member consists mainly of rusty weathering dark shale with siltstone and sandstone and is characterized by belemnite conglomerates and the *Teloceras* fauna. One of the major Jurassic inundations is represented by early Callovian Grey Beds, containing the *Corbula munda* and *Gryphaea* beds. Medium to light grey shales are present near Blairmore but calcareous sandstone and sandy limestone are more abundant farther west. Glauconitic sandstone and shale, with *Cardioceras,* of the Oxfordian Green Beds grade upward into the Passage Beds of interbedded fine-grained marine sandstone, siltstone, and black shale. These upper sandy sediments were deposited during the final retreat of the Fernie sea that accompanied the influx of clastics produced during the initial phase of Columbian Orogeny.

Toward the end of the Jurassic, non-marine conditions of deposition, which were established in the Crowsnest region, are represented by the Kootenay Formation. The Kootenay increases from an erosional edge on Interior Platform to more than 3,500 feet in the Fernie Basin. At Blairmore, the formation is divided into four members (Norris, D. K., 1959). The basal Moose Mountain Member is a quartz and chert sandstone transitionally overlying the Fernie shales. It contains a Portlandian ammonite near Fernie, B.C. The other three members, Adanac, Hillcrest, and Mutz, comprise a sequence of carbonaceous mudstones, siltstones, and sandstones, with several commercial coal seams. These beds are considered to be Jurassic on the basis of spore analysis although macrofloral evidence suggests a Cretaceous age (Bell, 1956). In the westernmost exposures, the upper beds of the Kootenay Formation include more than 1,400 feet of Elk conglomerate, composed mainly of chert, quartz, and quartzite, as large as 6 inches in diameter, that was derived from the rising southern Omineca Geanticline, expanding eastward during the Columbian Orogeny.

Marine deposits of Sinemurian age include coquina and conglomerate near Banff. Between Crowsnest Pass and Bow River they are locally absent, and the initial Jurassic transgression is represented by Toarcian limestones, sandstones, and black papery shales. The sequence is lithologically similar to that of the Crowsnest area, but also includes such prominent units as the easterly derived Bajocian Rock Creek sandstone in the vicinity of Highwood River (Douglas, 1959) and the Callovian Pigeon Creek sandstone at Mount Allan (Crockford, 1949). The latter appears to be a local sandy facies of the Grey Beds. The Kootenay attains a maximum thickness of 3,400 feet in the mountains southeast of Banff but to the east is eroded and thins rapidly to zero. The lower part contains fine-grained sandstone and coal-bearing beds. The upper part includes several thick, coarse-grained, sandstone beds, locally finely conglomeratic, interbedded with silty, carbonaceous shales. Those beds, termed the Pocaterra Creek Member, are probably a facies of the more massive Elk conglomerates.

Basal Fernie beds in the central Foothills of Alberta (Douglas, 1958; Mountjoy, 1962), between Bow and Athabasca Rivers, are included in the Nordegg Member, 30 to 100 feet of black, cherty, and phosphatic dolomite or limestone with siltstones and shales. They are overlain by 10 to 20 feet of dark papery fissile shales followed by the thin Rock Creek sandstone. In some areas, the Rock Creek is represented by dark grey shales with belemnite conglomerates overlying the Paper shales and underlying the dark to yellow-brown shales of the Grey Beds. In the more northerly areas, glauconitic, silty, concretionary shale comprising the Green Beds overlaps and truncates the Grey Beds in a northeasterly direction. This is probably the result of uplift of the margin of the craton at the end of the Middle Jurassic. At the top of the group, intercalated thin siltstones and silty shales grade into the

overlying Nikanassin Formation. The latter, about 1,200 feet thick, consists of thick beds of fine- to medium-grained, quartzose sandstone interbedded with dark grey shale, thin siltstone, and a few thin coal seams. It occupies a stratigraphic position similar to that of the Kootenay Formation but lacks commercial coal. Late Jurassic *Buchia* occur near the base of the formation. The Nikanassin increases from 2,000 feet near Jasper to nearly 4,000 feet in western exposures near Smoky River. Similar strata, more than 6,000 feet thick north of Smoky River (Zeigler and Pocock, 1960), are included in the Minnes Formation, reported to contain Jurassic microfossils near the base and early Lower Cretaceous spores in the upper part.

In the Peace River Foothills (Stott, 1967), the basal Fernie beds, 56 to 209 feet thick, consist of phosphatic, calcareous, and cherty shales with some limestone and siltstones, overlain by black calcareous papery shale. The middle succession of 65 to 273 feet of rusty weathering sideritic marine shale with a persistent glauconitic siltstone marker, is equated on the basis of microfauna and lithology with the Rock Creek, Grey Beds, and Green Beds of more southerly regions. Interbedded sandstone and shale intertongue with and grade upward into fine- to coarse-grained, quartzose sandstone of the Monteith Formation which spans the boundary between the Jurassic and Cretaceous.

The Jurassic of the adjacent Interior Platform is lithologically similar to the Fernie Formation (Lackie, 1958). The northeastern limit is a highly irregular erosional edge (Fig. VIII-33). Black, calcareous shales of the Nordegg Member, formed as a shelf deposit, were later covered in central Alberta by a fairly extensive chert and locally by accumulations of pebbly sandstone derived from Paleozoic beds to the northeast (Springer, *et al.*, 1964). The dark platy Poker Chip shales are overlain by glauconitic arenaceous beds of the Rock Creek Member. Thick black shale, with abundant glauconite pellets, of the Green Beds grade upward into siltstone and sandstone of the Passage Beds. Farther south, toward Sweetgrass Arch, much of the Middle and Upper Jurassic is not present, due partly to non-deposition and partly to later erosion.

Sweetgrass Arch and Williston Basin. The region of the Sweetgrass Arch in southern Alberta, emergent in Lower Jurassic time, was depressed during the Middle and Upper Jurassic and received a maximum of more than 600 feet of sediments (Thompson and Crockford, 1958). The Middle Jurassic Bajocian Sawtooth Formation, 0 to 235 feet, comprises two quartzose sandstone members separated by green shale with belemnite conglomerates. It was deposited on the irregular erosion surface carved on Mississippian rocks around several islands that remained along the crest of the arch. The epineritic Sawtooth sandstones grade eastward into limy sandstone of the Shaunavon Formation and westward into black shales of the Rock Creek Member. The Callovian Rierdon Formation contains as much as 250 feet of alternating grey to green calcareous shale, brown limestones, and grey to green argillaceous siltstones. It was deposited over Sweetgrass Arch, grading laterally westward into the shales of the Grey Beds of Alberta Trough and eastward into the Lower Vanguard calcareous shales of Williston Basin. The Oxfordian Swift Formation, ranging from 0 to 75 feet, of non-calcareous shale and fine quartzose silty sands, in part glauconitic, appears to lie disconformably on the Rierdon, the result of positive movements of Sweetgrass Arch at the end of the Middle Jurassic. The Swift beds grade into the light coloured shales of the Upper Vanguard and westward into the Green Beds and Passage Beds. Erosion in the late Jurassic and earliest Cretaceous removed the Swift beds from the crest of the arch, truncating older beds in a northward direction.

In Williston Basin, Jurassic rocks have a maximum thickness of more than 1,400 feet (Milner and Blakslee, 1958; Springer, *et al.*, 1964). Lower Jurassic red beds of the Lower Watrous Formation composed of siltstone, shale, and sandstone lie unconformably on Mississippian and Devonian rocks, filling depressions on a highly irregular erosional surface. These sediments, overlapped by the remainder of Jurassic deposits, were deposited in an oxidizing environment along the margin of an evaporitic embayment centred in the northern United States. Continued subsidence resulted in an expansion of the basin and the deposition of upper Watrous anhydrite, dolomite, and shales. Along Sweetgrass Arch an extensive peninsula separated the evaporitic waters of Williston Basin and the open seas of Alberta Trough. During Middle Jurassic time, silty and dolomitic marine shales of the Gravelbourg Formation accumulated in the central part of Williston Basin and quartzose sandstone and shale were deposited under shallow near-shore conditions along the eastern margin. The Gravelbourg is overlain with minor disconformity by the Shaunavon Formation as a westward shifting of the centre of downwarp resulted in the deposition of widespread, fine-grained sandstone and oölitic limestone in the west, at the same time as shale with minor sandstone was deposited in the east. Interbedded shales and sandstone derived from the west and southwest comprising the upper Shaunavon suggest regression during the Middle Jurassic, and subsequently transgression is recorded by the shales of the lower Vanguard Formation. Sandstones within the middle Vanguard are unconformably overlain by upper Vanguard marine shales, the result of a general regression at the end of the Callovian and subsequent transgression in the Oxfordian. Emergence during latest Jurassic to early Cretaceous time resulted in general widespread erosion of the Jurassic rocks, particularly around the margin of the basin.

Cretaceous

Tectonic Summary

Lower Cretaceous. Sedimentation in the deeply subsiding troughs of the Cordilleran Geosyncline continued from the Jurassic into the early Lower Cretaceous (Fig. VIII-35).

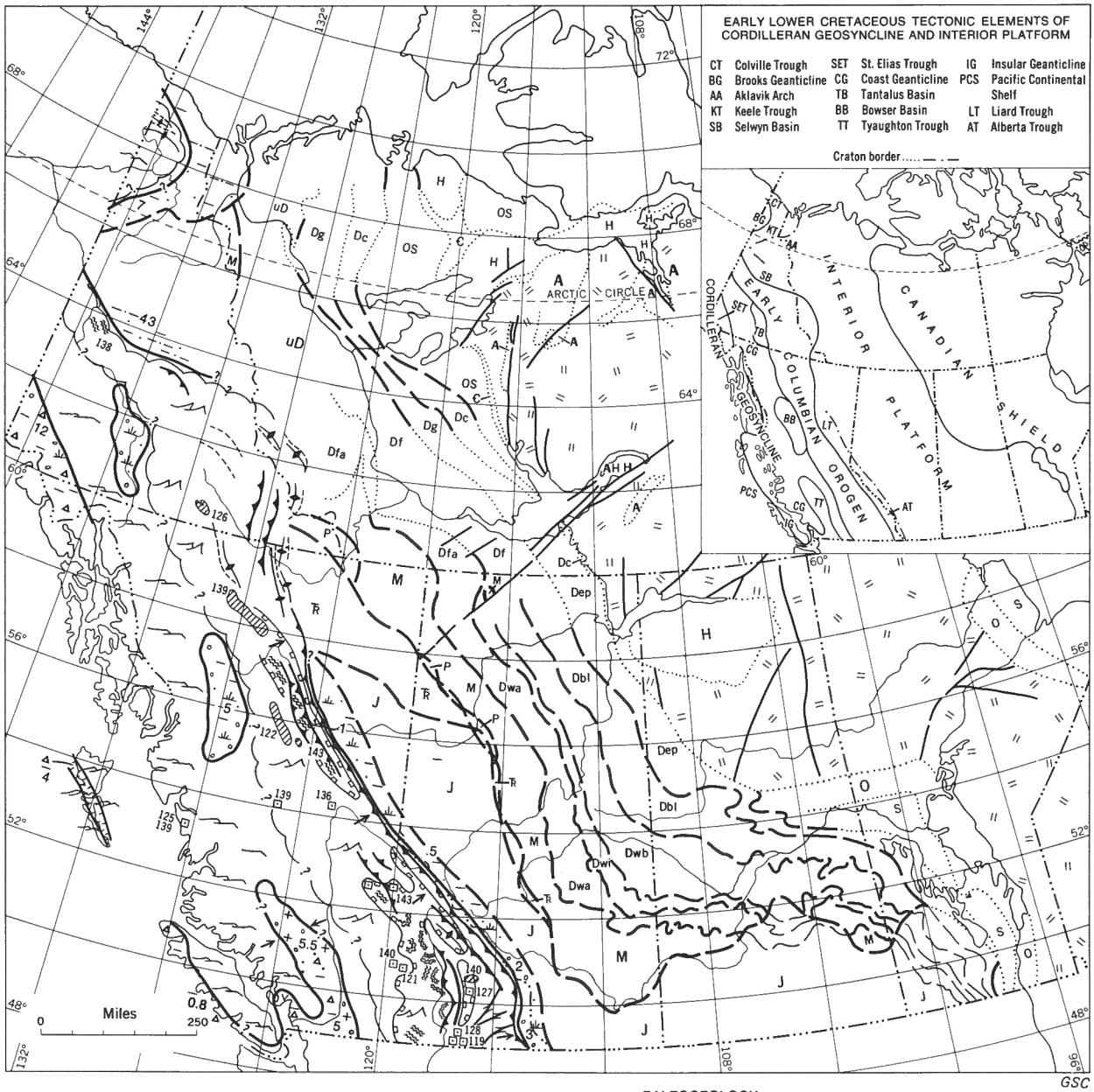
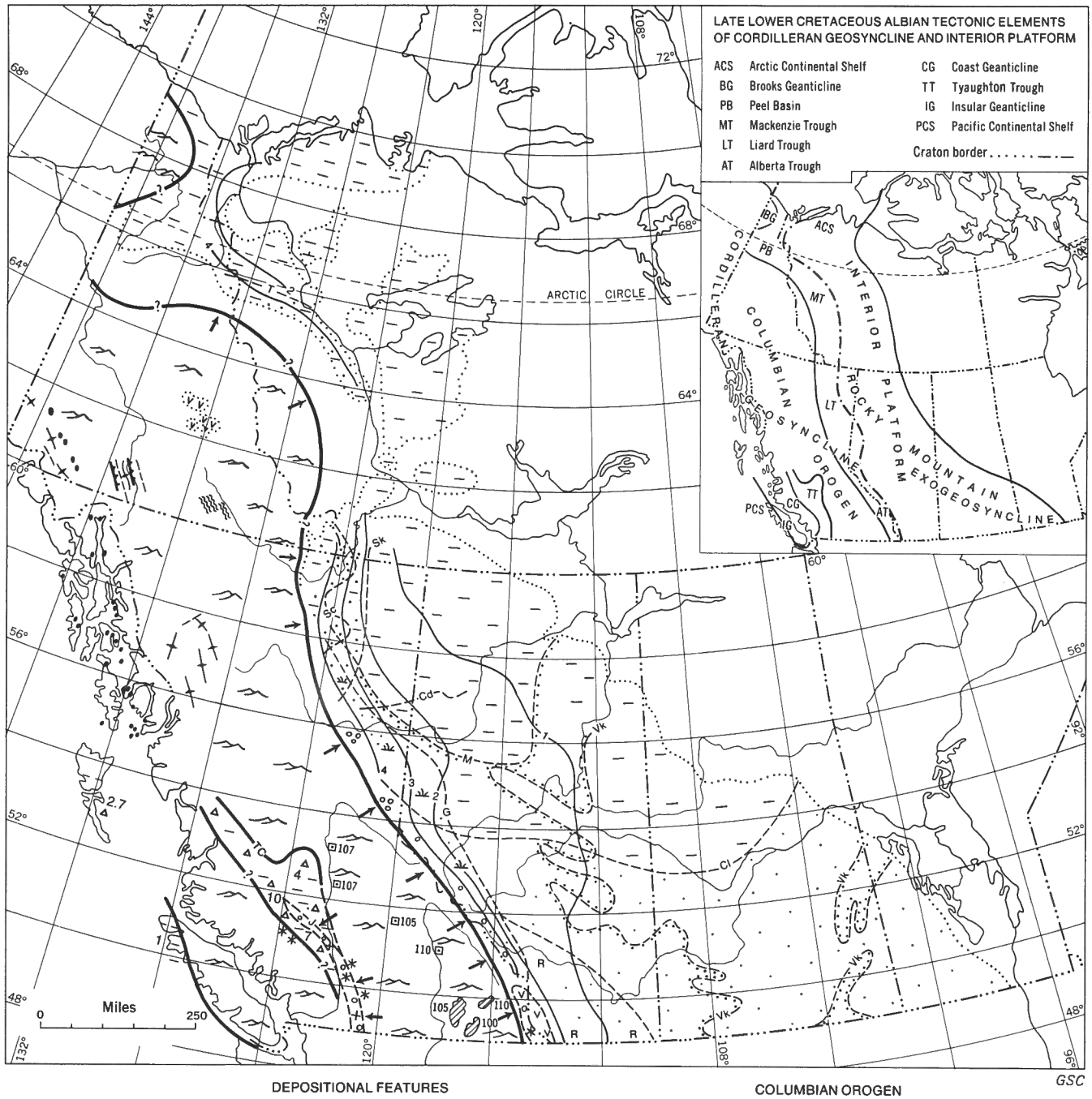


FIGURE VIII-35. Early Lower Cretaceous sedimentation and tectonism in Western Canada, and some features of the early phase of the Columbian Orogeny and of pre-Albian geology.



DEPOSITIONAL FEATURES

- Conglomerate
- Granitic clasts
- Greywacke
- Sandstone (marine)
- Sandstone, shale (non-marine)
- Shale (marine)
- Volcanic rocks
- Ultramafic rocks
- Coal
- Red beds
- Volcanoes (hypothetical)
- Depositional limit
- Outcrop limit
- Facies boundary
- Cadotte Cd
- Clearwater Cl
- Gething G
- Jackass Mtn. J
- Manville M
- Scatter Sc
- Sikanni Sk
- Trevor T
- Taylor Creek TC
- Viking Vk
- Isopach (thousands of feet)
- Thickness (thousands of feet)
- Direction of sediment transport

COLUMBIAN OROGEN

- Fault
- Thurst fault
- Anticline
- Syncline
- Granitic plutons (age in millions of years)
- Metamorphic trends

PHYSIOGRAPHIC FEATURES

- Mountainous land

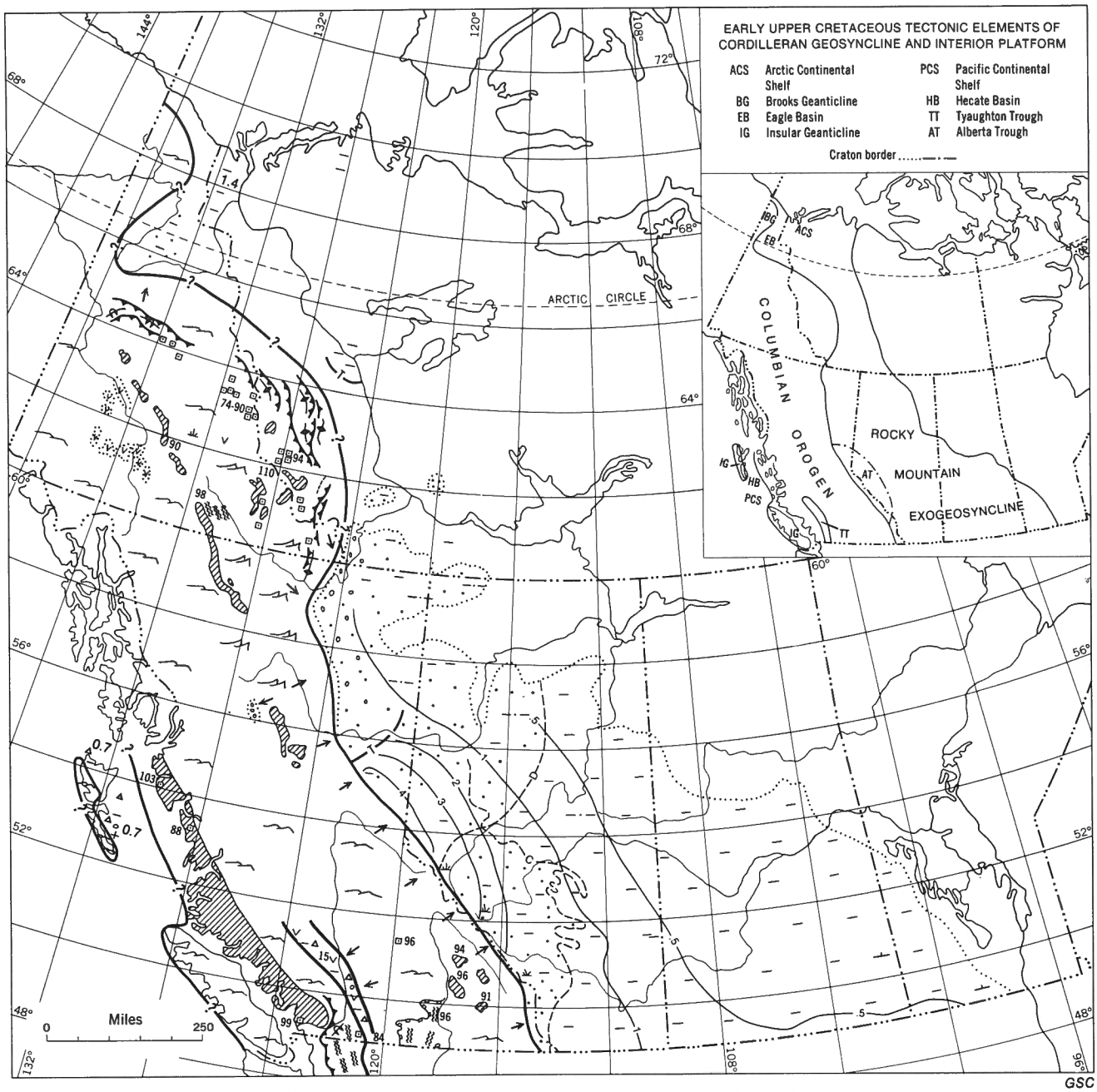
Blank areas are lowlands and plains developed on crystalline and sedimentary rocks of the craton, and where Albian is now absent through erosion, or unknown.

FIGURE VIII-36. Lower Cretaceous, Albian, sedimentation and tectonism in Western Canada, and some features of the middle phase of the Columbian Orogeny.

Some, such as Keele and Alberta Troughs, appear to have been foredeeps that were closely linked with adjacent rising geanticlinal welts and received thick clastic sediments deposited partly under non-marine conditions. In other elements, such as Bowser and Tantalus Basins and Tyaughton Trough, which lay between the geanticlines, deposition occurred contemporaneously with the early phase of Columbian Orogeny. Large granitic plutons were emplaced during the orogeny and extensive metamorphism occurred throughout the Omineca Geanticline. The orogeny and associated uplift were most intensive in the south and possibly also occurred in the Brooks Geanticline west of the Alaska Boundary. To what extent the orogeny affected the Coast Geanticline is not known but this belt was generally uplifted, much debris was supplied to St. Elias Trough and Pacific Continental Shelf, and the marine connection to Bowser Basin was greatly restricted. Sedimentation in Queen Charlotte Islands was restricted to a narrow, deeply depressed central graben. A boreal marine embayment, persisting from the late Jurassic, probably extended from the region of Keele Trough and Selwyn Basin into northern parts of Alberta Trough. By Aptian time, however, the sea had completely withdrawn from the craton and bordering troughs. Preserved beneath Albian sediments of the craton is a generally homoclinal succession of Paleozoic and early Mesozoic formations and many remnants of the topography that were established in the Aptian, or that persisted from earlier periods of cratonic uplift.

In the Albian (Fig. VIII-36) the central part of the western Cordilleran Geosyncline appears to have been land, marine conditions prevailing only in Tyaughton Trough and on the Pacific Continental Shelf. Tyaughton Trough was partly taphrogenic, and received much greywacke, shale, and volcanic rocks and some extremely coarse conglomerates that were derived from fault-bounded mountains to the east. Volcanism took place in the north, the basalt flows covering broad intermontane areas of moderate relief in central Yukon. East of the Columbian orogenic lands the early Albian sea transgressed rapidly southward over a plain of little relief. The adjacent basins and troughs were strongly depressed and the craton subsided evenly over broad areas. Peel Basin was partly exogeosynclinal, the clastic sediments derived from uplift of northern Columbian Orogen to the south. Mackenzie and northern Liard Troughs may have been more akin to foredeeps. In the south the vast Rocky Mountain Exogeosyncline developed and although much sediment was carried out over the craton, a substantial part was trapped in the strongly depressed linear Alberta Trough. The Albian clastic sediments were derived from the southern mountainous part of Columbian Orogen to the west and in small part from the east, from the Canadian Shield and its Paleozoic cover. The bulk was deposited under lacustrine, alluvial and neritic environments and formed a broad, low-lying, alluvial plain that bordered the Albian seas to the north. Variations in the rates of subsidence of the sea floor, of the influx of clastic material, and of uplift of the source region resulted in three major southward marine transgressions that alternated with three northward regressions and overlap of the marine shales by non-marine clastics, and also in several minor cyclical facies alternations. The non-marine facies do not extend beyond a northeast-trending hingeline that lies subparallel with the Devonian Peace River Arch, greater depression and marine shale facies prevailing to the north. Near the mountains, the alluvial plain was swampy and heavily forested by conifers in the early Albian and later also by angiosperms. The southwest was semi-arid and the site of small volcanoes. Tongues of sandstone of western source occur in northern Liard Trough, reflecting uplift of the central Omineca Geanticline, a prominent feature of Columbian Orogen in the early Upper Cretaceous. The latest of the Albian transgressions extended into the central United States joining the sea that had spread northward from the Gulf of Mexico.

Upper Cretaceous. Columbian Orogeny continued in the early Upper Cretaceous, the last phase represented by regional uplifts, minor folding and intrusion of post-orogenic granites (Fig. VIII-37). Synorogenic clastic sediments, including conglomerates, were deposited in the northern Rocky Mountain Exogeosyncline, Pacific Continental Shelf, Tyaughton Trough, and the northern elements of the geosyncline. These sediments conformably or disconformably succeed the Lower Cretaceous and are conformably succeeded by marine shales that reflect relative orogenic quiescence during the mid-Upper Cretaceous. A great deal of the Cordilleran Region was land, mainly mountains. In broad intermontane valleys and trenches, forming Sustut and Tyaughton Troughs and the Rocky Mountain and Tintina Trenches, non-marine clastic sediments accumulated and locally there was volcanism, partly pyroclastic and acidic. These rocks are not well dated; the age of the base of the successions probably varies generally, and may also be partly early Tertiary. In general, they lie nonconformably or with angular unconformity on older rocks, reflecting the effects of the local phase of Columbian Orogeny or the cumulative effect of all phases and older orogenies as well. The sequences are partly marine in Hecate and Georgia Basins and Tyaughton Trough. In Rocky Mountain Exogeosyncline the late Upper Cretaceous clastic sediments are very thick, representative of the widespread uplift of southern Columbian Orogen and the effects of the early phase of the Laramide Orogeny that continued into the Tertiary (Fig. VIII-44). The sediments grade from non-marine sandstone, shale, and conglomerate in the west to marine shale in the east and are conformably to unconformably overlain by the Tertiary. Regional uplift, including the northern part of the craton, resulted in the southward withdrawal of the seas from the continental interior.



DEPOSITIONAL FEATURES

- Conglomerate
- Granitic clasts
- Greywacke
- Sandstone
- Siltstone
- Shale
- Limestone
- Volcanics
- Coal
- Depositional limit
- Truncated limit of Turonian rocks
- Facies boundary
- Cardium sandstone
- Dunvegan sandstone
- Outcrop limit
- Direction of sediment transport
- Isopach (thousands of feet)
- Thickness (thousands of feet)

COLUMBIAN OROGEN

- Anticline
- Anticline overturned
- Syncline
- Thrust fault
- Granitic plutons (age in millions of years)
- Trend of metamorphic rocks
- Metamorphic front

PHYSIOGRAPHIC FEATURES

- Mountainous land (high, low)

Blank areas are lowlands and plains developed on crystalline and sedimentary rocks of the craton, and where the early Upper Cretaceous is now absent through erosion, or unknown.

FIGURE VIII-37. Early Upper Cretaceous sedimentation and tectonism in Western Canada, and some features of the late phase of the Columbian Orogeny.

Columbian Orogeny

From Late Jurassic to earliest Upper Cretaceous time parts of the Cordilleran Geosyncline underwent major deformation, regional metamorphism, granitic intrusion, and uplift—the Columbian Orogeny—during which the following structural provinces were established: (1) *Eastern Columbides,* a northeasterly directed alpine-type orogen in the south, whose core zones of plutonic rocks constitute much of the Omineca Geanticline and in the north include the arcuate Selwyn Fold Belt and the eastern termination of Brooks Geanticline; (2) *Columbian Zwischengebirge,* characterized by a heterogeneous mosaic of northwesterly elongate, uplifted blocks with intervening troughs and successor basins; (3) *Western Columbides,* comprising the Coast, Cascade, Insular, and St. Elias segments, dominated by plutonic rocks and steeply dipping, southwesterly directed structures.

Columbian Orogeny consisted of three phases. The earliest, occurring in the latest Jurassic, was intense but restricted mainly to Omineca Geanticline. The middle phase, in Aptian and Albian time, was more widespread. The late phase in earliest Upper Cretaceous was manifested mainly by intrusion of granite bodies.

Eastern Columbides. While the syntectonic basins were receiving clastic deposits in latest Jurassic time, deformation and plutonism occurred in the Omineca Geanticline. The youngest rocks deformed are the Middle and Upper(?) Jurassic upper part of the Rossland Group whereas the oldest rocks, not involved, are those of the Uslika Formation, probably of Aptian age. Radiometric dates on metamorphic and granitic rocks in the southern Omineca Geanticline and southern Quesnel Trough and evidence from textural studies, such as alignment of metamorphic minerals parallel with the axes of early folds, indicate that the early Columbian deformation began in the Portlandian or Kimmeridgian. Two K–Ar ages of 140 m.y. and one of 135 m.y. have been obtained from metamorphic rocks of the Mount Ida Group in the western part of the Omineca Geanticline, and one age of 146 m.y. from northeast Selkirk Mountains. The latter appears to represent the time of growth of biotite poikiloblasts in a schist, the muscovite matrix of which gave an age of 205 m.y. Elongate, syntectonic or oval, late tectonic biotite–hornblende granodiorite plutons such as the Cassiar and Nelson batholiths, and the Porcupine, Ruby Range, and Glacier Creek stocks yield K–Ar dates ranging from 136 m.y. to 152 m.y. These plutons are locally foliated and in gross aspect parallel the structural grain. Furthermore, the discordant, post-tectonic biotite granodiorite plutons, roughly circular in plan, yield dates of less than 143 m.y.

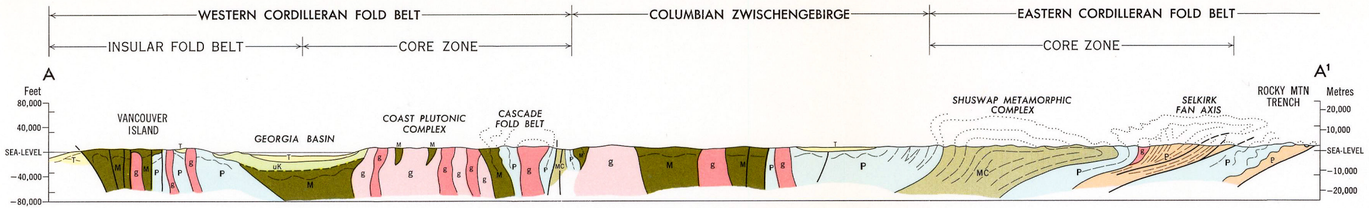
The core zone of the Eastern Columbides contains several structural culminations and areas of high-grade metamorphism (Fig. VIII-28). The overall pattern of structural trends, as displayed by the long dimensions of the metamorphic complexes, the fault traces, and by the fold axes in the rocks adjoining them, is that of a gentle arc, convex eastward.

The deeper structural levels exposed within the culminations are featured by multiphase folding in which earlier attenuated isoclines have been redeformed by later folds to produce several huge, partly interdigitating and backfolded nappes with profiles resembling that of an anvil. The frontal parts of the nappes correspond to the horn and the backfolds to the rear of the anvil (Fig. VIII-38). The higher structural levels, well displayed in Cariboo Mountains, are characterized by single phase, broad, open folds and fault blocks.

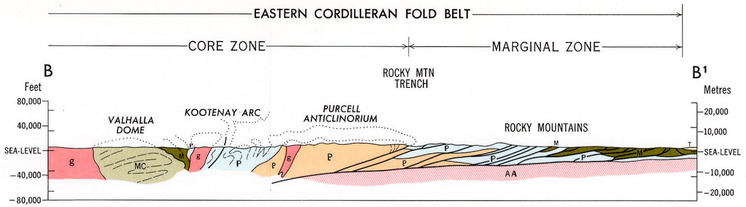
In cross-section the anvil structures are characterized by broad undulating arches whose western parts have subsidiary northeast-dipping reverse faults and late parasitic folds, commonly asymmetrical or overturned to the southwest. The late folds in the western part of the Shuswap anvil structure, however, are upright whereas those in the northwesternmost part are tight with axial planes and associated cleavage progressively changing attitude westward from northeasterly dips to southwesterly. The eastern part of the Shuswap anvil structure is represented by the enveloping surface of the late folds in the Shuswap Complex which dips eastward under the back-folded part of the Selkirk anvil structure lying to the east. The axial planes of the late folds and attitudes of thrust faults in the Selkirk anvil structure are arranged in the form of a fan; those in the southwestern part dip northeast, pass through vertically along the fan axis, and then dip southwest in the northeastern part of the structure. Structures in the northeastern part are characterized by overturned and undulating recumbent folds that are part of thrust masses that truncate and override structures of the westernmost Rocky Mountains. The Selkirk anvil structure is probably largely allochthonous and westerly directed structures within it may have resulted from underthrusting of the nose of the Shuswap anvil structure into the back of the Selkirk anvil structure.

The Purcell anticlinorium is composed of several north-plunging segments separated by faults. In cross-section the anticlinorium is crudely anvil-shaped. The structures in its western part merge with those of the Kootenay Arc which are featured by complex repetitive folding whereas those in the north plunge under the eastern part of the Selkirk anvil structure. The central part is dominated by a broad, gently undulating anticline transected by southwest-dipping thrust faults, north-northwesterly trending normal faults, and smaller folds co-axial with those formed during the Hadrynian East Kootenay Orogeny. The Moyie, Hall Lake, and St. Mary's faults existed prior to the intrusion of the White Creek and Bayonne batholiths. Some of the early-formed faults acted as oblique thrusts as the intervening segments moved differentially eastward (Leech, 1962b).

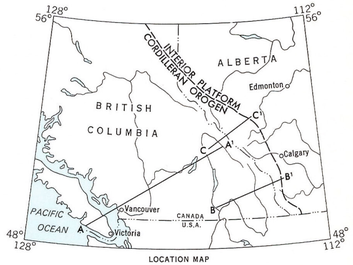
In the western ranges of southern Rocky Mountains a thin sequence of lower Paleozoic carbonate, quartzite, and pelitic rocks is thrown into several southwesterly over-



a) Vancouver Island to Rocky Mountain Trench (by J.O. Wheeler)



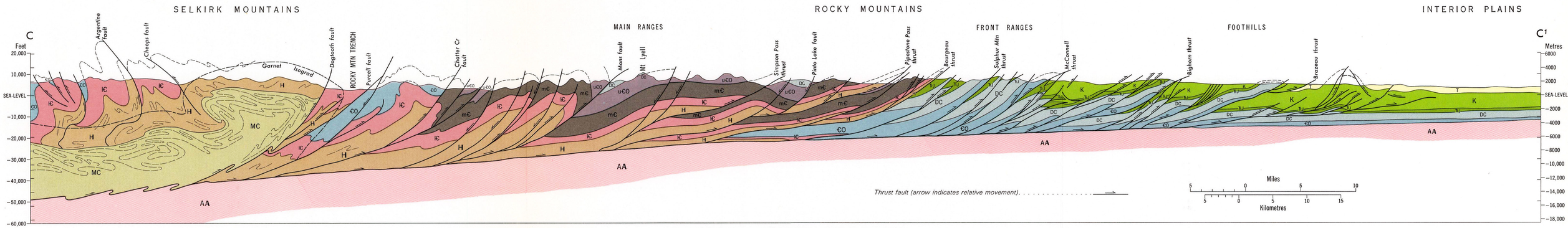
b) Southern Selkirk Mountains to Rocky Mountains (by J.O. Wheeler)



TERTIARY — T
UPPER CRETACEOUS — uK
MESOZOIC — M (Undivided)
PALEOZOIC — P
PROTEROZOIC — P (Sedimentary rocks)
ARCHEAN AND/OR APHEBIAN — AA (Crystalline rocks)
g g (Granites)
MC (Metamorphic Complex)

Fault

Miles: 20 0 20 40
Kilometres: 25 0 25 50



Thrust fault (arrow indicates relative movement)

Miles: 5 0 5 10
Kilometres: 5 0 5 10 15

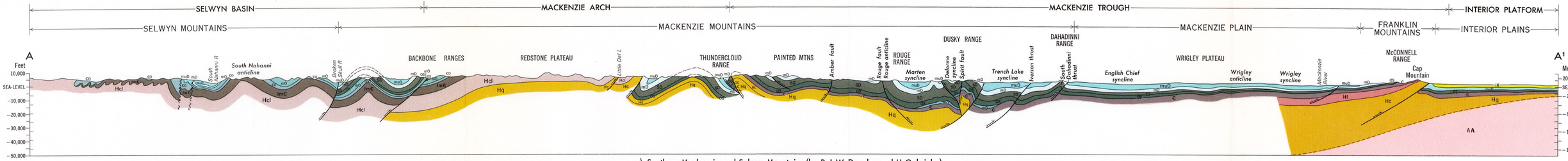
c) Southern Rocky Mountains (by R.A. Price and E.W. Mountjoy)

TERTIARY — T — Paskapoo Formation
CRETACEOUS — K — Brazeau Formation, Alberta Group, Luscar, Nikanassin Formations
TRIASSIC AND JURASSIC — TrJ — Spray River, Fernie Groups
LATE PALEOZOIC — DC — Fairholme Group, Sassenach, Palliser, Exshaw, Banff Formations, Rundle, Rocky Mountain Groups
UPPER CAMBRIAN AND ORDOVICIAN — uЄO — Lynx, Survey Peak, Outram, Skoki, Owen Creek Formations
MIDDLE CAMBRIAN — mЄ — Mount Whyte, Cathedral, Stephen, Eldon, Pika, Arctomys Formations
CAMBRIAN AND ORDOVICIAN — ЄO — Lardeau Group, Chancellor, Ottertail Formations, McKay Group
LOWER CAMBRIAN — lЄ — Gog, Hamill Formations
HADRYNIAN — H — Miette, Kaza Groups
PRECAMBRIAN OR PALEOZOIC — MC — Metamorphic Complex
ARCHEAN AND/OR APHEBIAN — AA — Crystalline rocks

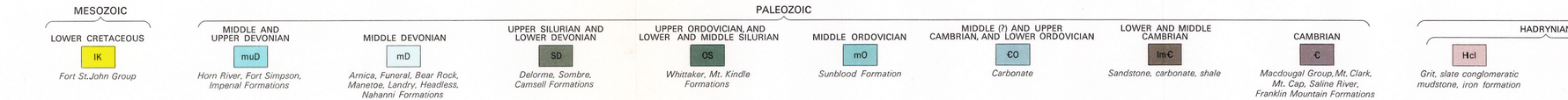
Cartography by the Geological Survey of Canada

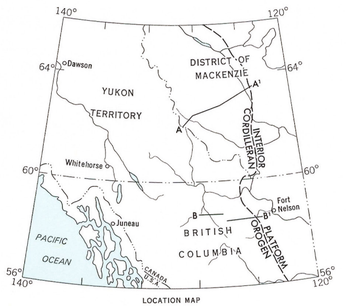
Figure VIII-38. Structure-sections through southern Cordilleran Orogen

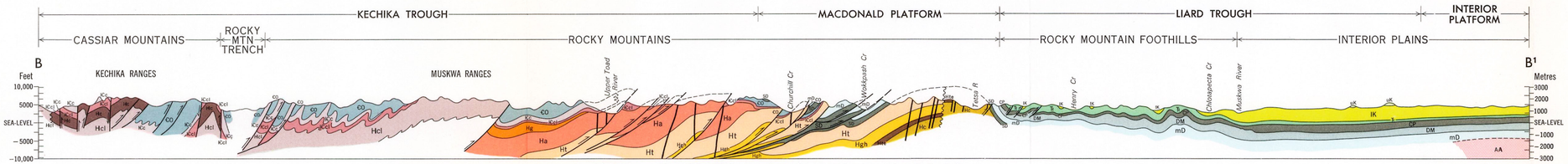
Printed by the Surveys and Mapping Branch



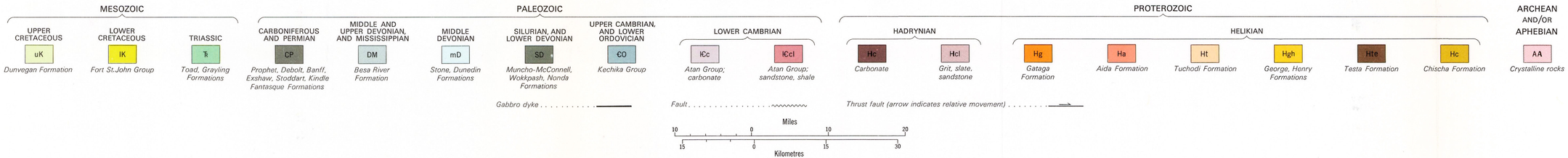
a) Southern Mackenzie and Selwyn Mountains (by R. J. W. Douglas and H. Gabrielse)







b) Cassiar and northern Rocky Mountains (by H. Gabrielse and G. C. Taylor)



Cartography by the Geological Survey of Canada

Figure VIII-39. Structure-sections through northern Cordilleran Orogen

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PLATE VIII-12. Folds with steeply west-dipping axial plane cleavage in Cambrian, Arctomys and Waterfowl Formations, near Sullivan River, southern Rocky Mountains, British Columbia. Mount Tsar in right background.

turned and recumbent folds with southeast-dipping axial plane cleavage (Henderson, 1954). These folds are cut by the Redwall and other high-angle strike-slip faults, commonly marked by zones of breccia and mylonite. Between the faults the strata are imbricated by thrusts or segmented by small cross-faults.

To the northeast in western Main Ranges, across a fan-axis marked by vertical cleavage, the dominantly pelitic Cambro-Ordovician succession is intricately folded, the cleavage and axial planes dipping southwest. Comparable folds also occur locally in the alternating shale and limestone succession of the central Main Ranges (Pl. VIII-12). To the southeast the Columbide structural front is limited by the presence of Jura-Cretaceous clastic rocks in Fernie Basin (Fig. VIII-35).

About 20 miles north of the Big Bend of Columbia River, in a culmination in the main ranges of the Rocky Mountains, a mass of pre-Hadrynian gneiss is thrust over Lower Cambrian and Hadrynian strata (Fig. VIII-38a). The enveloping staurolite–kyanite, garnet, and biotite isograds gradually cut across the westernmost thrusts and folds of the main ranges. Post-kinematic, porphyroblastic biotite has yielded a K–Ar age of 111 m.y. which provides an upper limit on the time of formation of the structures (Price and Mountjoy, 1970).

The structures of the stratified rocks of Cassiar Mountains (Gabrielse, 1963) are extremely heterogeneous and certain ones are confined to the individual northwesterly trending blocks that are bounded by major steeply dipping faults (Fig. VIII-39b). In some blocks the folds are overturned to the northeast whereas in others they are upright or overturned southwest. Some faults bounding the blocks coincide with marked changes in thickness and facies of the Lower Paleozoic strata. Numerous small cross-faults contribute to the iscontinuous distribution of stratigraphic units. In southern Cassiar Mountains northeasterly overturned folds with steeply dipping axial planes occur with northeasterly directed thrust faults (Gabrielse, 1962b). The structures are not nearly as continuous as the thrusts and folds in western Rocky Mountains but otherwise their style is similar and they may have been produced during the same phase of deformation. In western Rocky Mountains this deformation predates deposition of the Sifton Formation of Paleocene and possibly Upper Cretaceous age in Rocky Mountain Trench, and in Cassiar Mountains predates deposition of the Santonian to Campanian continental sediments that occur in a northwesterly trending trench that forms an acute angle with Rocky Mountain Trench. Accordingly, the structures were formed during Columbian Orogeny, and could be as young as late Columbian.

The southern part of the Cassiar batholith from which a muscovite–biotite pair has yielded K–Ar ages of 123 m.y. and 139 m.y. lies subparallel with the regional structure but is intrusive into previously metamorphosed and deformed rocks. The Marker Lake batholith in northern Cassiar Mountains with a K–Ar date of 126 m.y. is possibly also related to the early phase of Columbian Orogeny. It is mainly a medium-grained uniform granodiorite or quartz monzonite.

In Pelly Mountains two areas of different structural style are separated by a major northeasterly directed thrust fault. Southwest of the thrust, the structure is characterized by moderately west-dipping homoclines and low-angle thrust faults, some of which are folded. Displacement is directed mainly to the northeast. Displacement on some steeply dipping thrust faults in the southwestern part of the region is southwesterly. Locally, crystalline rocks have been thrust northeasterly over Devono-Mississippian clastic strata. The thrust sheets have been broken subsequently by a north-northwesterly trending normal fault. Northeast of the main thrust and flanking Tintina Trench, structures are featured by steep dips, tight upright folds, and fault slices bounded by steeply dipping faults.

In Ogilvie Mountains and Stewart Plateau near Tombstone and Mayo, tightly folded, sheared, foliated and locally metamorphosed strata possibly as young as early Lower Cretaceous display a marked northwesterly and northerly directed structural asymmetry (Tempelman-Kluit, 1966). Two major thrust faults and numerous subsidiary thrust faults produce an imbrication of thrust slices above a basal *décollement*. The rocks are also involved in overturned folds (Fig. VIII-40). One major overturned syncline was later deformed by folds having northeasterly trending axes. Coarse- to medium-grained syenite, monzonite, and quartz diorite intrude the deformed sequence of Lower Cretaceous sediments and have K–Ar dates of 102 and 106 m.y. (Albian). K–Ar dates on the metamorphic rocks near Keno Hill range from 81 to

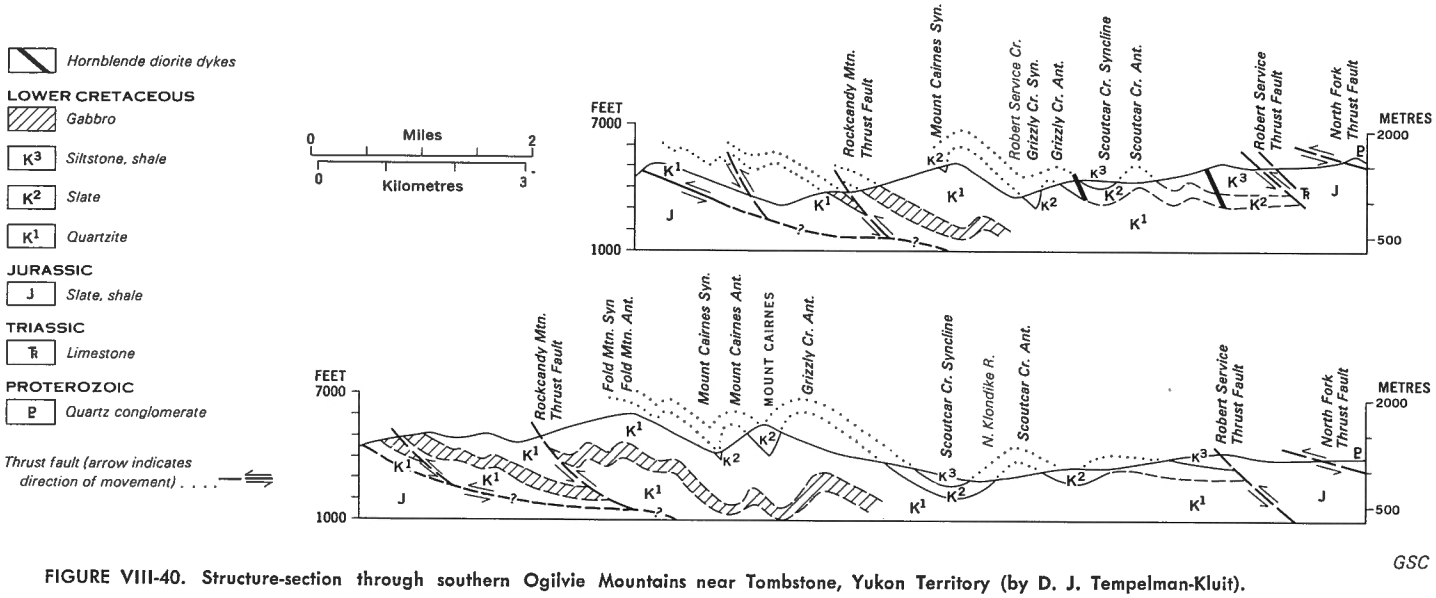

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FIGURE VIII-40. Structure-section through southern Ogilvie Mountains near Tombstone, Yukon Territory (by D. J. Tempelman-Kluit).

93 m.y. and indicate a later metamorphic event that may be related to the early Laramide Orogeny.

Dacitic and basaltic flows on MacMillan Plateau dated at 100 and 117 m.y. lie with marked angular unconformity on Paleozoic strata that had been strongly deformed along northwesterly trending axes. The flows were apparently extruded on a surface of considerable relief. Similarly, strata beneath the Hutshi Group were strongly deformed along northwesterly trending axes before deposition of the volcanics. The numerous late-orogenic plutons in Selwyn Fold Belt northeast of Tintina Trench (Pl. VIII-13) are characteristically coarsely megacrystic, homogeneous, quartz monzonites. K–Ar ages from 74 to 110 m.y. suggest a considerable range in time of emplacement.

The extent of Columbian deformation in Selwyn and Mackenzie Mountains is difficult to assess. In Selwyn Fold Belt the structures are cut by plutons, some of which have yielded K–Ar dates as old as early Upper Cretaceous. In southern Mackenzie Fold Belt the Columbide structural front is limited by the presence of Albian clastic rocks and Cenomanian conglomerates (Figs. VIII-36 and 37). Major deformation in northern Mackenzie and Wernecke Mountains preceded deposition of the late Cretaceous to early Tertiary sediments of Bonnet Plume Basin but had not occurred in the region of Monster syncline in southwestern Porcupine Plateau. Probably then, the deformation in Selwyn and much of Mackenzie Mountains took place during the Columbian Orogeny.

Columbian Zwischengebirge. The Mesozoic formations in the zwischengebirge have structural characteristics that are related partly to their lithology, and hence are dependent on their competency, and partly to the behaviour of adjoining geanticlines and fault blocks. In the Mesozoic rocks of the Nechako and Quesnel Troughs the structures of the competent predominantly volcanic and volcanic-clastic rocks are broad folds, warps, and steep faults, and generally reflect relatively little shortening. They trend northwesterly or north-northeasterly in the Nicola Group at the south end of the Pinchi Geanticline and northeasterly across Skeena Arch. Tight folds and the local development of cleavage occur near the Pinchi fault and other faults bounding the Pinchi Geanticline. More tightly compressed folds are probably related to the movements between the major fault blocks, which, in the case of the Pinchi fault, continued until the mid-Tertiary.

The early Mesozoic formations of Whitehorse Trough and the Upper Jurassic and Lower Cretaceous beds of Tantalus Basin were closely folded and faulted, probably during the mid-Columbian phase, prior to being bevelled and overlain unconformably by the Upper Cretaceous Hutshi Group. The intensity of the folding is governed partly by their lithology, as tight folds are characteristic of the turbidite facies of the Laberge Group and broad, open folds are typical of the conglomerate, near-shore facies. Regional metamorphism within Columbian Zwischengebirge was slight. The Mesozoic and older rocks are generally of low greenschist facies or less. Zeolite and prehnite–pumpellyite facies occur locally and glaucophane schist facies is developed in the late Paleozoic volcanics near Pinchi fault and in Atlin Horst. The age of the latter metamorphism is uncertain and could be contemporaneous with the associated Permo-Triassic ultramafic intrusions.



PLATE VIII-13. Late Columbian (Cretaceous) granitic pluton, forming peaks, intrusive into dark weathering Ordovician and Silurian Shales and Upper Cambrian carbonates (foreground), Ragged Range, Selwyn Mountains, Yukon Territory.

Western Columbides

The effects of Columbian Orogeny in the Coast and Cascade Mountains and Yukon Plateau have not generally been separated from those of other orogenies. They are described in the section on Plutonic Rocks.

Lower Cretaceous of Western Cordilleran Geosyncline

Tyaughton Trough. The early Lower Cretaceous, Berriasian to Barremian, part of the Relay Mountain Group has about the same facies and distribution as the Upper Jurassic part. Shaly rocks occur in the central part of the trough, and greywacke occurs on the flanks. In the northeast, the group is conformably overlain by the non-marine Jackass Mountain Group of Aptian to early Albian age. A thick lower division of greywacke and shale is overlain by a middle division of 2,000 to 3,000 feet of granite-bearing boulder conglomerate, succeeded by an upper division of several hundred feet of greywacke. The conglomerate was apparently derived from uplift to the northeast, in the southwest part of Columbian Orogen (Fig. VIII-35). In the southwest, the Relay Mountain Group is overlain by a shale and greywacke unit that is overlain conformably by the marine, late Aptian and Albian Taylor Creek Group. The basal conglomerate of the latter contains no granitic material and, although approximately equivalent in age to the conglomerate division of the Jackass Mountain Group, was probably derived from Coast Geanticline southwest of the trough. The western facies of the Taylor Creek Group is tuff and volcanic breccia with abundant black shale conformably overlain by the Upper Cretaceous Kingsvale Group.

In Fraser Valley, equivalents of the Valanginian part of the Relay Mountain Group are the Lillooet and Brew Groups (Duffell and McTaggart, 1952). Granite-bearing conglomerate occurs in the upper part of the latter. Farther south, the Dewdney Creek Group consists of several thousand feet of interbedded siltstone, volcanic arenite,

greywacke, and conglomerate containing Hauterivian and Barremian fossils. A conglomerate, as much as 2,000 feet thick, occurs 300 feet above the Barremian fossils and is probably of Aptian age. It contains granitic detritus which is dissimilar to the Mount Lytton batholith and Eagle granodiorite outcropping a short distance to the northeast; the source is unknown. The Albian part is marine siltstone. The predominantly non-marine Pasayten Group to the east contains Albian plants and is interbedded with eastward protruding tongues of the marine Albian strata of the Dewdney Creek Group (J. A. Coates).

In the Coast Mountains, near Harrison Lake, Lower Cretaceous strata contain much pyroclastic debris. The Peninsula and Brokenback Hill Formations are early Berriasian to Hauterivian and lie disconformably on Middle and Upper Jurassic rocks. The Fire Lake Group farther northwest occurs as roof pendants in the Coast Plutonic Complex. It is a heterogeneous assemblage of andesitic volcanic rocks, argillite, greywacke, and granite-bearing conglomerate with Berriasian fossils. Similar roof pendants on the west side of the complex are referred to the Gambier Group (Roddick, 1965). As the debris is coarser in southernmost Coast Mountains than farther northeast the source area appears to be islands within the southernmost Coast Geanticline composed of granitic and metamorphic terrain and explosive volcanoes.

Mainly non-marine volcanic rocks of the Spences Bridge and Kingsvale Groups occur northeast of the belt of sedimentary strata deposited in Tyaughton Trough. The Spences Bridge Group comprises varicoloured porphyritic lavas of andesite, dacite, basalt, and porphyritic rhyolite and pyroclastics with blocks a foot to 2 feet wide and fragments of plant stems and twigs. Associated clastic sediments contain Aptian plants (Duffell and McTaggart, 1952). The group lies unconformably on the Cache Creek Group or the Guichon batholith and is in turn unconformably overlain by the Kingsvale Group. The basal beds of the latter are characterized by coarse clastic sediments as much as 1,000 feet thick. They contain Albian fossil plants and clasts of the Nicola Group and granitic rocks. The main part is basaltic and andesitic lavas and pyroclastic rocks that are relatively undeformed.

Two isolated occurrences of late Lower Cretaceous rocks were possibly deposited in a northern extension of Tyaughton Trough. Marine and non-marine Aptian and Albian clastic sediments were encountered in drillholes beneath Tertiary strata in the Interior Plateau west of Quesnel (H. W. Tipper). Marine Albian shales occur east of the Coast Mountains near Ootsa Lake. They are not in contact with older formations and are overlain by 2,000 to 3,000 feet of arkose succeeded by andesitic and basaltic volcanic rocks probably equivalent to the Kingsvale Group (Duffell, 1959).

Tantalus and Bowser Basins. Clastic sediments of Lower Cretaceous age occur in the upper part of the mainly Jurassic Tantalus Formation and Bowser Group. Near Hazelton, in southeastern Bowser Basin the alternating marine and non-marine Red Rose Formation, equivalent to part of the Bowser Group, is overlain conformably by the Brian Boru Formation of porphyritic andesite flows, breccias, tuff, and volcanic arenite (Sutherland Brown, 1960).

Omineca Geanticline. The Uslika Formation, which occurs in a fault-slice in Omineca Mountains, contains conglomerate with clasts as large as 10 inches in diameter of andesite and diorite (Roots, 1954; Armstrong, 1949). The sources of the latter are the early phases of the Omineca Intrusions that outcrop about 20 miles to the west. Fossil plants provisionally assigned to the Aptian occur in argillites associated with the conglomerate. Gently folded dacitic, andesitic, and basaltic flows with an aggregate thickness of more than 5,000 feet underlie a large area northeast of Tintina Trench, south of Tay River (Roddick and Green, 1961a, b). Individual flows range in thickness from 10 to 300 feet. The volcanic sequence unconformably overlies deformed Paleozoic strata and the flows are thought to have been extruded onto an erosion surface of moderate relief. Two samples of porphyritic dacite have given K–Ar dates of 117 m.y. and 100 m.y.

Pacific Continental Shelf. The One Tree Formation, which outcrops on islands off the west coast of Vancouver Island, lies disconformably on Upper Jurassic strata (Jeletzky, 1950). It is composed of clastic sediments containing marine fossils ranging in age from early Berriasian to mid-Valanginian. Near Quatsino Sound, greywacke and conglomerate contain fossils of late Valanginian to Hauterivian age and lie unconformably on the Bonanza Group or Jurassic granite. The Hauterivian is absent in the east where coquina limestone of possible early Barremian age lies disconformably or possibly unconformably on older Cretaceous rocks. The limestone is succeeded by late Barremian siltstone and sandstone which in the eastern part of Quatsino Sound are overlain by 2,000 feet of conglomerate. The conglomerate intertongues westward with quartzose sandstones. It is derived from the east and reflects uplift of the Insular Geanticline on Vancouver Island during the middle phase of the Columbian Orogeny. It is overlain by marine shales and sandstone similar to those in the Albian Haida Formation of Queen Charlotte Islands.

Late Valanginian to late Barremian clastic sediments on Queen Charlotte Islands form the Longarm Formation (Sutherland Brown, 1966). The lower part contains greywacke, conglomerate, and some volcanic rocks and the upper part consists principally of graded-bedded calcareous siltstone and greywacke that were deposited in a graben-like trough. It lies unconformably on the Jurassic Yakoun Formation and older rocks but is faulted against the Haida Formation and overlain, possibly unconformably, by the Upper Cretaceous Honna Formation. The Haida Formation comprises a lower sandstone member with Albian fauna and an upper shale member with Cenomanian and

Turonian marine fossils. It contains granitic detritus and lies unconformably on all older rocks except the Longarm Formation. Aptian beds are apparently absent on the Queen Charlotte Islands.

St. Elias Trough. The Dezadeash Group in St. Elias Mountains is composed of thick successions of graded-bedded greywacke, siltstone, and shale (Kindle, 1953). The lower beds in the eastern exposures contain coal and conglomerate with granitic clasts apparently derived from the Coast Geanticline to the northeast. The group has yielded marine fossils ranging from late Tithonian to middle Valanginian.

Selwyn and Keele Troughs. An apparently conformable sequence of clastic rocks of possible Lower Cretaceous age occurs in the Tintina–Tombstone area, Yukon Territory (Tempelman-Kluit, 1966). The succession comprises: a lower member of thick-bedded quartzite and minor black slate about 1,200 feet thick; a middle member of pale green and bright red cherty slate 100 to 200 feet thick; and an upper member of crosslaminated, lime-cemented siltstone with interbedded black slate at least 1,500 feet thick. These rocks may conformably overlie Jurassic strata and are tentatively correlated with similar sequences in part of Valanginian age, in Kandik Basin of Alaska and southern Keele Range, Yukon Territory (Frebold, *et al.*, 1967). A tholeiitic diabase and gabbro sill, 300 to 800 feet thick, intrudes the quartzite member.

Lower Cretaceous of Northern Cordilleran Geosyncline

Keele Trough. A generally conformable sequence of shale and sandstone, partly non-marine, Valanginian to Aptian in age, 4,000 to 6,000 feet thick, conformably overlies Upper Jurassic–Berriasian shales and is disconformably succeeded by Albian shales. These rocks, which were deposited in an embayment of Arctic Sea, are thought to have been derived mainly from uplift of the Brooks Geanticline during Columbian Orogeny, and partly from cratonic land adjacent to southern Richardson Mountains. In northern Richardson Mountains the lowermost Albian, 350 feet of buff, fine-grained, thinly bedded, hard, quartzose sandstone, is overlain by about 350 feet of whitish grey, fine- to coarse-grained, quartzose sandstones and associated carbonaceous siltstone, sandstone, shale, and coal. In central Richardson Mountains the North Branch conglomerate, 600 feet thick, marks the northern flank of Aklavik Arch during the Valanginian and Hauterivian. Dark grey concretionary Barremian shales and siltstones, 3,000 feet thick, conformably succeed and overlap southerly the older strata and are in turn overlain and overlapped southerly by about 600 feet of light grey, buff, fine- to medium-grained, well-sorted quartzose and micaceous sandstones. In southern Keele Range and Eagle Plain the succession is comparable to that of northern Richardson Mountains but considerably thicker.

Peel Basin. On Peel Plateau, the Albian comprises four units: a basal glauconitic sandstone as much as 350 feet thick; black to brownish grey, soft, fissile, concretionary shale with *Cleoniceras, Beaudanticeras,* and *Lemuroceras,* about 500 feet thick; some 500 to 1,000 feet of dark grey silty and bentonitic shale; and an upper silty shale and glauconitic siltstone at least 500 feet thick (E. W. Mountjoy). A fine-grained calcareous sandstone, 1,200 feet thick, in part laterally equivalent, grades northwards into the upper units. The sandstone unit contains *Gastroplites* and is overlain by 400 feet of undated silty dark brown shale and dark bluish grey concretionary shale. The Albian shales of Anderson Plain, northern Richardson Mountains, and Eagle Plain are similar and generally lie disconformably or with regional unconformity on strata as old as the Precambrian.

Mackenzie Trough. In Mackenzie Plain, between Franklin and Mackenzie Mountains, the Sans Sault Group lies unconformably on Devonian strata. It varies from 1,000 to almost 4,000 feet in thickness and consists of a basal sandstone, a middle shaly member, and an upper sandstone, all of Albian age. The overlying Slater River Formation is about 1,500 feet of marine shale, and is succeeded by the Little Bear Formation of conglomerate, sandstone, and coal, tentatively considered to be late Albian.

Lower Cretaceous of Rocky Mountain Exogeosyncline

Pre-Albian (Fig. VIII-35). Lowermost Cretaceous strata, dominantly alluvial sandstones, and shales in southern Alberta, grade northwesterly into marine mudstones and siltstones near Peace River and transitionally overlie the Jurassic Fernie Formation (Fig. VIII-41). They are bevelled in an easterly and northerly direction by a regional erosional unconformity beneath the Albian and do not extend far beneath the plains nor beyond Muskwa River in the Foothills (Figs. VIII-42, 43). The unconformity represents a major erosional interval related to epeirogenic uplift during the early phase of Columbian Orogeny.

Although partly Jurassic, the Kootenay and Nikanassin Formations probably include Lower Cretaceous strata in the more westerly and thickest sections. In southern Rocky Mountains near Fernie and Banff, coal-bearing beds and thick conglomerates with chert and quartz clasts occur at the top of the Kootenay. Carbonaceous sediments characterize the upper Nikanassin in the vicinity of Smoky River (Irish, 1951). The Minnes Formation, between Smoky and Pine Rivers, comprises sandstones, shales, and carbonaceous mudstones, 6,000 feet thick (Ziegler and Pocock, 1960). The lower part includes much quartzose sandstone considered to be marine and late Jurassic in age. The upper part, including a thick succession of coal-bearing beds similar to those of the Kootenay and Nikanassin, appears to be dominantly non-marine and contains early Lower Cretaceous spores.

In the Carbon Creek synclinorium between Pine and Peace Rivers and to the north, almost to Prophet River, the Jurassic Fernie shales and siltstones grade upward into quartzose marine sandstone of the Monteith Formation (Mathews, 1947), which is dated as late Jurassic to early

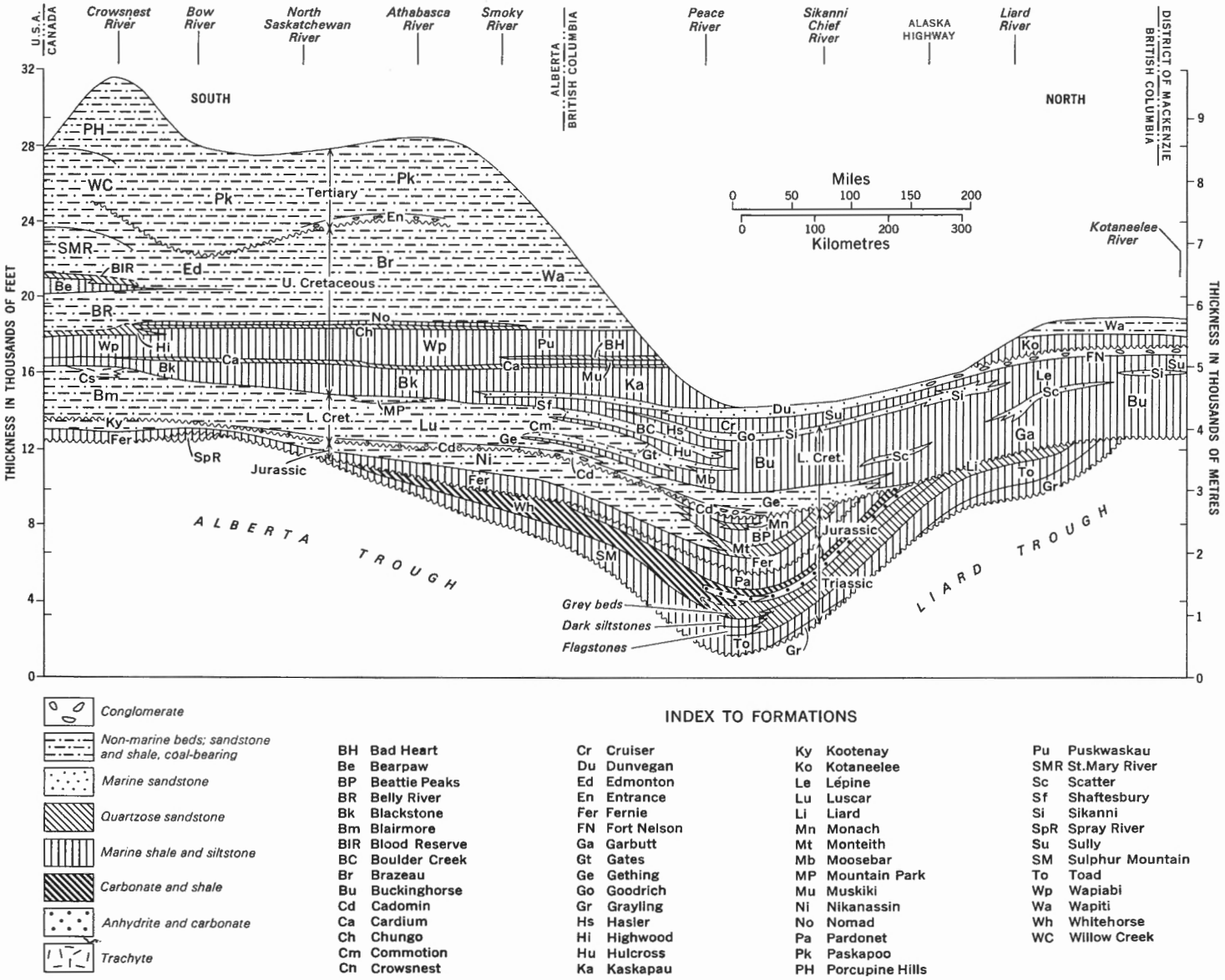

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FIGURE VIII-41. Restored longitudinal stratigraphic section of Mesozoic and Tertiary rocks along the Foothills of Alberta and British Columbia (by D. F. Stott).

Valanginian. Those beds and the overlying Beattie Peaks Formation are considered to be equivalent to most of the Minnes strata north of Smoky River. The Monteith sandstones, best developed near Peace River, grade northwesterly into marine shales and siltstones similar to the overlying Beattie Peaks. The latter contains a succession of silty marine mudstones with many channel sandstones and a Valanginian fauna of *Buchia*. The middle to late Valanginian Monach Formation contains medium- to coarse-grained sandstones near Pine River which are not present north of Peace River but may be represented by some fine-grained sandstones. The uppermost beds of the succession, unnamed, include interbedded mudstone, fine-grained sandstone, and locally, thin coal seams. The southern limit of the dominantly marine, lowermost Cretaceous sediments lies along the line of Peace River Arch.

Albian (Fig. VIII-36). The Blairmore Group of the southern Foothills of Alberta is 2,500 feet thick in the west, thinning to 1,300 feet in the east. It is 6,500 feet thick in Fernie Basin in southern Rocky Mountains. The group comprises a thick wedge of non-marine detrital strata that is divisible into three formations (Mellon, 1967). The Gladstone Formation, 250 feet in more easterly exposures, comprises a widespread basal siliceous conglomerate of quartzite, chert, and argillite, or a quartzose sandstone locally termed Dalhousie sandstone, a middle unit of non-marine, dark grey, green, and red shales interbedded with calcareous siltstone, and fine-grained sandstone, and an upper "calcareous" member of dark grey, calcareous shale and silty limestone with abundant freshwater pelecypods, gastropods, and ostracods. The middle Blairmore, or Beaver Mines Formation, consists of 800 to

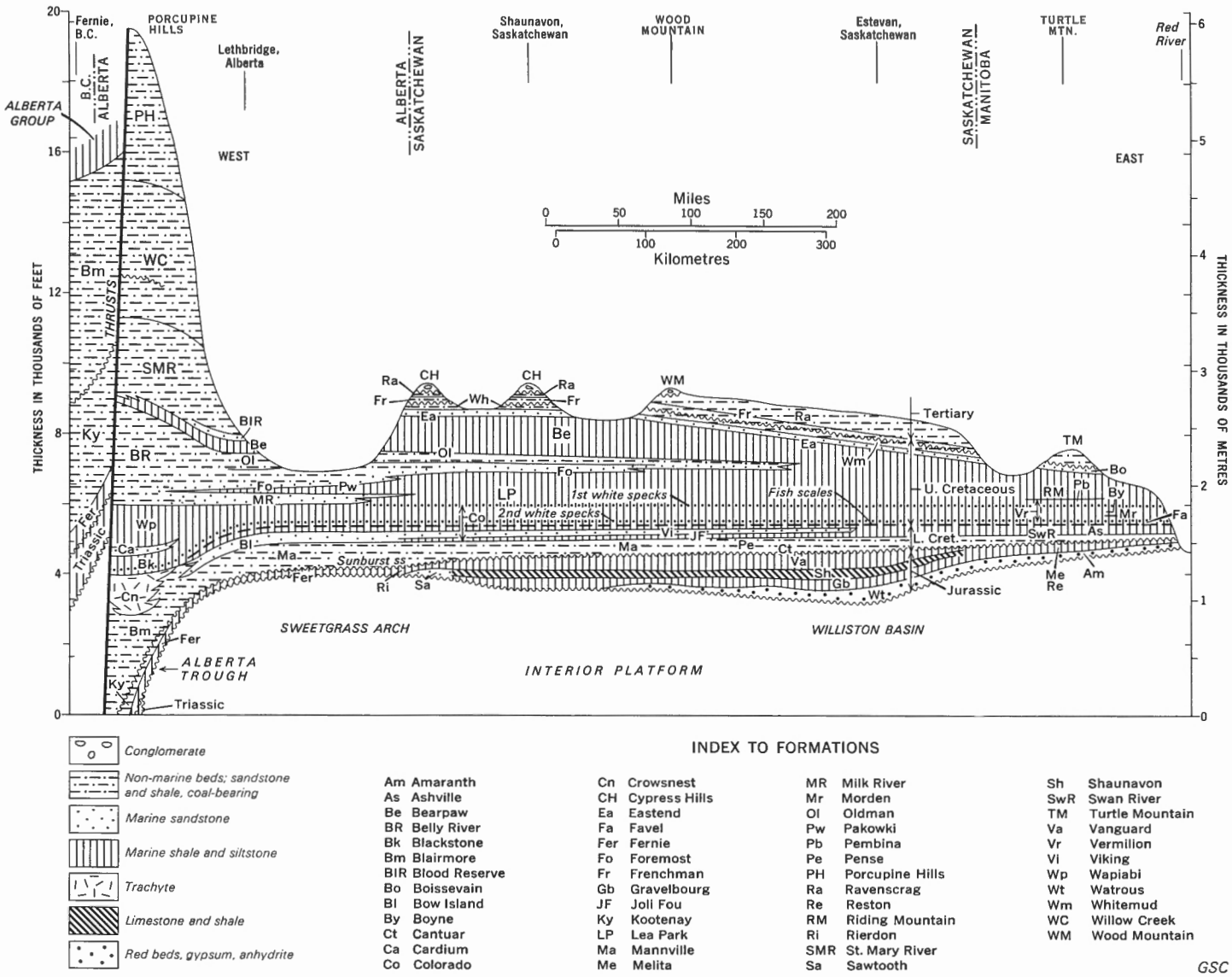


FIGURE VIII-42. Stratigraphic section of Mesozoic and Tertiary rocks from Fernie, British Columbia, across southern Alberta and Saskatchewan to Manitoba (by D. F. Stott).

1,200 feet of non-marine, varicoloured shales and green feldspathic sandstone, such as the local McDougall–Segur, that contain abundant volcanic detritus and chlorite cement, and conglomerates with igneous pebbles derived from the unroofing of the batholiths of the Omineca Geanticline to the west. A non-dicotyledonous flora, termed the Lower Blairmore flora, from Gladstone and Beaver Mines strata heretofore has been considered Aptian but recently has been found to extend upward into rocks of middle Albian age. The upper beds of the Blairmore Group in the southern Foothills differ from the lower ones, both in composition of the sandstone and in floral content. In the Crowsnest Pass region, the Mill Creek Formation, 580 feet thick, comprises quartzose cherty sandstone and green and red mottled shale that grade upward into trachytic tuff and agglomerate of the Crowsnest Formation. Near Coleman, the pyroclastic beds attain a maximum thickness of 1,600 feet but thin to the north and east, becoming interbedded with and grading into the cherty quartzose sandstone and varicoloured shales of the Mill Creek. Sanidine from the Crowsnest Formation has yielded a K–Ar date of 96 m.y. A predominantly dicotyledonous flora comprises the Upper Blairmore or Mill Creek flora. It occurs in both the sedimentary and pyroclastic beds, and is of Albian age.

In the Foothills between North Saskatchewan and Smoky Rivers, strata equivalent to the Blairmore are included in the Cadomin, Luscar, and Mountain Park Formations, which are equivalent to the Gladstone and Beaver Mines Formations of Mellon (1967). In that region, the Blairmore equivalents total 900 to 2,000 feet in thickness. The basal Cadomin conglomerate, 35 to 40 feet thick, is composed of well-indurated pebbles of quartz, chert, quartzite, and argillite. It is overlain by the lower

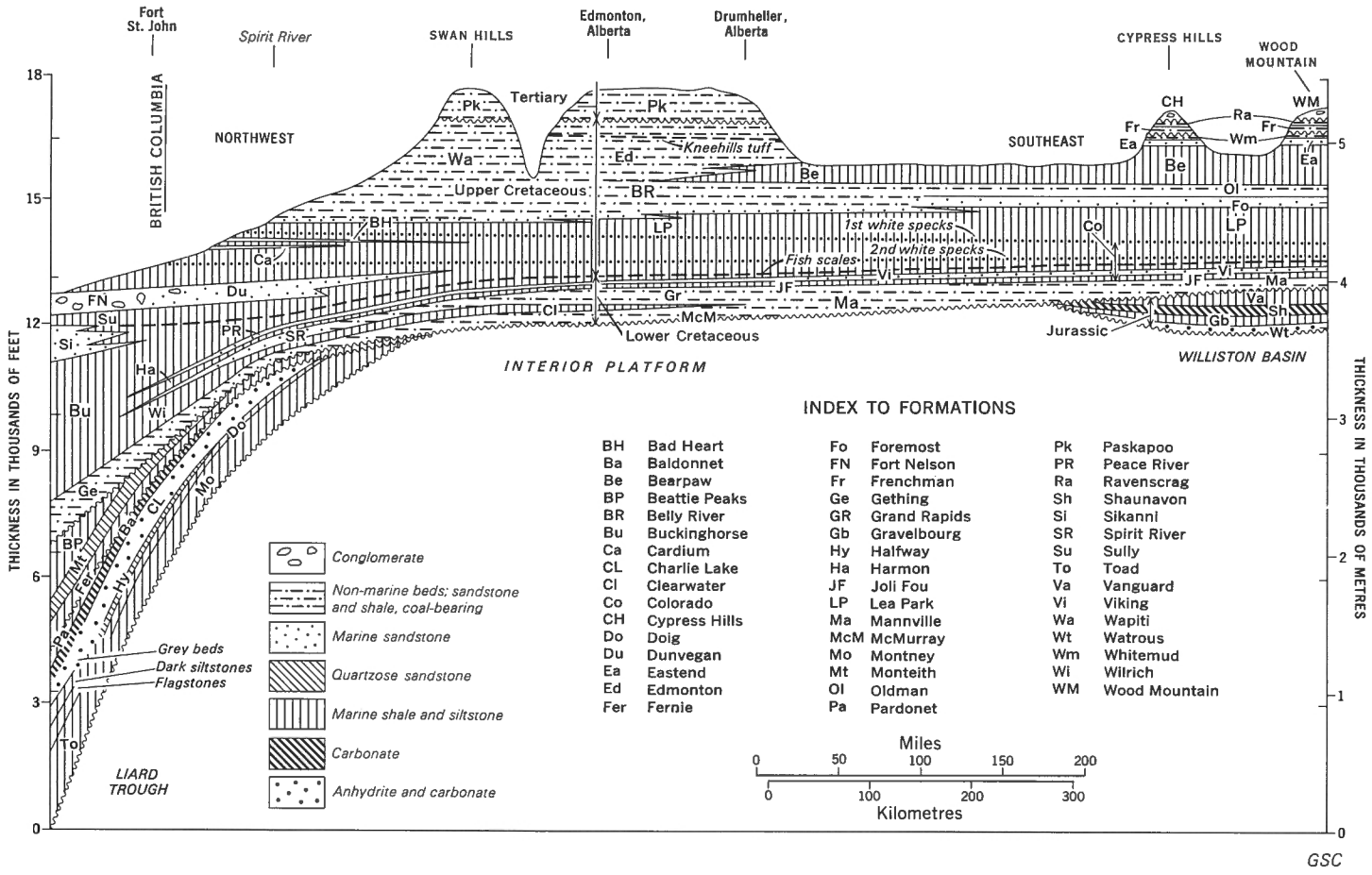


FIGURE VIII-43. Stratigraphic section of Mesozoic and Tertiary rocks from Fort St. John, northeastern British Columbia, southeastward to Saskatchewan (by D. F. Stott).

Luscar Formation equivalent of the Gladstone Formation, comprising thin-bedded, dark grey, carbonaceous shale, siltstone, and fine-grained sandstone, with 15 to 20 feet of dark grey shale with ostracods at the top. Basal Beaver Mines equivalents include 20 feet of dark grey concretionary marine shale with foraminifera correlative with that of the Moosebar Formation and Clearwater shales. The remainder of the Beaver Mines Formation consists of two facies: one previously referred to as the upper Luscar Formation, characterized by pale to medium grey rocks, cemented by illite or kaolinite and quartz, and containing commercial coal beds; and the other, the Mountain Park Formation, of green to greenish grey shales and sandstone, cemented in part by chlorite and locally conglomeratic. Beds equivalent to the Mill Creek Formation are absent in the central Foothills, owing to non-deposition or erosion.

Between Smoky and Peace Rivers, beds equivalent to the Blairmore Group are represented by sediments of the Bullhead and Fort St. John Groups, their combined thickness more than 6,000 feet (Stott, 1968). The Cadomin Formation, consisting of piedmont gravels and conglomerates, interfingers eastward with sediments of the alluvial plain, represented by the Gething Formation. The overlying marine Fort St. John shales were deposited conformably or with only a minor hiatus. The intertonguing of near-shore sandstones of the Commotion Formation with alluvial plain sediments on the one side and marine shales on the other shows that the sea was relatively shallow. The Cadomin conglomerate attains a maximum thickness of 600 feet and grades laterally northward and eastward into fine marine sandstones and argillaceous siltstones. The Gething Formation, about 1,800 feet thick on Peace River, thins rapidly eastward. It contains brown siltstones, silty and carbonaceous sandstones, grey to brown carbonaceous mudstones, and some commercial coal. The overlying Moosebar shale, merging northward with the Buckinghorse Formation, represents a major marine incursion. It thins southward to Smoky River, and is replaced by the chert sandstones of the Gates and lower Commotion Formations. The Commotion Formation contains numerous coal seams, chert sandstones, and carbonaceous shales. The lower Commotion or Gates Formation is dated as early to middle Albian and also contains the Lower Blairmore flora. The dark grey, silty marine shales of the Hulcross Member of the Commotion are overlain by and are in part laterally equivalent to the Boulder Creek deposits of well-sorted chert sandstone, carbonaceous sediments, and conglomerate deposited on an alluvial deltaic plain. The upper Commotion, with the

middle Albian ammonite *Gastroplites*, contains an angiosperm flora possibly slightly older than the typical Upper Blairmore flora. The northeasterly trend of marine and non-marine facies boundaries is subparallel with the trend of faults bordering the Devonian Peace River Arch which suggests differential subsidence in the depositional basin. Marine sediments deposited during late Albian time are the dark grey, concretionary marine shales included in the Hasler Formation, the fine-grained marine sandstones of the Goodrich Formation, and the dark grey marine shales of the Cruiser Formation. The Goodrich sandstones disappear eastward and southward resulting in one continuous succession of dark grey, concretionary, silty, marine shales known as the Shaftesbury Formation. The shales overlap the non-marine beds southeastward along the Foothills. Their thickness correspondingly decreases until they are not represented in the southernmost Foothills.

Lower Cretaceous rocks between Tetsa and La Biche Rivers lie unconformably on Triassic to Mississippian sediments (Stott, 1968b). They comprise a thick succession of intertonguing deltaic and marine sandstone and shale and grade transitionally upward into the alluvial and deltaic conglomeratic sandstones of the Dunvegan Formation. The basal Garbutt dark grey, concretionary and silty marine shales, about 900 feet thick, are succeeded by the Scatter Formation, a 1,200-foot wedge comprising two major units of fine-grained, glauconitic sandstones separated by a thick shale member. The Scatter is approximately equivalent to the Commotion Formation farther south, including and overlain directly by beds containing the middle Albian *Gastroplites* fauna. The Scatter Formation is overlain transitionally by 3,000 feet of concretionary shales and thin argillaceous siltstone of the Lepine Formation. The Sikanni Formation of fine-grained nearshore sandstone is about 1,000 feet thick at Muskwa River but grades northward and eastward into silty shale and platy grey siltstone. The overlying Sully Formation comprises 1,000 feet of silty mudstone with sideritic concretions, fissile black shales, and mudstone with interbedded silty sandstone and siltstone.

The Albian succession in the Peace River region is similar to that of the Foothills to the west (Rudkin, 1964). The Cadomin extends eastward as a thin fine pebble-conglomerate or coarse-grained sandstone. The Gething coal-bearing beds are much reduced in thickness and the upper part grades laterally into glauconitic marine sandstones and shales of the Bluesky Formation. Within the overlying Spirit River Formation, the basal Wilrich marine shales gradually replace alluvial and deltaic sandstones and shales of the Falher Member. The upper Notikewin Member of well-sorted sandstone is equivalent to upper beds of the Gates Formation. The Harmon shales of the Peace River Formation represent the late middle Albian marine transgression. Well-sorted sandstones of the Cadotte and the coarse-grained sandstone and carbonaceous sediments of the Paddy Member are part of the nearshore and alluvial deltaic plain sediments associated with the transgression. They are overlain by 400 to 900 feet of Shaftesbury marine shale.

In the central and southern Interior Platform, the Albian comprises, in ascending order, the Mannville Group, Joli Fou Formation, Viking Formation, and an unnamed shale beneath the Fish Scale beds which are considered to mark the base of the Upper Cretaceous. All these beds above the Mannville are commonly included in the lower part of the Colorado Group.

The Mannville Group, ranging from 200 to 1,000 feet in thickness, has recently been subdivided in central Alberta into the McMurray and Fort Augustus Formations (Mellon, 1967). The McMurray Formation lies unconformably on Mississippian and Devonian carbonates over most of the central Alberta Plains, overlapping older Cretaceous or Jurassic detrital strata in the western part of the region. The thickness, controlled by relief on the underlying erosion surface, varies from zero to about 400 feet. The McMurray comprises three members: the basal or Deville Member, not always present, is a residual detritus formed of red and green waxy shale, chert fragments, dark brown shale, siltstone, and sandstone; the middle or Ellerslie Member consists of dark grey, silty shale, siltstone, and white kaolinitic quartz sandstone; and the upper or "calcareous" member of hard dark grey, calcareous shale and siltstone with abundant ostracods. The Fort Augustus Formation can be traced into the Beaver Mines Formation of the Foothills. The lower or Wabiskaw Member of the Fort Augustus is marine, dark grey, glauconitic shale and fine-grained, glauconitic sandstone, containing a Clearwater foraminiferal assemblage. The remainder of the Fort Augustus is of shoreline and non-marine origin, consisting of laminated shale and siltstone, and fine- to medium-grained, feldspathic sandstone, and containing a "Lower Blairmore" megaflora. The marine beds thicken to the north, grading into the Clearwater and Loon River Formations, whereas the overlying non-marine beds thin and grade laterally in the northern plains into shoreline beds, such as the Grand Rapids and Notikewin sandstones. In east-central Alberta, the basal quartzose sandstone of the Mannville was previously termed the Dina Member. It is overlain by the Cummings Member, a marine shale that includes equivalents of the Ostracod zone. In that area, the upper Mannville was divided into Islay, Tovell, Borradaile, and O'Sullivan Members on the basis of small differences in sandstone composition (Nauss, 1945).

In northeastern Alberta, the McMurray sediments were derived from the Canadian Shield to the east. They consist of 100 to 200 feet of well-sorted quartzose sandstone that is impregnated with heavy oil over a wide area, and constitutes the Athabasca Oil Sands (Fig. IX-29). A thin shale, equivalent to the Ostracod zone separates the basal beds from the glauconitic Wabiskaw sandstone. The overlying Clearwater marine shales are transitionally succeeded by carbonaceous sandstones and shales of the Grand Rapids Formation. These upper Mannville sand-

stones, in the Cold Lake area of Alberta, contain heavy hydrocarbons, transitional in properties between the bitumen of the Athabasca Oil Sands to the north and heavy crude oil of Lloydminster area to the south.

The Joli Fou marine shales, 60 to 110 feet thick, are widely distributed throughout the Interior Plains but disappear westward into the non-marine beds of Mill Creek Formation of the Blairmore Group. The Viking sandstone, 50 to 100 feet thick, grades eastward into siltstone and shale and into the marine shales of the Ashville Formation in Manitoba. The Viking is correlated also with the Pelican sandstone of northeastern Alberta. In southwestern Alberta, the Viking sandstones grade into dark marine shales and sandstones of the Bow Island Formation that attains a thickness of more than 400 feet. Those beds and tuffaceous beds with heulandite in the extreme southwest are equivalent to the Mill Creek or upper Blairmore and Crowsnest strata of the Foothills. The relationship of the Joli Fou and Viking Formations of the Plains with the Peace River Formation is controversial. Although stratigraphic evidence strongly favours their correlation, micropaleontological evidence suggests that the Viking and Joli Fou are slightly younger.

In southeastern Saskatchewan, the non-marine Cantuar and Pense Formations are overlain by the Ashville Group. Coarse-grained sandstones and accumulations of chert detritus occur locally at the base. In southwestern Saskatchewan the Roseray quartzose, fine- to medium-grained sandstones lying unconformably on Jurassic sediments may be of late Jurassic or early Cretaceous age. Sandstones of the Cantuar, ranging from greywacke to quartzose sandstone, are interbedded with silt, grey and green shale, coal, and variegated clay. The Pense comprises fine-grained sandstone with some interstitial white clay and interbedded silt and dark grey shale. The Ashville succession of grey marine shale is divided by the Viking sandstone into the Joli Fou shale below and an unnamed shale above. Two other sands are recognized, the Spinney Hill near the base and the widespread but lenticular Fish Scales near the top. The latter beds, although easily recognized on electric and gamma-neutron logs, have only minor lithological expression, as they are composed of shale with some sand and fish remains.

In Manitoba, the Swan River Group includes all the sandstone in the lower part of the succession. The basal part consists of discontinuous fine- to coarse-grained quartzose sandstones and light coloured clay. Shaly glauconitic sandstone, presently included in the upper part in the Swan River region, appears equivalent in part to the Viking and Joli Fou Formations of the Ashville Group farther south and west.

The latest Albian marine shales are included in the Shaftesbury Formation of Peace River Plains, the lower Colorado Group of the western Plains, and the upper Ashville Group of Manitoba. The boundary between Lower and Upper Cretaceous beds is taken as the base of the Fish Scale Sand which probably coincides with the uppermost range of the late Albian ammonite *Neogastroplites*.

Upper Cretaceous of Western Cordilleran Geosyncline

Pacific Continental Shelf. On eastern Queen Charlotte Islands the Lower Cretaceous Haida Formation is succeeded, generally conformably, by the coarse clastic Honna Formation that contains granite-bearing conglomerates and by the shales of the Skidegate Formation. These sediments totalling 6,000 feet, and a granite-bearing conglomerate near Bella-Bella, were deposited in Hecate Basin which probably extended southeast of Queen Charlotte Islands, and formed slightly earlier than Georgia Basin.

The Nanaimo and Comox Groups in Georgia Basin on Vancouver Island consist of marine and intertonguing marine, shoreline, and coal-bearing non-marine clastic sediments that were derived from the erosion of the mountains in central Vancouver Island (Buckham, 1947). The rapidly subsiding basin formed between Coast and Insular Geanticlines, on the site of the present Strait of Georgia and western Coast Mountains. Non-marine tongues are typically represented by conglomerate and sandstone units, and the marine facies by shales. The sediments, which range from late Santonian to late Maestrichtian in age, lie with angular unconformity or nonconformity upon older rocks and Jurassic plutons.

In the eastern part of Georgia Basin, in the Mount Garibaldi region north of Vancouver, a conglomeratic basal unit, the Cheakamus Formation, containing Campanian marine fossils rests unconformably on altered quartz diorite of the Coast Plutonic Complex (Mathews, 1958). The total thickness of the Cheakamus and overlying Empetrum and Helm Formations may be more than 20,000 feet. They occur in a fault block and are intruded by later phases of the Coast Plutonic Complex. Near Vancouver the base of the Burrard Formation is a conglomerate, with granitic boulders, which lies unconformably on the Coast Plutonic Complex. Fossil plants and microflora from the lower part of the Burrard Formation are mainly Campanian, whereas the upper part is possibly early Tertiary. In a drillhole on Fraser Delta the Burrard Formation is 14,000 feet thick and lies on granitic rocks.

Columbian Zwischengebirge. The lower part of the Ootsa Lake Group which lies unconformably on all older rocks may contain Upper Cretaceous strata. Freshwater molluscs from the base of the andesitic unit near Ootsa Lake may be Upper Cretaceous age although associated fossil plants are regarded as Paleocene (Tipper, 1963).

The Sustut Group comprises non-marine clastic sediments with abundant granitic detritus derived from the Omineca Intrusions to the northeast (Lord, 1948). In places the group is separable into a lower division of conglomerate and pebbly sandstone and an upper division of sandstone and shale, with several thin bands of white-weathering dacitic tuff near the base. It forms a northwest-trending belt and was apparently deposited in a broad

intermontane valley or trench. The group has yielded fossil plants probably of Cenomanian age and others that are Paleocene. It lies unconformably on older rocks in the valleys but occurs also as upland remnants in northern Omineca Mountains.

The Hutshi Group in southwestern Yukon Territory overlies Lower Cretaceous and older strata with marked angular unconformity and is intruded by granodiorite and pink granophyric quartz monzonite. The group comprises gently folded flows and flow breccias that range from basalt to rhyolite, with agglomerate, tuff, and minor clastic sedimentary rocks. Conglomerates and breccias commonly occur at the base. Locally, the group is more than 3,000 feet thick.

Eastern Columbides. Near Rossland in the Monashee Mountains, the Sophie Mountain Formation of quartzitic conglomerate and minor argillite contains fossil plants of Cenomanian to Campanian age. The formation is not observed in contact with other units but probably rests unconformably on older rocks. It contains no fragments of the nearby Nelson batholith and the source of the quartzite pebbles is unknown.

The Upper Cretaceous and Paleocene Sifton Formation and related rocks in northern Rocky Mountain Trench comprise conglomerate, sandstone, shale, and minor coal. Most of the clasts in the conglomerate are limestone with some sandstone, quartzite, chert, dolomite, and argillite that appear to have been derived from the east. In places the conglomerates become coarser and more abundant towards the east. Some with coarse, angular clasts along the eastern margin of outcrop may be in part of fanglomeratic origin.

In Cassiar Mountains, two northwesterly trending trench-like valleys bounded by major faults are underlain by gently dipping sandstones, conglomerates, and shales, variably carbonaceous, and locally containing seams of coal. Near Turnagain River a flora, dated as Santonian–Campanian, has been found.

Upper Cretaceous of Northern Cordilleran Geosyncline

The Eagle Plain Formation lies unconformably on Albian shales and the Mississippian. It consists of 2,000 feet of fine- to medium-grained sandstone, light brownish grey and arkosic, with crossbedding indicating a generally northeasterly direction of sediment transport (Mountjoy, 1967b). These strata are overlain by 1,200 feet of silty shale with thin interbeds of siltstone and sandstone and plants of Cenomanian age *Inoceramus* ex gr. *dunveganensis* have been collected.

Soft, dark grey shales of Cenomanian and Turonian age occur in northern Richardson Mountains (Jeletzky, 1961). They are about 1,000 feet thick, weather orange or bluish grey, and rest unconformably on Aptian sandstone. The shales are the youngest rocks cut by dextral wrench faults bounding Richardson Mountains, some displacement having taken place during the Laramide Orogeny. Grey marine Turonian shales occur in a fault slice at Fort Norman in Franklin Mountains.

In northern Yukon Territory and northwestern District of Mackenzie late Upper Cretaceous and early Tertiary non-marine clastic strata are present in several small areas. These strata may be of various ages as most available plant material is not diagnostic, but all lie with angular or regional unconformity on deformed or disturbed strata and are deformed to various degrees as the result of the late phase of Laramide Orogeny in early Tertiary. They are considered to be post-orogenic or synorogenic. It is improbable that the remnants were part of a once continuous sheet of sediments, although those of Arctic Coastal Plain may be, and, accordingly, would grade offshore into marine equivalents on the Arctic Continental Shelf and the Mackenzie Delta. The Monster Formation, in Ogilvie Mountains, is composed of 4,000 feet of dark brownish grey, medium- to coarse-grained, crossbedded arkosic sandstone (Mountjoy, 1967b). Thick beds alternate with covered intervals that are presumed to be argillaceous rocks. Some beds are conglomeratic, the pebbles of quartz and chert being well rounded and as large as 3 inches in diameter. Upper Cretaceous plants are present in the lower half. These are the youngest rocks in the region and rest with regional unconformity on the Triassic and Permian. The lithologically similar Bonnet Plume Formation on Peel River rests with angular unconformity on deformed lower Paleozoic strata. It may lie in a fault-bounded basin. The lignite, with late Cretaceous or early Tertiary spores, lies near the top of the formation which has a maximum thickness of 5,000 feet. The Moose Channel Formation west of Mackenzie Delta includes a 7-foot, sub-bituminous coal seam, utilized locally. The formation is more than 200 feet thick and consists mainly of feldspathic, coarse-grained sandstone.

Upper Cretaceous of the Rocky Mountain Exogeosyncline

Upper Cretaceous strata immediately underlie much of the Interior Platform, extending from within the Foothills almost to the Canadian Shield (Figs. VIII-37, -44). At the western limit of their exposure, they attain a maximum of 12,000 feet but thin rapidly to less than 4,000 feet in the east (Figs. VIII-42, -43). The sediments, mostly sandstone and shale, were derived mainly from late Columbian and early Laramide uplifts in the Cordilleran Geosyncline; a small amount of detrital material was probably contributed by the Shield. The rocks of Cenomanian to Santonian age are, except for the deltaic deposits of the Dunvegan Formation, almost entirely marine and constitute the Alberta, Smoky, and Colorado Groups (Fig. VIII-41). Extensive alluvial sediments deposited after mid-Campanian time are found in the Belly River and Edmonton Formations.

The Alberta Group extends along the Foothills from Montana to Athabasca River (Stott, 1963). Within it, the Blackstone and Wapiabi Formations, consisting of

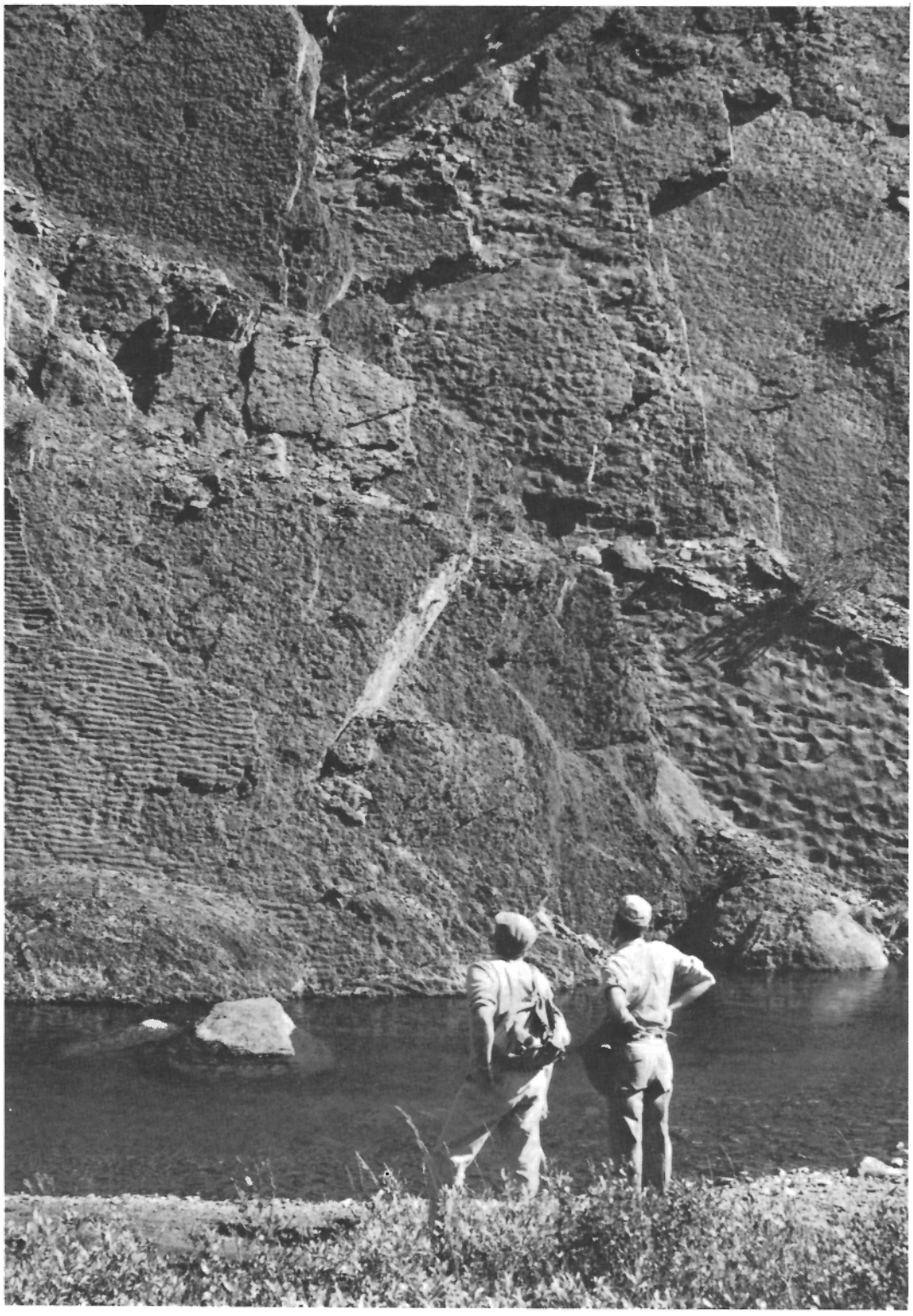


PLATE VIII-14

Ripple-marks on several bedding surfaces on sandstones of Upper Cretaceous Sunkay Member, Blackstone Formation, Burnt Timber Creek, Foothills of Alberta.

marine shales, are separated by arenaceous epineritic and shoreline beds of the Cardium Formation. The Blackstone Formation is characterized by fossil zones of Cenomanian to Turonian age. The basal Sunkay Member, which includes the Fish Scale beds in the central Foothills, overlaps the Blairmore Group in a southerly direction, disappearing almost completely in southwestern Alberta where the Blackstone disconformably overlies the Crowsnest volcanic rocks. The Sunkay Member, representing a marine transgression, has a thin layer of chert pebbles at the base and

contains sideritic, rusty weathering, silty shale, mudstone and sandstone (Pl. VIII-14). The Vimy Member, developed during the inundative phase, contains shales that are calcareous to dolomitic, gypsiferous, and pyritic. The Haven and Opabin Members, formed by regressive sediments similar to the transgressive deposits, grade upward into thick-bedded, fine- to medium-grained sandstone. In the central Foothills, the relatively thin Cardium Formation contains littoral marine beds of thick-bedded, well sorted, fine-grained cherty sandstone; the basal unit extends eastward beneath the Plains and at Pembina oil field forms one of the largest reservoirs in Canada (Fig. IX-24d). The Cardium also includes marine shales and lagoonal and marsh deposits of carbonaceous shales, siltstones, argillaceous sandstone, and some coal. The Cardium sandstones grade laterally eastward into shaly siltstones and shales. The overlying marine shales of the Wapiabi Formation include several zones of *Scaphites* ranging from Coniacian to Santonian. The transgressive deposits of rusty weathering sideritic shales and siltstones are included in the Muskiki, Marshybank, and Dowling Members. The inundative facies is represented by calcareous shales of the Thistle Member. A major regression is recorded in the blocky to rubbly, silty mudstones of the Hanson Member and in the overlying fine-grained sandstone to argillaceous siltstone of the Chungo Member. A non-marine facies of greyish green shales with thin coal beds occurs in the Chungo at a few places along the western Foothills. A short-lived transgression, probably related to Lea Park of the Plains, is represented by dark grey shales, thin siltstones, and sandstones of the Nomad Member. South of Highwood River, beds equivalent to the upper three members of the Wapiabi contain much sandstone and are included in the Belly River Formation of the Foothills.

North of Athabasca River, some shales and siltstones of the Sunkay Member of the Blackstone Formation grade into massive sandstone that constitutes the Dunvegan Formation. The overlying marine strata form the Smoky Group and the marine shales between the Blairmore Group and the Dunvegan Formation are the Shaftesbury Formation (Stott, 1967). The Dunvegan Formation contains interbedded marine and non-marine shales and sandstones of Cenomanian age that thicken northward to more than 1,200 feet. Several major sandstone lobes extend from the Peace River Foothills into the Plains of British Columbia and Alberta forming a complex delta. The formation averages about 500 feet in thickness north of Peace River and becomes conglomeratic.

The Smoky Group is divided into the Kaskapau, Bad Heart, and Puskwaskau Formations. Along the Foothills, the Kaskapau contains much siltstone in the lower and middle parts that grades northward near Peace River into several prominent sandstones. These near-shore sandstones may be related to the underlying Dunvegan Formation and may also be associated with a broadening of the uplift of the central part of the Omineca Geanticline to the west. Sideritic shales of the upper Haven and Opabin Members of the Kaskapau grade upward into epineritic sandstones of the Cardium Formation. The Cardium Formation is present in the Foothills below the Bad Heart sandstone and the intervening shale is called Muskiki. The Cardium Formation grades laterally eastward into silty shales and is represented on lower Smoky River only by a pebble bed. The Bad Heart Formation of the Foothills contains well-sorted sandstone similar to that of the Cardium. In the Plains, it is highly glauconitic and is succeeded by sideritic and calcareous shales of the Dowling and Thistle Members of the Puskwaskau Formation. The upper beds of the Puskwaskau are transitional with overlying coarse-grained sandstones and coal-bearing beds of the Wapiti Formation. As in the Wapiabi Formation of the southern Foothills, the uppermost marine shales of the Puskwaskau intertongue westward with arenaceous sediments, grading laterally into non-marine strata of the Wapiti Formation.

In the Liard River region, the Dunvegan Formation, formerly referred to as the Fort Nelson Formation, consists of 500 to 600 feet of massive conglomerate, coarse-grained sandstone, carbonaceous shales, and a few thin coal seams (Stott, 1960). A regional disconformity, representing the Turonian stage, separates it from the overlying Kotaneelee marine shales. The hiatus is possibly related to uplifts associated with the late Columbian Orogeny, the intrusion of the Cassiar batholith, and the granitic bodies in Yukon Territory (Fig. VIII-37). The Kotaneelee dark grey, sideritic shales, exposed in only a few localities, grade upward into the Wapiti Formation of dominantly continental sandstones, shales, and minor coal.

In the Plains, marine beds above the Mannville Group are generally assigned to the Colorado Group, the lower part of which is Lower Cretaceous (Williams and Burk, 1964). Equivalent beds are the Lloydminster shale in eastern Alberta and the La Biche shales of northeastern Alberta. They are dark grey, bentonitic and concretionary shales, and include the Fish Scale marker horizon toward the base and two units of calcareous shale, termed the Second (Lower) and First (Upper) White Specks. In southern Alberta, the Colorado shales grade upward into well-sorted sandstones of the Milk River Formation that in turn are overlain by marine shales of the Pakowki Formation. The Milk River sandstones merge westward with the lower part of the Belly River Formation and, in east-central Alberta, grade into about 1,000 feet of grey marine shales constituting the Lea Park Formation. The Belly River Group of the Plains, averaging about 900 to 1,000 feet, comprises the Foremost and Oldman Formations (Russell and Landes, 1940). In eastern Alberta, nine members representing tongues of marine shales and deltaic sediments are recognized within strata equivalent to the Foremost (Shaw and Harding, 1954). Throughout most of southern Alberta and as far north as Edmonton, the Belly River is overlain by marine shale of the

Bearpaw Formation (Lines, 1963). Progressively younger shales occur at the top of the Bearpaw from west to east in southeastern Alberta and Saskatchewan, and the thickness of the formation ranges from less than 100 feet near the Foothills to as much as 1,200 feet in the east.

The Belly River Formation and its equivalents form a thick wedge of clastic sediments that extends from the Foothills into Saskatchewan. These continental beds, several thousands of feet thick in the west, thin to a few hundred feet in the east and grade laterally into marine rocks (Fig. VIII-42). In the southern Foothills, the Belly River Formation, about 2,000 feet of green sandstone and shale with some coal, is overlain by 800 feet of marine shales included in the Bearpaw Formation. These beds lie below 2,500 feet of non-marine, hard sandstone and shale of the St. Mary River Formation (Douglas, 1950). Well-sorted marine sandstone immediately overlying the Bearpaw is locally known as the Blood Reserve Formation. The Willow Creek Formation, lying transitionally above the St. Mary River is about 4,000 feet thick. The lower part comprises grey-weathering sandstone with interbedded varicoloured shales that grade into the upper part of massive buff-weathering sandstone and brown shale. No hiatus is recognized within the formation although the lower part contains Upper Cretaceous non-marine invertebrates and the upper part contains a typical Paleocene fauna (Tozer, 1956).

The Edmonton Formation of central Alberta, 1,100 to 2,500 feet thick, includes all Cretaceous non-marine beds equivalent to the Bearpaw shale, or lying above it. (Ower, 1960; Williams and Burk, 1964). It consists of soft sandstones, bentonitic and silty shales. Coal seams are common in the lower and upper parts and a marine limestone tongue, the Drumheller Member, occurs locally in the lower part. The Kneehills Tuff, the most distinctive and persistent marker, is in dark bentonitic shales underlain by white clay and sandstone near the middle of the formation. An abundant dinosaurian fauna has been obtained from the formation in the Drumheller badlands.

The marine shales of the Bearpaw Formation of southwestern and central Alberta are not recognized in the central and northern Foothills, and separation of Belly River strata from similar non-marine beds of the younger Edmonton Formation is not possible, the entire succession being included in the Brazeau or Wapiti Formations. The lower beds of the Brazeau, 800 to 1,500 feet thick, include quartz pebble sandstone and conglomerates. The remainder of the formation consists of about 5,000 feet of interbedded, medium- to coarse-grained, greenish grey sandstone with interbedded shale, some tuff, and thin coal seams. It is overlain by the Entrance basal conglomerate and thick deposits of Paleocene age. The Wapiti Formation, approximately equivalent to the Brazeau has been divided into five members (Allan and Carr, 1946). The equivalent to the Kneehills Tuff, a widespread time and stratigraphic marker, occurs near the base of the uppermost member.

In the Cypress Hills of southeastern Alberta and southwestern Saskatchewan, the Bearpaw shales are overlain by very fine grained marine sandstones of the Eastend Formation. The succeeding Whitemud, 15 to 75 feet thick, includes white to light grey refractory clays, siltstones, and sandstones. The Battle Formation contains dark grey to purplish bentonitic shale with the Kneehills Tuff occurring near the top. In southwestern Saskatchewan, the Battle, Whitemud, and Eastend Formations are bevelled beneath *Triceratops*-bearing beds of the Frenchman Formation which grade transitionally upward into the Paleocene Ravenscrag (Furnival, 1946). The Frenchman consists of about 200 feet of medium-grained sandstone with carbonaceous and bentonitic shales.

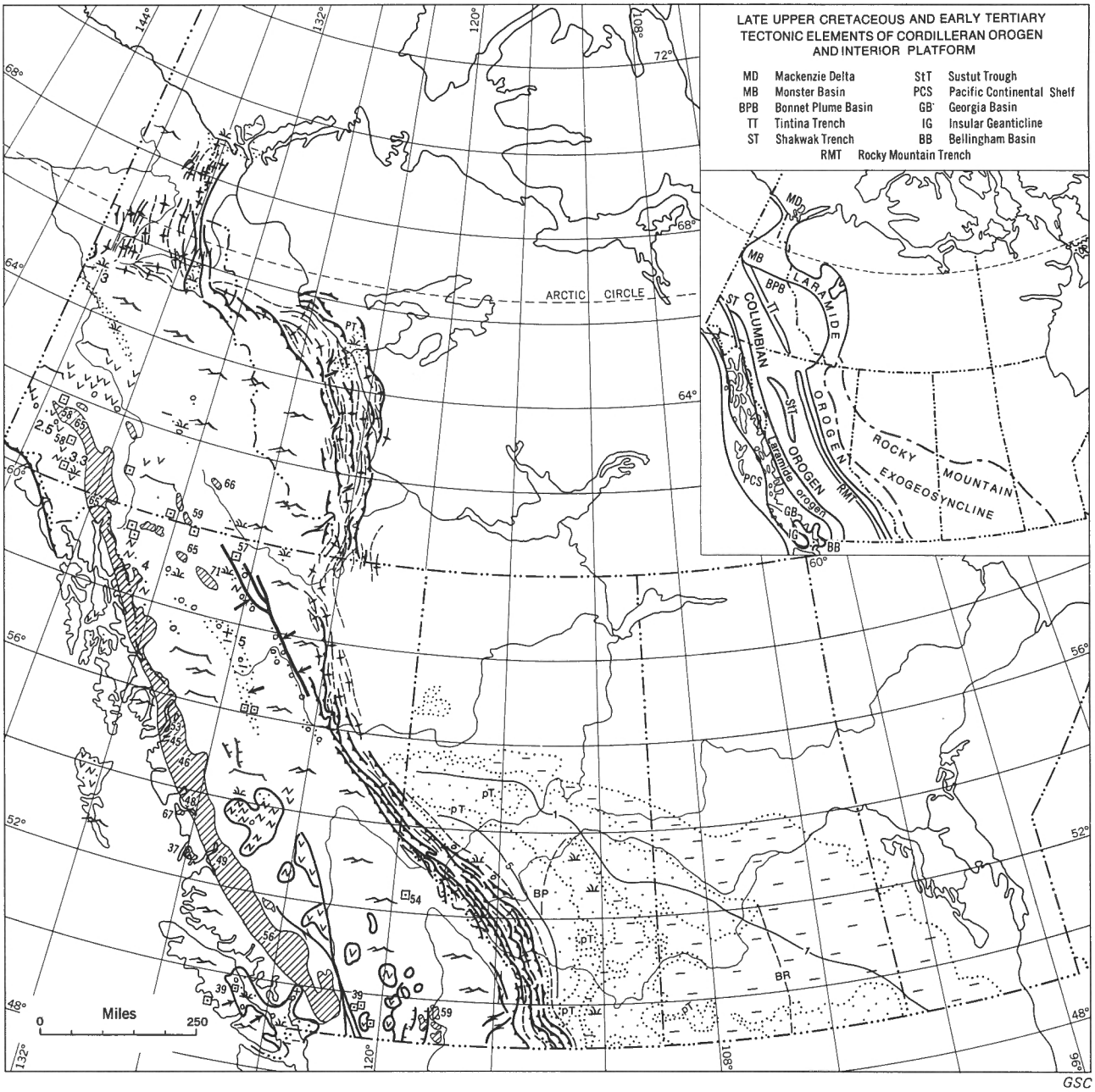
Upper Cretaceous rocks of eastern Saskatchewan and Manitoba, about 1,800 feet thick, are dominantly marine shales (Wickenden, 1945). The Fish Scale beds are in the Ashville Group which extends stratigraphically to the top of the Second White Specks shale. The latter forms the Keld Member of the Favel Formation. The Vermilion River Formation, lying above the Favel with apparent conformity, consists of the Morden, Boyne, and Pembina Members. The Boyne includes the First White Specks shale. The Pembina dark grey, non-calcareous shales are post-Colorado in age. The Riding Mountain Formation probably contains beds equivalent to most of the Milk River, Pakowki, Foremost, Oldman, and Bearpaw Formations. It consists of green clays, grey siliceous shales, and dark grey bentonitic shales, and ranges in thickness from 700 feet in southern Manitoba to 1,500 feet in southern Saskatchewan. The succeeding Boissevain Formation, about 100 feet thick, contains greenish grey sandstone.

CENOZOIC

Tertiary

Tectonic Summary

In the Tertiary almost all of western Canada was land. The mountain belts formed during Columbian Orogeny were rejuvenated and new mountain belts were produced by the Laramide Orogeny, the early phase of which began in the late Upper Cretaceous and the late phase ended probably early in the Oligocene (Fig. VIII-44). The principal effects of the orogeny were the development of most of the easternmost and northernmost structural elements of the Cordilleran Region: the linear Rocky Mountain Thrust Belt, the arcuate Mackenzie Fold Belt, and the Northern Yukon Fold Complex. The structures



DEPOSITIONAL FEATURES

- Conglomerate
- Granitic clasts
- Sandstone
- Shale
- Volcanic rocks
  - rhyolite
- Coal
- Depositional limit
- Outcrop limit,
  - Paleocene (pT)
- Facies boundary,
  - Belly River (BR)
  - Bearpaw (BP)
- Isopach (thousands of feet)
- Thickness (thousands of feet) 3.3
- Direction of sediment transport

LARAMIDE OROGEN

- Thrust fault
- Transcurrent fault
- Normal fault
- Fault
- Anticline
- Syncline
- Granitic plutons (age in millions of years) 59

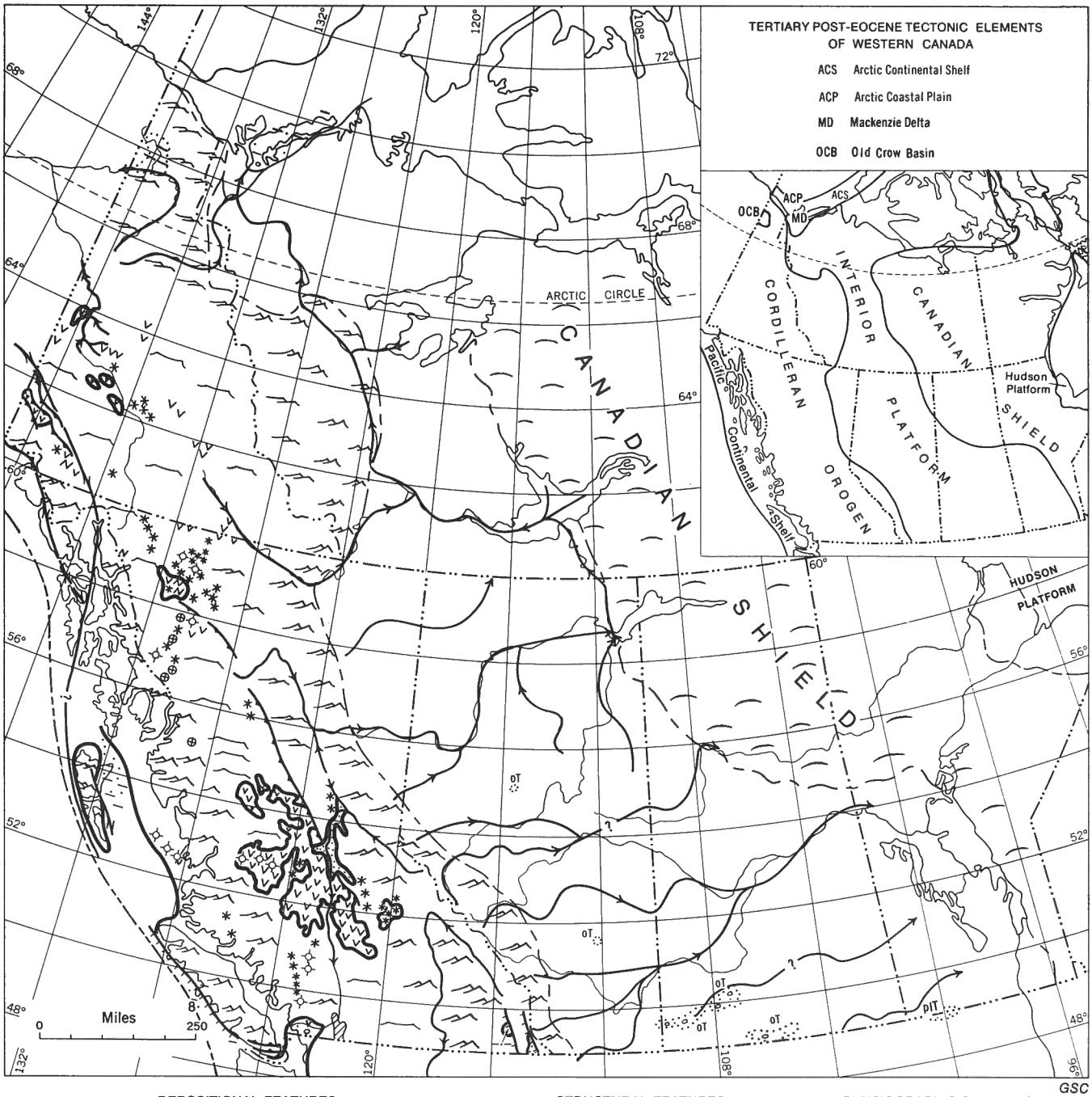
Note: Late Upper Cretaceous and Tertiary strata involved in Laramide structures are not shown in southern Laramide Orogen; on Vancouver Island the structures are not shown.

PHYSIOGRAPHIC FEATURES

- Low mountainous land
- Plateau

Blank areas are lowlands and plains developed on crystalline and sedimentary rocks of the craton, and where the Late Upper Cretaceous and Early Tertiary is now absent through erosion, or unknown. Parts of Laramide Orogen were mountainous, particularly the southern part from which much of the clastic sediment of the Rocky Mountain Exogeosyncline was derived.

FIGURE VIII-44. Late Upper Cretaceous and early Tertiary sedimentation and tectonism in Western Canada, and some features of the Laramide Orogeny.



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DEPOSITIONAL FEATURES

- Conglomerate
- Sandstone
- Siltstone
- Shale
- Volcanics
  - acidic
  - basic
- Volcanoes
  - General
  - Pleistocene
  - Recent
- Outcrop limit
  - Pliocene (pIT)
  - Oligocene (oT)
- Depositional limit
- Edge of continental shelf
- Thickness (thousands of feet) 8

STRUCTURAL FEATURES

- Transcurrent fault
- Normal fault
- Thrust fault
- Fault
- Granitic pluton

PHYSIOGRAPHIC FEATURES

- Province boundary
- Main pre-Pleistocene rivers (taken partly from Figure XII-1)
- Mountainous land (high, low)
- Plateau
- Crystalline lowland

Blank areas on the continent are plains.

FIGURE VIII-45. Tertiary, post-Eocene, sedimentation and tectonism in Western Canada.

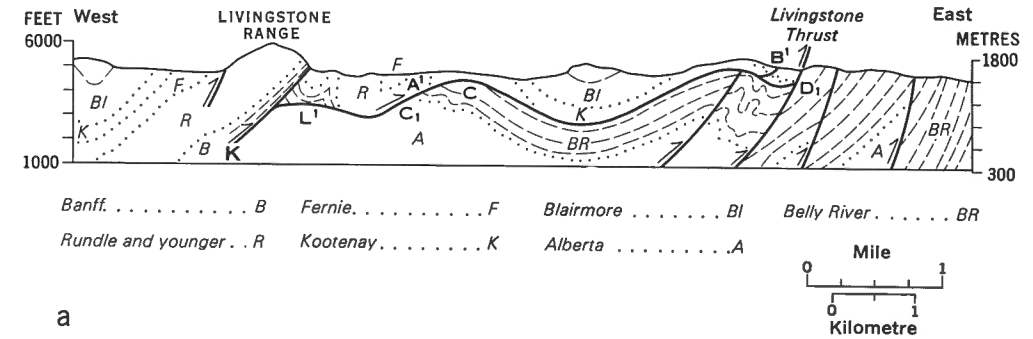
of the Columbian and older orogens were slightly modified and some movement took place on previously formed faults. Scattered, discordant bodies of quartz monzonite, granite, syenite, and quartz diorite were intruded in the western Cordillera. They include small stocks and sills related to volcanic extrusives in Columbian Zwischengebirge and Insular Fold Belt, small batholiths in the southeastern Columbides, and several batholiths of the eastern Coast Plutonic Complex that were accompanied by faulting and minor folding.

During the early Tertiary, deposition continued in most of the negative tectonic elements that were established during the latest Upper Cretaceous: the Rocky Mountain Exogeosyncline, the Pacific and Arctic Continental Shelves, and the intermontane valleys and trenches of the Cordilleran interior. The deposits were mainly nonmarine, synorogenic, clastic sediments. Succeeding volcanic sequences are mainly Eocene. They are mostly subaerial, occurring contemporaneously with tilting and faulting, or related to the intrusion of some of the Laramide granitic stocks. They form several distinct provinces: partly marine basalt and rhyolite that accumulated on the Pacific Continental Shelf; basalt and rhyolite of the St. Elias Mountains; dacite and rhyolite of the central Interior Plateau; andesite, and trachyandesite of the southern Interior Plateau.

During the mid-Tertiary erosion was dominant, both in the Cordilleran Region and on the Interior Plains (Fig. VIII-45). Widespread surfaces of low relief and peneplains were produced, and some of the larger elements of the present drainage system were established, but little of the debris produced is preserved except possibly on the Arctic and Pacific Continental Shelves. Some gravels occur on the southern Interior Plains. In the mountains, faulting contributed to the formation of small intermontane basins where fluviatile and lacustrine sediments accumulated. Marine clastic sediments were deposited on the Pacific Continental Shelf along the west coast of Vancouver Island and along the east coast of northern Queen Charlotte Islands. In the late Tertiary, basalt and olivine basalt erupted from fissures and shield volcanoes. The flows covered large parts of the undulating erosion surfaces of the Interior Plateau and penetrated into the valleys of Columbia Mountains. Some thrusting and folding took place in St. Elias Mountains. Transcurrent movements occurred on many of the major faults, and on some displacement has continued to the present. Plutonism occurred in the core of Cascade Fold Belt. Differential uplift in the Pliocene resulted in the elevation of Coast Mountains, parts of the Interior Plateau, and Rocky Mountains. Many small volcanoes formed, some apparently aligned along faults or zones of weakness. The activity continued locally into the Pleistocene, eruption taking place beneath or on top of the ice sheets.

Laramide Orogeny

During the Laramide Orogeny the layered miogeosynclinal and exogeosynclinal sequences on the eastern and northern elements of the Cordilleran Geosyncline were deformed and many of the structural features of the Eastern Cordilleran Orogen were produced (Fig. VIII-44). The orogeny is herein considered to have taken place during and immediately subsequent to the deposition of the late Upper Cretaceous and early Tertiary non-marine clastic



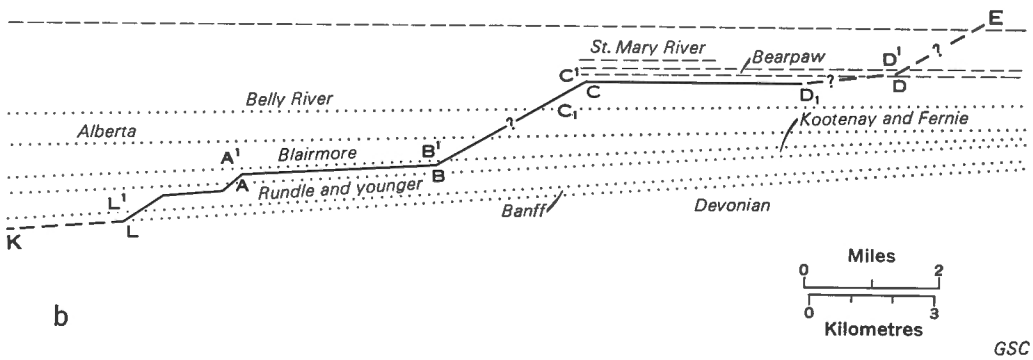
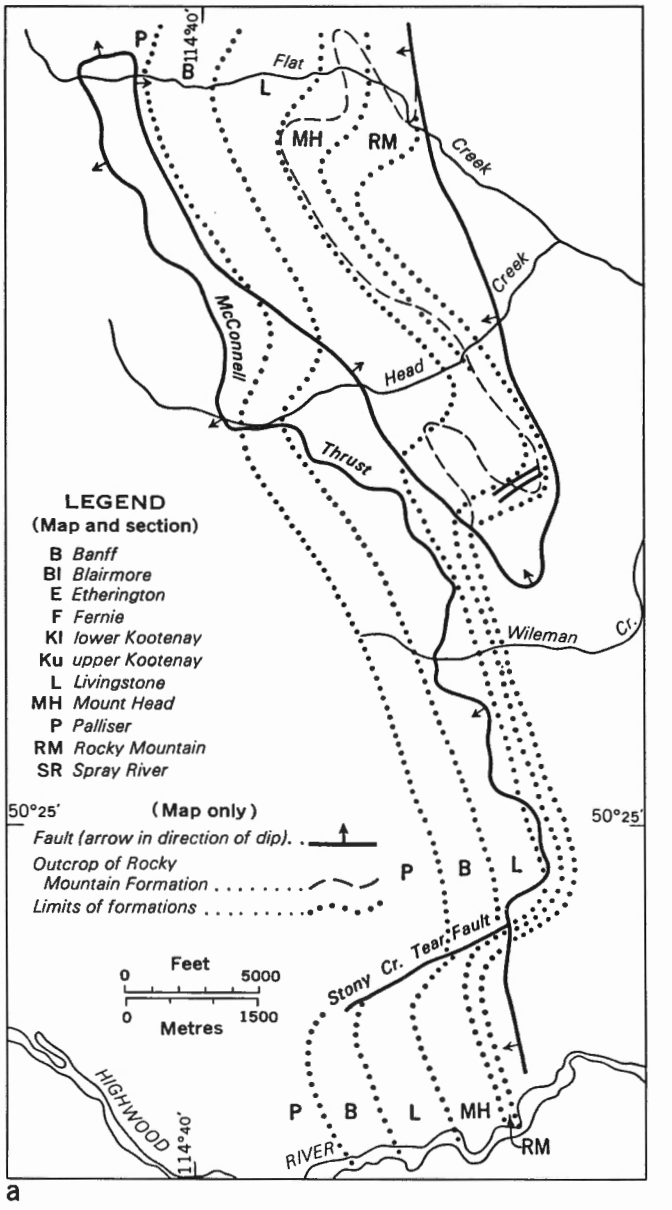


FIGURE VIII-46
Livingstone thrust, southern Foothills of Alberta (Douglas, 1950): (a) structure-section; (b) path (K-E) followed by Livingstone thrust through upper Paleozoic and Mesozoic strata, as inferred from structure-section.



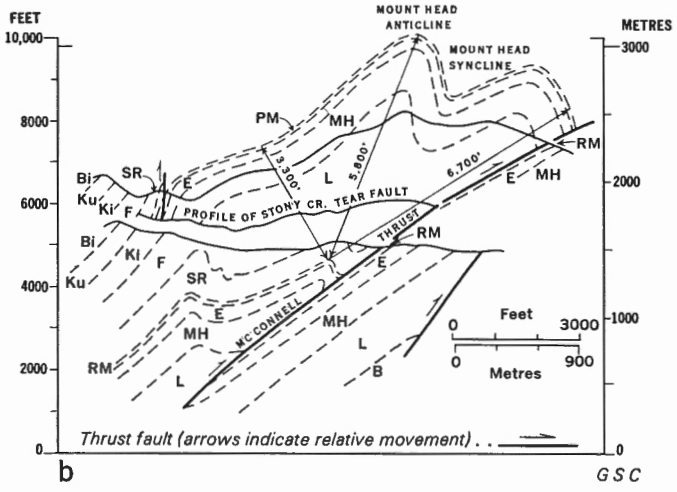


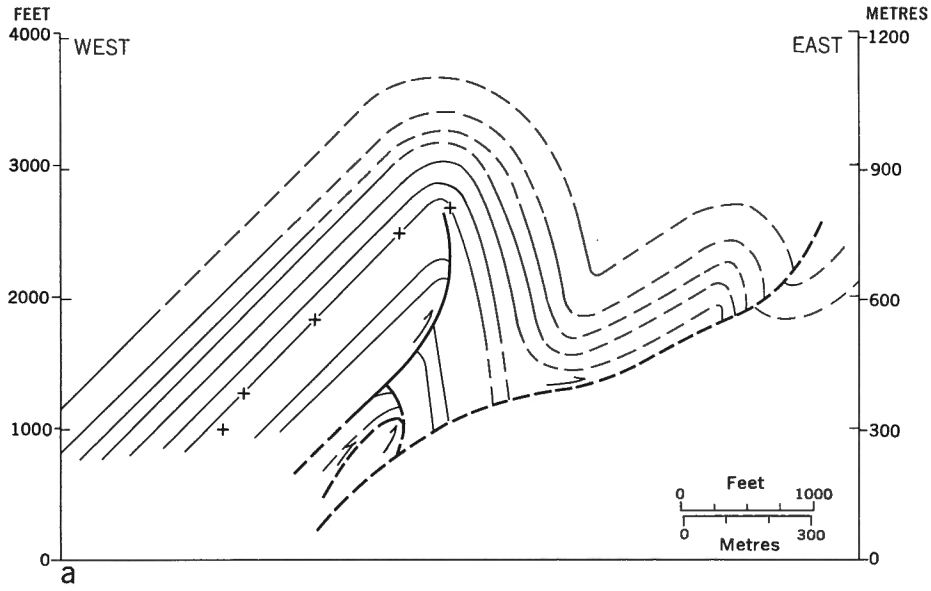
FIGURE VIII-47. Limits of formations above McConnell thrust plane and association with tear faults, southern Rocky Mountains of Alberta (Douglas, 1958).

sequences that are involved in the structures of the southern Rocky Mountains and Foothills and are partly post-orogenic in northwestern Mackenzie and southern Richardson Mountains. The Oligocene and Miocene beds are post-orogenic (Fig. VIII-45). In the western Cordillera, in the mountainous region produced during the Columbian Orogeny, the effects of the Laramide Orogeny appeared to have been limited to transverse and normal faulting, the intrusion of small granite plutons and batholithic intrusions in the Cascade Mountains.

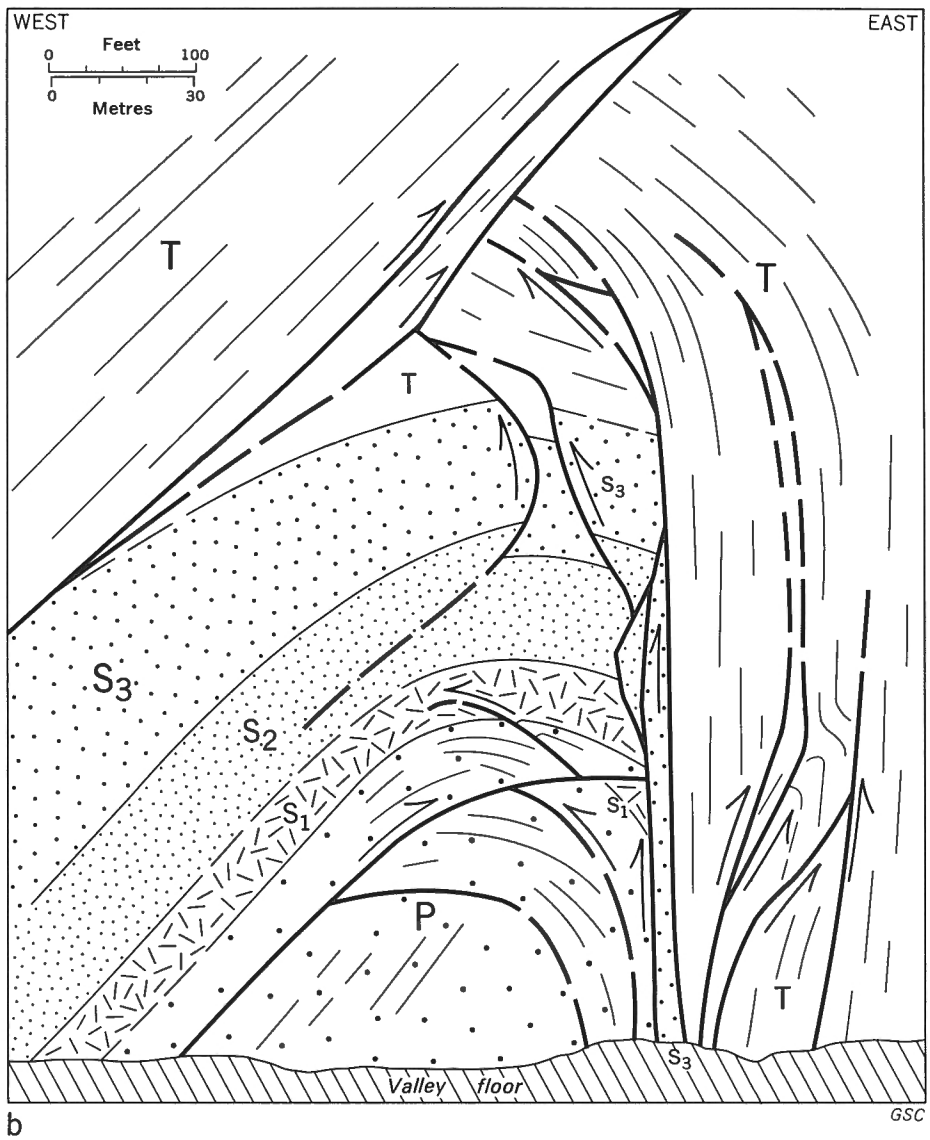
Rocky Mountain Thrust Belt. The Rocky Mountains and Foothills are segmented into three gentle salients separated by shallow re-entrants. The loci of these features are related to the presence of Precambrian basement buttresses that were positive tectonic elements in the early Paleozoic: the Liard, Peace River, and Sweetgrass Arches. The youngest rocks outcrop in the Foothills and the oldest formations in the cores of the salients. Thus, the Foothills are underlain mainly by Mesozoic strata with outliers of Tertiary and inliers of Paleozoic rocks; the more easterly ranges of the Rocky Mountains are formed of Paleozoic and early Mesozoic strata; and the main and westernmost ranges consist of Paleozoic and Proterozoic formations.

The southern Rocky Mountains and Foothills are formed of several large, west-dipping thrust sheets, each of which has been displaced relatively eastwards many miles (Fig. VIII-38a). The strata within the thrust sheets are broken by many closely spaced, subparallel, steeply dipping faults of small displacement, and the folds are complex and disharmonious. In the extreme south total displacement is about 100 miles, representing a shortening of 50 per cent (Bally, *et al.*, 1966), and the thrust belt, as indicated by seismic data, is allocthonous above an undeformed, gently westerly dipping basement of Precambrian crystalline rock (Fig. IX-32). Many of the more westerly structures are considered to have formed in the Columbian Orogeny and to have been modified slightly in the Laramide.

The more westerly thrusts are considered to have originated first, initially as fractures low within the Proterozoic and Paleozoic successions (Fig. VIII-46) and then to have been extended, preferentially following the bedding in zones of easy gliding such as shale or coal, upwards through the Mesozoic and Tertiary strata to the surface and laterally in both directions along the strike (Douglas, 1950). New breaks formed and the extensions of the various thrusts either interlock, merge, or are connected by subsidiary faults. Tear faults are present where the strata initially are transversely cut off by the thrust, the locus being a fault or fracture zone that was produced in the rocks at an earlier time (Fig. VIII-47). The great lateral displacements of the thrust sheets are produced by the riding out of little deformed sheets of rock, the lower zones of gliding being brought into juxtaposition with higher zones. The maximum known displacement is that of the Lewis thrust which has a minimum displacement of more than 45 miles. Initial folds are



Right section through Centre Peak anticline (left) and Green Creek syncline and anticline (right) at Green Creek, Livingstone Range, Alberta. All three concentric folds have developed above a detachment surface that is a west-dipping 'stepped' thrust fault, and above inflection points in the path of this fault through the stratigraphic layering. The axis of curvature or focal line for the upper level of Centre Peak anticline lies along the terminus of an axial thrust fault that rises out of a surface of interstratal slip.



Detail of axial region of Centre Peak anticline. Heavy lines are faults; dashed lines are bedding. P-Pekisko Member; S_1 S_2 S_3 – Shunda Member; T-Turner Valley Member. The thrust fault in upper left is that which terminates as an axis of curvature for Centre Peak anticline, shown in (a). The vertical faults in the centre of the figure are older axial faults that have been rotated in conjunction with the external rotation of bedding produced by continued folding. The normal faults on the right are contraction faults (thrust faults) that have been rotated into an overturned position.

FIGURE VIII-48. Flexural-slip folds in southern Foothills of Alberta (Price, 1964c).

induced concomitantly with displacement of the strata and sequentially, drag-folds are developed with *décollement* at the thrust plane. Splays from the initial thrust plane, sigmoidal in cross-section, merge asymptotically downwards with the thrust and upwards with an earlier formed thrust. The successions of thin thrust slices or imbrications so produced are considered to have formed in sequence from east to west within the overthrust mass, the easternmost slices being detached and over-ridden by the main overthrust mass along the newly formed splay. Where associated with folds some faults occur on the less steeply dipping or back-limb of the anticline rather than the fore-limb, and dip subparallel with the strata. Collapse of the core of the folds and the concave character of the fault plane impart rotation to the strata even to the extent of overturning (Pl. VIII-15). Displacement on a new fault in the footwall strata of a fault results in an induced anticline, a superimposed anticline in both the hanging-wall strata and the fault plane itself, the formation of another fault on the back-limb of this anticline, and cessation of displacement on the folded portion of the first fault. The sequence of faulting and folding of strata and fault planes is considered to be the result of a continuous, self-perpetuating process (Douglas, 1958), or the result of two stages of deformation, firstly the formation of several fault slices and subsequently the folding of them as a group (Scott, 1951).

Folds within the more southerly thrust sheets are essentially concentric and flexural-slip (Price, 1964c). Individual folds are attenuated vertically along their axial surfaces by thrust faults that originate as surfaces of interstratal slip on one limb, and which lie, at their upper terminus, in the constricted core of the overlying concentric fold (Fig. VIII-48).

In the southern Foothills of Alberta the thick late Upper Cretaceous and Paleocene formations are involved in the easternmost structures, mainly as steeply west-dipping homoclinal slices. In the central Foothills, these strata are compressed into long narrow synclines, the upturning of the western flanks of which was consequent upon the development of east-dipping thrust faults. Between these faults and other, west-dipping thrusts are wedge-like zones of greatly thickened, fissile marine shales and sandstones. The thickening is accomplished by the production of many, very closely spaced thrusts which are generally folded concordantly with the strata.

In southern Rocky Mountains are several post-orogenic normal faults that strike subparallel with the thrusts. Most have small displacements, west side down (Pl. VIII-16). They are generally considered to flatten downward and merge with the pre-existing thrusts (Bally, *et al.*, 1966). The Flathead fault in the Flathead River region (Price, 1966) is large and forms part of a graben system that postdates Laramide structures in Lewis thrust sheet, and that formed while the late Eocene to early Oligocene Kishenehn Formation was being deposited. A deep well has been drilled close to the Fording River fault,



PLATE VIII-15. Westerly overturned folds in Miette Range, near Jasper, Front Ranges of southern Rocky Mountains, Alberta. Northern end of vertically dipping McConnell thrust lies beneath dark band on right; folds in Cambrian and Devonian carbonates and shales to left are rotated into overturned position by displacement on concave, upward steepening thrust surface. In background, tree-covered valleys in Mesozoic clastics separate parallel ranges of Paleozoic thrust sheets.

a northern element of the system (Dahlstrom, *et al.*, 1962). The interpretation that the Fording River fault merges with the Lewis thrust is shown on Figure VIII-49; alternatively, the normal fault offsets the thrust at depth.

Normal faults also border Rocky Mountain Trench, particularly the southernmost segment (Leech, 1966). The trench is a major, linear topographical feature (Pl. VIII-17), which is partly underlain by the easterly derived, fault-bounded, late Cretaceous and early Tertiary Sifton Formation in the north, and by Miocene silts in the south. Although the loci of many longitudinal faults are speculatively considered to have been produced by transcurrent faulting, certain structural and stratigraphic elements extend from Purcell Mountains into southern Rocky Mountains without offset other than that by the normal faults. The trench transects structures produced in the Columbian Orogeny and terrain that is allocthonous with respect to the Precambrian crystalline basement. It is thought by Leech to mark an old, deep, fracture zone that reasserted itself through the allocthonous veneer. Price and Mountjoy (1970) suggest that the site of the trench corresponds with the locus of the hinge zone that bordered the Cordilleran Geosyncline prior to the thrusting.

In northern Rocky Mountains, very broad, long, and high-amplitude folds have gently dipping west flanks and vertical or faulted eastern flanks (Fig. VIII-39b). They are cored by a thick west-dipping sequence of Helikian strata, gently folded by the Racklan Orogeny (Pl. VIII-2), and cut by numerous, northerly trending gabbro dykes. According to G. C. Taylor, west-dipping thrusts in the Paleozoic strata pass into *décollement* surfaces parallel

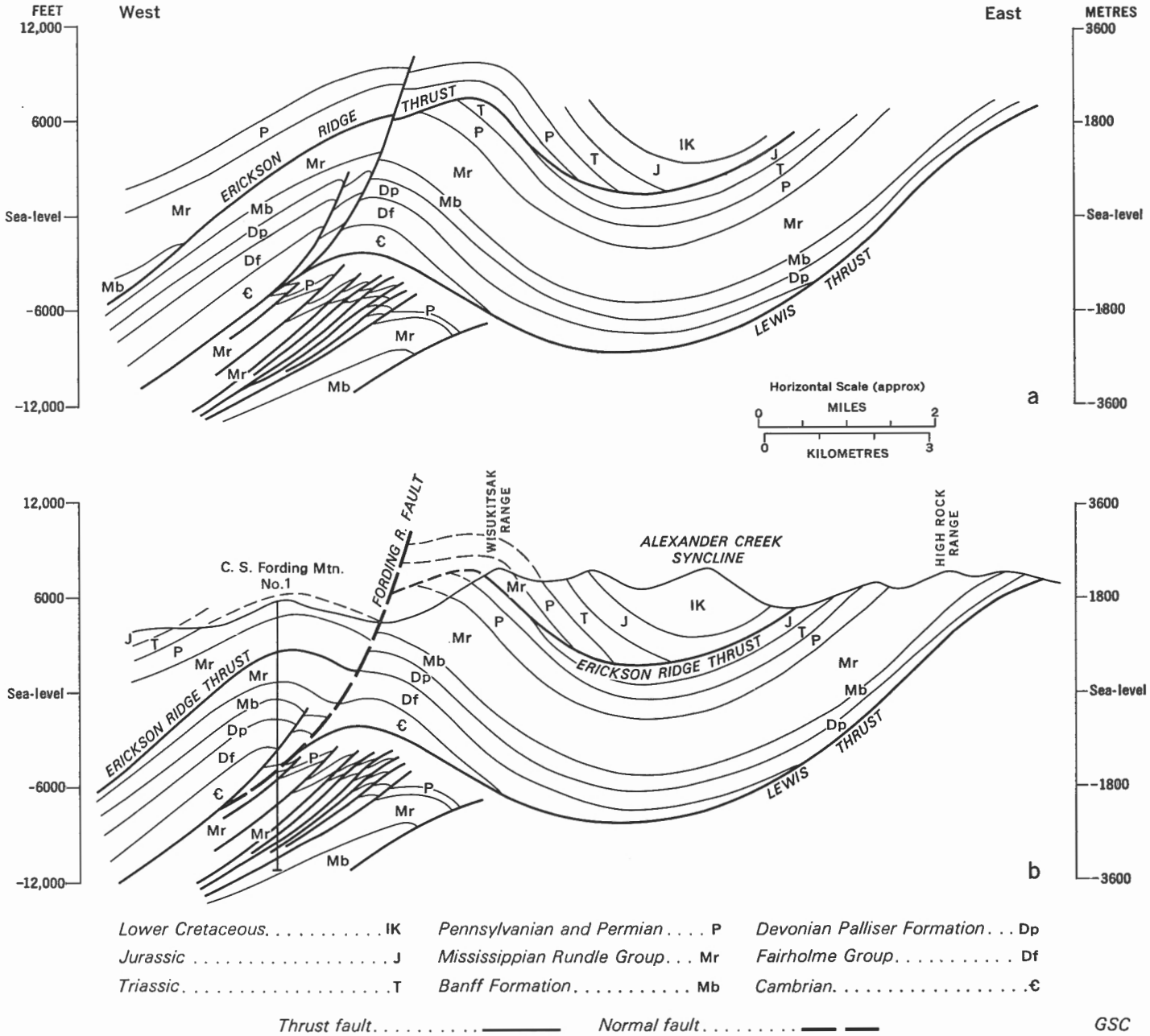


PLATE VIII-16

Normal fault in Paleozoic carbonates, Cirrus Mountain, Main Ranges of southern Rocky Mountains, Alberta. Ml—Mississippian, Livingstone; Mb—Banff; Dp—Devonian Palliser; Df—Fairholme; Omw—Ordovician, Mount Wilson; Ooc—Owen Creek; Osk—Skoki.



PLATE VIII-17. Rocky Mountain Trench. Looking south across a central part of the trench near the junction of Fraser and Torpy Rivers. Cariboo Mountains are on the right; Rocky Mountains on the left with Mount Robson, elevation 12,972 feet, on skyline beneath wing tip.



Lower Cretaceous . . . . . IK	*Pennsylvanian and Permian* . . . . P	*Devonian Palliser Formation* . . . Dp
Jurassic . . . . . J	*Mississippian Rundle Group* . . . Mr	*Fairholme Group* . . . . . Df
Triassic . . . . . T	*Banff Formation* . . . . . Mb	*Cambrian* . . . . . €

Thrust fault . . . . . *Normal fault* . . . . . GSC

FIGURE VIII-49. Cross-section through Lewis thrust sheet and California Standard, Fording River No. 1 well, southern Rocky Mountains; (a) Structure before normal faulting; (b) Structure after normal faulting, the Fording River normal fault decreasing dip at depth and merging with the Lewis thrust (Dahlstrom, Daniel, and Henderson, 1962).

with the bedding of Silurian dolomites or at the unconformity at the base of the Silurian where it lies in contact with the Helikian. The main thrust belt lies west of the broad folds, involving pelitic facies of the lower Paleozoic and the Hadrynian. Structures in the Mesozoic rocks of the Foothills are long, narrow, tightly compressed similar folds in Triassic sandstones, broken by fore-limb thrust faults. The folds are steeply inclined and in the overlying Cretaceous shales become less persistent and more numerous. They are considered to lie discordantly above the structures of the Paleozoic carbonates, the zone of *décollement* being in the Devono-Carboniferous black shales of the Besa River Formation (Fitzgerald and Braun, 1965).

Mackenzie Fold Belt. The Mackenzie Fold Belt is composed of broad, short folds, the crests and troughs nearly flat with sharply upturned and faulted flanks (Fig. VIII-39a). Some folds are long, narrow, and tightly compressed with faults lying on either or both flanks and in the axial region. Some of the broad anticlines are cored with Helikian and Hadrynian strata in northern and western Mackenzie Mountains and Franklin Mountains and with

phyllite in Wernecke Mountains and Taiga Ranges. The lower Paleozoic contains much carbonate, the contact of the uppermost, generally Middle Devonian massive beds with the overlying, mainly Upper Devonian shales forms extensive stripped bedding surfaces that sharply delimit the flanks of the anticlinal mountain ranges and the intervening synclinal valleys (Pl. VIII-4). In the extreme south, in Liard Plateau, a low east-facing homocline dominates the structure. It is bounded on the west by east-dipping thrust faults and extends for a width of 50 miles interrupted only by a few gentle folds, asymmetrical westward. To the north anticlinal mountain ranges appear right hand *en echelon*. In the southern part of Mackenzie Fold Belt the folds are markedly segmented. The individual segments are oriented along west, northwest, and northeast lines and are commonly broken by faults with a small amount of thrust or normal displacement that also have the same trend. A fracture system appears to have been present in the rocks as faults or zones of small fractures prior to the Laramide deformation. Old faults with normal displacement may have been rotated along strike into attitudes where thrust displacement occurs. In western Mackenzie Mountains some tightly folded anticlines are broken by what appears to be two episodes of folding, manifested by faulting of older faults and by overturning of the folds to the east or west in the opposite direction to an earlier overturning. These structures may represent some of the easternmost elements of Columbian Orogen. Most thrust faults are associated with folds and with their middle or fore-limb. The amount of deplacement is small; shortening across the Mackenzie and Franklin Mountains is only about 10 per cent. The thrust faults dip either eastward or westward and commonly end at a small tear fault which is oriented in the direction of the old fracture system, or which may be linked by a short segment of it. On these segments little horizontal displacement has occurred; rather, the beds have undergone only rotational translation. Thrust faults with homoclinal hanging-wall and footwall strata are rare. Generally, subsidiary structures indicate some dextral motion in addition to the dip-slip where the trend of the thrust is northeasterly, and sinstral where northwesterly.

In the Mackenzie River valley near Fort Norman, non-marine beds, poorly dated as Eocene, are involved in the folding and thrusting, particularly near Mackay Range. As the beds lie with strong regional unconformity on Lower Cretaceous to Middle Devonian strata, some tilting, warping, and possibly also folding and faulting may have taken place during the early phase of the Laramide Orogeny or earlier. Some small complex structures in Mackenzie Plain and Franklin Mountains appear to have been produced with *décollement* at the level of the Cambrian salt and anhydrite. The larger structures, particularly the major anticlines with cores of Helikian and Hadrynian strata probably formed with *décollement* close to the crystalline basement.

Northern Yukon Fold Complex. In the vicinity of Bonnet Plume Basin where the Richardson and Mackenzie Mountains merge, very gently folded and tilted non-marine beds of the late Upper Cretaceous or early Tertiary Bonnet Plume Formation lie with angular unconformity on deformed Paleozoic and Proterozoic rocks (Fig. VIII-50), the bulk of the orogenic deformation having taken place prior to their deposition. The structural grain around the periphery of the basin is considered by R. A. Price in terms of two sets of structural elements whose antecedents have controlled patterns of deformation and sedimentation since the Precambrian. The structure of southern Richardson Mountains is dominated by northerly and northwesterly trending faults, most or all of which have a large right-hand separation. These faults bound a complex horst and broad, north-plunging anticlinorium that forms the core of the mountains. In northwestern Mackenzie Mountains and southern Porcupine Plateau the dominant system of east-west trending folds and faults is associated with a less conspicuous system of northerly and northwesterly trending, right-hand faults and folds that join those of Richardson Mountains. The anisotropy inherited from the structural elements that were produced in the Hadrynian and Paleozoic exerted a strong influence on the character of the structures produced during the early phase of Laramide Orogeny. The older faults were reactivated and assumed new kinematic roles related to their orientation relative to the direction of maximum principal stress. For example, the western sides of the faults bounding eastern Richardson Mountains were downthrown during the Paleozoic, particularly during deposition of the Lower Cambrian and the Middle Devonian. The faults now have the northeast side downthrown and have been subjected to right-hand transcurrent displacement. In northern Mackenzie Mountains they merge along strike with easterly trending thrust faults downthrown to the north. The kinematic roles of both easterly and northerly trending faults can be integrated in terms of a single deformation arising from a north-south shortening in conjunction with relative tectonic transport northward. In northernmost Richardson Mountains near Aklavik some displacement on the wrench faults on the east side can be seen to be later than mid-Upper Cretaceous. Albian shales are involved in the folds fronting northwestern Mackenzie Mountains. Inasmuch as these structures are considered to be kinematically related and are in part older than the late Upper Cretaceous Bonnet Plume Formation, a large part of the structures of the northern part of the Eastern Cordilleran Orogen are considered to be early Laramide.

Long broad open anticlines and synclines, some displaying *en echelon* linkage are present on Eagle Plain and Nahoni Range. There, northerly trending structures appear to be truncated against a major thrust on the southeast side of Keele Range. Culminations and depressions mark their intersection with the west-trending structures in the Taiga Ranges.

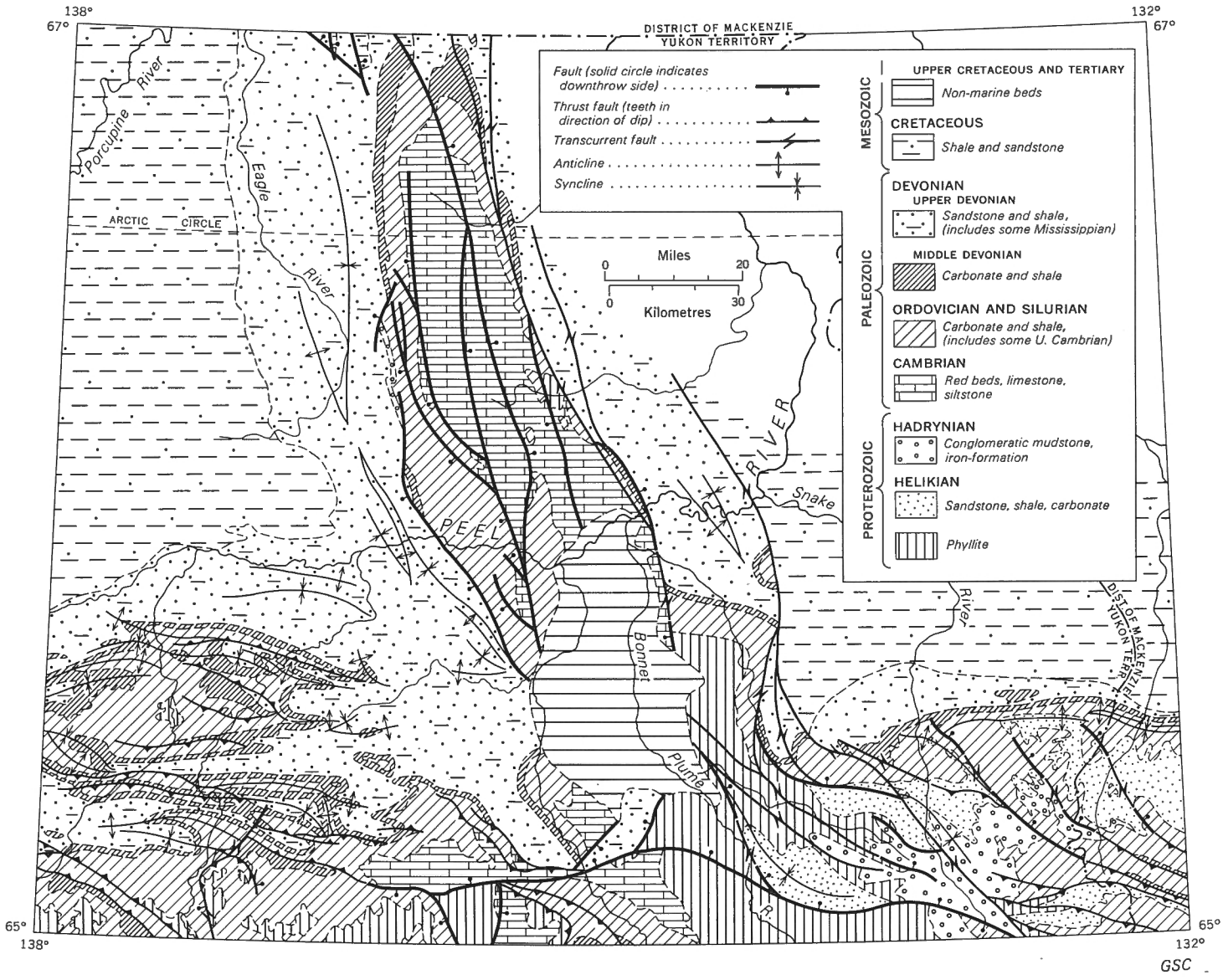


FIGURE VIII-50. Geology, provisional, in the vicinity of Bonnet Plume Basin, northern Yukon Territory, showing early Laramide structures of southern Richardson and northwestern Mackenzie Mountains (by R. A. Price and D. K. Norris).

Yukon Plateau. In the northern Coast and Omineca Geanticlines and in the intervening area, large, sharply cross-cutting bodies of quartz monzonite and alaskite or miarolitic granite were emplaced near the end of the Cretaceous. The granites are characterized by highly variable texture from fine to very coarse grained, and in places porphyritic, abundant smoky quartz, low mafic-mineral content, and dull creamy colour of the feldspars. Small pegmatitic streaks and clots are common and drusy cavities are abundant. Traces of fluorite and apatite are widespread whereas allanite and topaz are rare. Tourmaline and axinite are locally abundant in one batholith. The quartz monzonites are similar to the miarolitic granites but contain less potash feldspar and more mafic minerals. These granitic rocks were apparently emplaced at relatively shallow depths during a time of crustal tension and no significant regional deformation appears related to them.

St. Elias Fold Belt. The time of pre-Tertiary deformation in the Canadian part of the St. Elias Mountains is not known. However, Paleocene and later volcanic and sedimentary rocks are displaced by thrust and counterthrusts (Muller, 1967) and possible tillites in easternmost Alaska equivalent to part of the St. Clare Group are tilted and gently folded. The age of the 'tillites' is not known, but they may be equivalent to marine tillites of Miocene and Pliocene(?) age on the Gulf of Alaska that were involved in latest Pliocene or earliest Pleistocene deformation. Movement along faults on the southwest side of the St. Elias Mountains continues to the present day, if one judges by the strong earthquakes of 1899 and the last decade. The central crystalline core of St. Elias Mountains with fault blocks of Lower Cretaceous sediments is



PLATE VIII-18. Late Tertiary thrusts in Kluane Ranges, St. Elias Mountains, Yukon Territory. Shakwak Trench, floored with Pleistocene deposits, separates the younger rocks of St. Elias Fold Belt from Yukon Crystalline Platform which is underlain by early Paleozoic and late Precambrian metamorphic rocks and granitic plutons of Mesozoic and Tertiary ages.

bounded on both sides by inward-dipping faults that separate it from adjacent regions that are characterized by folds and thrust faults directed away from the axial region of the fold belt. On the northeastern side, the Duke River thrust zone is marked by numerous lenses of gypsum and converges southeastward with the Shakwak fault that bounds the fold belt on the northeast. The intervening zone is intensely deformed and has at least two counter-thrusts directed to the southwest (Pl. VIII-18).

Insular Fold Belt. Structures in Upper Cretaceous rocks of Georgia Basin, produced in the early Tertiary, appear to reflect movements on faults in the basement and are related to two prominent, steep northwest-trending faults on which beds are downdropped on the southwest. The Upper Cretaceous rocks are intruded by sills and stocks of quartz diorite with K–Ar dates of 35 to 59 m.y. At the south end of Vancouver Island, folds in the Tertiary formations trend easterly and are associated with the north-

dipping, east-trending thrust on which displacement took place in the Oligocene. On Queen Charlotte Islands gentle warps in the Tertiary volcanics may merely represent sagging above an evacuated magma chamber. The faults that affect Tertiary rocks are generally steep and trend northwesterly, parallel with the prevailing fault and fold trends in older rocks. On Queen Charlotte Islands they have been active since at least the Lower Cretaceous and combine right-hand wrench movement with an east-side down displacement. On Vancouver Island the northwesterly faults were established in Middle Jurassic time but the nature of their subsequent history is unknown except that movement on those along the west coast downdropped Tertiary sediments in post-Miocene time.

The northwest-trending faults also include one on the continental margin off the west coast of the Queen Charlotte Islands (Fig. VIII-51). This fault is considered by Sutherland Brown (1966) to dip steeply under the continent and by Wilson (1965) to be one of a series of transform faults within the San Andreas–Denali linkage. It separates northwesterly trending continental structures to the northeast from a region on the ocean floor to the southwest that is characterized by north to northeasterly trending magnetic anomalies (Fig. VIII-52). Vine (1966) considered that the oceanic crust has spread laterally away from the axis of the East Pacific Rise in the region of the Juan de Fuca ridge. He estimated that the rate of spreading may have decreased within Pliocene time from a rate of 4 or 5 to 2.9 cm/yr for the last 5 m.y.

Interior Plateau. Early Tertiary structures are extensional features such as normal faults and dyke-filled fractures superimposed on Columbian structures. Movement along the faults took place partly contemporaneously with the deposition of the early Tertiary sediments and volcanic rocks, and partly later. The normal faults have a zig-zag pattern in southwestern Monashee Mountains. Tertiary strata in the horsts, grabens, and half-grabens are locally folded and commonly tilted eastward. The fault system controlled the direction of basalt and lamprophyre dyke swarms and influenced the distribution of the Coryell Intrusions. Northerly trending topographical elements in the poorly exposed Tertiary volcanics of the central Interior Plateau may also reflect similar northerly trending block faults. More variable trends and attitudes of tilting and folding occur in the northern part near Stikine Arch. Some reverse movement took place on Pinchi fault subsequent to the deposition of the Upper Cretaceous to Paleocene Sustut Formation.

Cascade and Coast Mountains. Recurrent movements on the Fraser River fault zone that transects the core of the southern Cascade Mountains resulted in the localization of Eocene continental clastics and their subsequent deformation prior to the emplacement of Chilliwack batholith in the mid-Tertiary. Similar events affected the southern part of Coast Mountains where late Upper Cretaceous beds are tilted, faulted, and intruded by granitic plutons.

Paleocene and Eocene (Fig. VIII-44)

Pacific Continental Shelf. In Bellingham Basin beneath the lower Fraser River Delta, the upper beds of non-marine sandstones and conglomerate comprising the Burrard Formation are early Tertiary. The conformably overlying Kitsi-

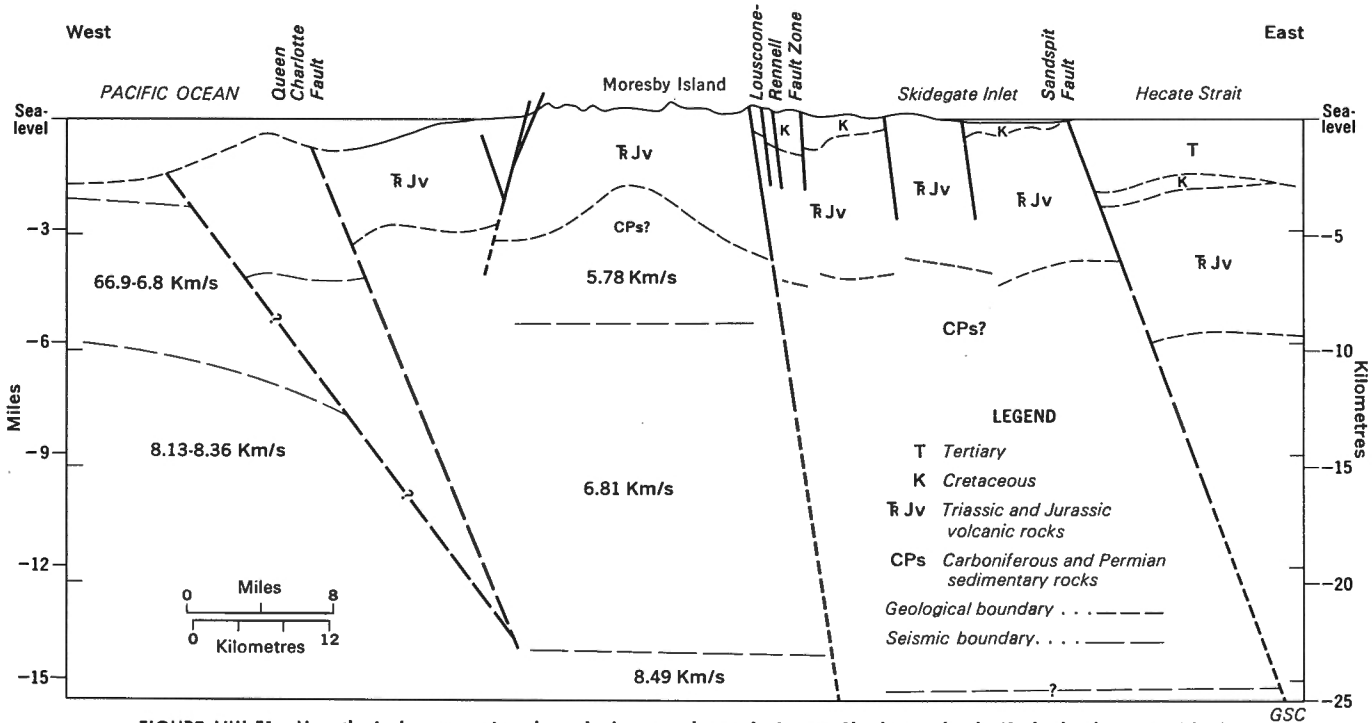


FIGURE VIII-51. Hypothetical cross-section through the crust beneath Queen Charlotte Islands (Sutherland Brown, 1967).

lano Formation and the equivalent Huntingdon Formation contain a Middle Eocene flora. In Washington, on the southern margin of the basin, the Huntingdon Formation lies unconformably on the Chuckanut Formation, the lower part of which is equivalent to the upper Nanaimo Group. At the southern end of Vancouver Island the Eocene Metchosin Formation comprises 7,500 feet of submarine pillow basalt, minor pyroclastics, and sediments. It lies unconformably on the Lower Cretaceous on Olympic Peninsula of Washington. On southern Queen Charlotte Islands the Massett Formation has several distinctive facies (Sutherland Brown, 1966). The Dana Inlet facies comprises subaqueous pyroclastic breccias with mixed rhyolitic and basaltic clasts. The subaerial equivalent, the Kootenay Inlet facies, consists of rhyolitic ash flows, tuffs, and breccias, and dacitic and basaltic flows. The Tartu Inlet facies of subaerial volcanics in northern Queen Charlotte Islands is composed of three members totalling 18,000 feet in thickness. The lowest member is a mixture of basalt breccias and flows and rhyolite ash flows, tuff, and flows. The middle member is mainly rhyolitic flows and ash. The upper member consists of slightly sodic columnar basalt flows. Mica from a sill related to the lower member yielded a K–Ar date of 62 m.y.

Cascade Mountains. Conglomerates of local derivation lie unconformably on the Custer gneiss and granitic and metamorphic rocks along Fraser River south of Lillooet (McTaggart and Thompson, 1967). They are also faulted against similar rocks to the east thus indicating faulting along Fraser River during and after the Eocene. Along the same trend south of Hope remnants of Eocene conglomerate are engulfed by the Miocene Chilliwack batholith that yields K–Ar dates of 18 m.y.

Interior Plateaux. Subaerial volcanic rocks with subordinate clastic rocks are 5,000 feet thick in the interior of British Columbia. They are preserved in isolated structural basins and fault troughs, or occur as upland remnants lying with angular unconformity on older rocks.

In the southernmost Interior Plateau the Kamloops and Princeton Groups, and the Springbrook, Marron,

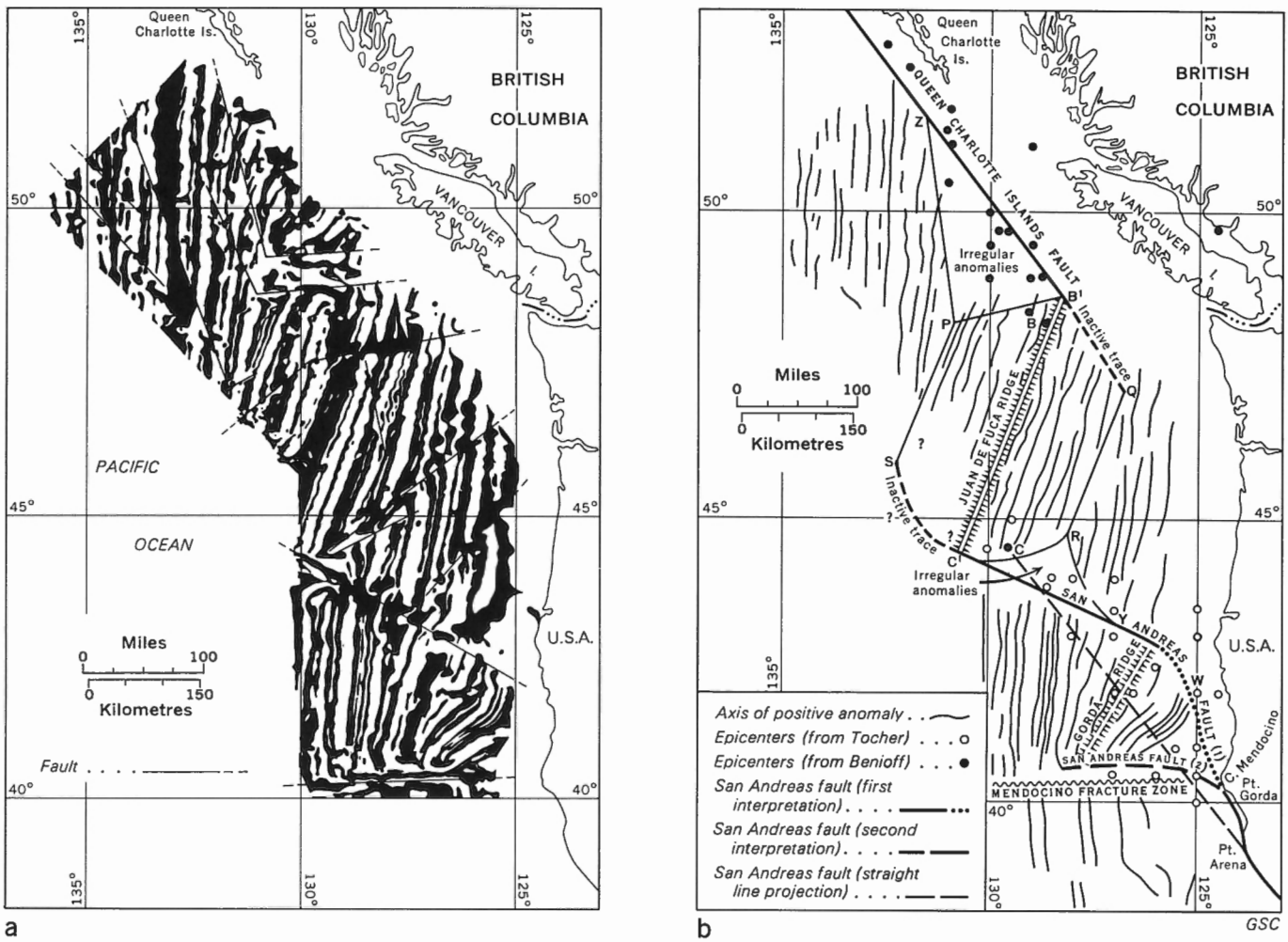


FIGURE VIII-52. Magnetic anomalies southwest of Vancouver Island: a) Total magnetic field: areas of positive anomaly shown in black; straight lines indicate faults (Vine, 1966); b) Interpretation of transform faults (Wilson, 1965).

White Lake, and Kettle River Formations are considered Middle Eocene on the basis of flora from the sediments of the Kamloops and Princeton Groups and the Kettle River Formation, and mammal teeth near Princeton. The volcanic rocks have yielded K–Ar dates ranging from 45 m.y. to 49 m.y. (Rouse and Mathews, 1961). They are mainly andesites and trachy-andesites with subordinate basalt and rhyolite that occur as lenticular flows and pyroclastic sheets. The felsic types commonly are ash flows. Tilting and faulting took place contemporaneously with the deposition of sediments and eruption of the volcanic rocks. The Kettle River Formation increases in thickness towards some bordering faults. Feeder dykes for the middle Marron lavas have the same trends as normal faults, and also appear to be related to the Coryell Intrusions.

The Ootsa Lake Group of the central Interior Plateau is composed principally of spherulitic, perlitic rhyolite, and subordinately of andesite, basalt, and clastic sediments. Flows are about as common as pyroclastic rocks. Some volcanic rocks, particularly felsic types, originally mapped with the younger Endako Group (Armstrong, 1949) have yielded K–Ar dates of 53 m.y. in common with early Tertiary volcanics of the Ootsa Lake Group and those in southern British Columbia (Mathews, 1964).

The Sloko Group (Aitken, 1960) of northern British Columbia is mainly pyroclastic dacite and rhyolite with subordinate andesite and basalt, and contains poorly preserved plants of Cretaceous or Tertiary age. The pyroclastic rocks range from coarse explosion breccias, some of which consist entirely of granitic fragments, to delicate vitric tuffs and ignimbrites. Flows are rare. The associated sediments are derived from nearby volcanic material and contain coal, plant fragments, and logs. The Sloko strata are flat or gently tilted and are commonly bounded by normal faults or ring fractures. Some dykes follow the faults or form parallel swarms, suggesting that volcanism was contemporaneous with faulting and locally accompanied by cauldron subsidence. Some of the trachytic volcanics are apparently genetically related to quartz monzonite plugs, one of which has a K–Ar date of 69 m.y. Locally, the pyroclastics grade downward to felsite and thence to quartz monzonite through a difference in elevation of less than 5,000 feet.

In southwestern Yukon Territory the Carmacks and Skukum Groups contain sediments of Paleocene and Eocene age, overlain by volcanic rocks, mainly basalt, but including trachyte, rhyolite, andesite, and pyroclastic rocks. The rocks are tilted and truncated by an erosion surface probably of Miocene(?) age.

St. Elias Mountains. Southwest of Shakwak Trench in Kluane area Paleocene sandstone and conglomerate with minor shale and coal are as much as 2,500 feet thick. They have been involved in middle or late Tertiary thrusting. In Dezadeash area flat-lying and gently dipping beds of conglomerate, sandstone, and shale, 3,500 feet thick, lie unconformably on porphyritic granite or on Mesozoic volcanics and sediments. The strata are cut by a stock and numerous sills of soda syenite. In the Squaw Creek–Rainy Hollow area (Watson, 1948) gently to moderately dipping conglomerates, sandstones, and shale overlie porphyritic rhyolite flows in a down-faulted block. The sedimentary and volcanic units, each about 1,500 feet thick, are in part intercalated and both are considered Paleocene.

Coal seams are present in well-stratified Paleocene or upper Eocene conglomerates, sandstones, and siltstones along northwestern Tintina Trench and also in the Carmacks area. Similar rocks, lacking coal, are present along Pelly River in Tintina Trench and along MacMillan River to the northeast. Lignitic coal seams occur with typical Tertiary clastic rocks on Liard and Coal Rivers in southeastern Yukon Territory and along Rapid River near McDame.

Arctic Coastal Plain. West of Mackenzie Delta the Moose Channel Formation consists of 1,200 feet of non-marine sandstone, loosely consolidated, fine- to coarse-grained, feldspathic, greenish grey to brown, and ripple-marked, interbedded with silty grey shale (Mountjoy, 1967). A seam of sub-bituminous coal 7 feet thick occurs on Moose Channel. Microflora indicate a general Upper Cretaceous to early Tertiary age. East of the delta, the Paleocene to Eocene Reindeer Formation, more than 700 feet thick, contains non-marine silty shales, siltstone, common sandstone, quartzite, and chert conglomerate and coal.

Mackenzie Plain. In the vicinity of Fort Norman Paleocene or Eocene shales, forming a small basin, rest with regional unconformity on Lower Cretaceous to Middle Devonian strata. The lower part, 4,000 feet thick, comprises soft sandstones, shales, and lignite. The upper part, capping hills south of Little Bear Lake, is 2,000 feet of massive conglomerate.

Rocky Mountain Exogeosyncline. The Paleocene of the Interior Plains and Rocky Mountain Foothills ranges from more than 5,000 feet in the west to only a few hundred feet in southwestern Manitoba. Massive buff-weathering sandstones, included in the Paskapoo Formation of central Alberta and in the Ravenscrag in Saskatchewan, overlie the Cretaceous coal- and dinosaur-bearing beds. An erosional unconformity apparently separates the Edmonton and Paskapoo Formations although deposition may have been continuous in southern Saskatchewan and Alberta. In the southern Foothills, the boundary between the Cretaceous and the Paleocene lies within the apparently conformable sequence of the Willow Creek Formation.

The Willow Creek Formation consists of soft, medium-grained grey sandstone and variegated shales overlain by buff-weathering beds. It contains two molluscan faunas (Tozer, 1956), the older being correlated with that in the upper Edmonton Formation. The younger fauna is correlated with that in the Paskapoo and Ravenscrag Formations. The Willow Creek is more than 4,000 feet

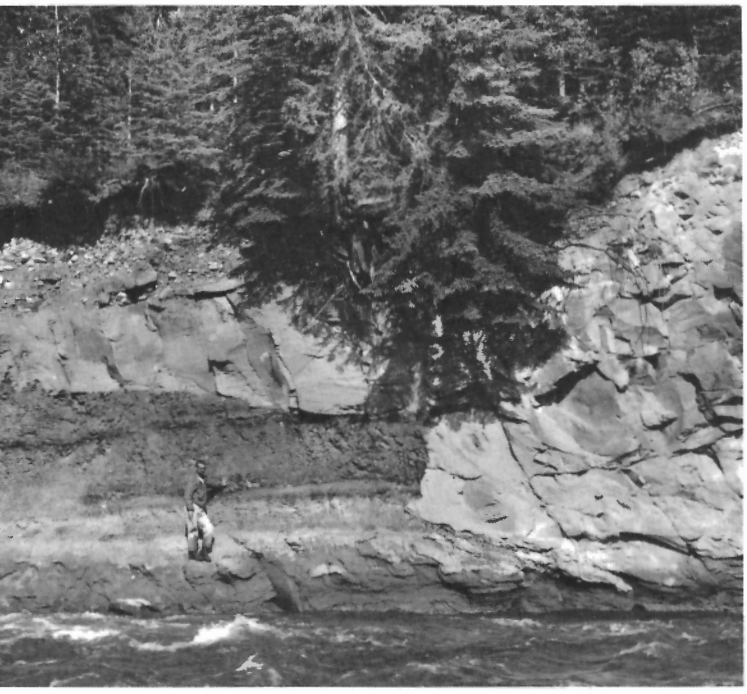


PLATE VIII-19. Sandstone channel-filling in mudstones of Tertiary Paskapoo Formation, Burnt Timber Creek, Foothills of Alberta.

thick near Castle River and thins northward, the lower part merging with the upper Edmonton, and the upper part bevelled beneath the unconformably overlying Porcupine Hills Formation (Douglas, 1950). The Porcupine Hills include about 4,000 feet of coarse-grained, massive, crossbedded, grey and green sandstone and grey and brown shales.

The Paskapoo Formation is 3,000 feet thick in the southern Foothills and more than 6,500 feet in the central Foothills (Pl. VIII-19). It contains massive, crossbedded, coarse-grained, brown sandstone, grey and green shales, and some limestone. The thin, lenticular Entrance conglomerate commonly occurs at the base between North Saskatchewan and Athabasca Rivers. Thick coal seams of economic importance occur near Edmonton. The Paleocene age has been established by a well-preserved molluscan fauna, mammalian remains, and plants.

In the Cypress Hills along the Alberta–Saskatchewan border, Paleocene beds with a mammalian fauna are included in the Ravenscrag Formation (Furnival, 1946). The base is drawn below the lowest commercial coal seam and appears to be conformable with the underlying, dinosaur-bearing Frenchman Formation. The Ravenscrag is a cyclic succession of silts, clays, and lignite that ranges in thickness from about 500 feet to 250 feet, thinning eastward across southern Saskatchewan.

The Turtle Mountain Formation in southern Manitoba is considered to be early Tertiary, and probably Paleocene in age. The formation, estimated to be about 100 feet thick, contains fine sand, silt, clay, and thin coal seams. Paleocene foraminifera from the Cannonball Formation, which occurs on the west flank of Turtle Mountain in North Dakota, indicate that marine strata may lie between the Turtle Mountain Formation and the Cretaceous Boissevain Formation. This short-lived Cannonball Sea had connections with the Gulf of Mexico and perhaps with the Arctic Ocean. It may be possibly related to the post-Ravenscrag subsidence of Williston Basin.

Oligocene and Younger (Fig. VIII-45)

Pacific Continental Shelf. Clastic marine sediments of the middle and Upper Oligocene Carmanah Formation and of the Lower Miocene Sooke Formation occur along the west coast of Vancouver Island. They unconformably overlie older rocks and are unconformably overlain by the Pleistocene. The beds strike parallel with the coast and dip west, oceanward. The Skonun Formation of Miocene and Pliocene age on northern Queen Charlotte Islands comprises intertonguing marine and non-marine sands and shales, 6,000 feet thick in the east. It overlaps Lower Tertiary volcanics westward (Sutherland Brown, 1966).

Interior Plateaux. Miocene silts, sands, and gravels occur along the major valleys such as that of the Fraser River and its tributaries near Prince George (H. W. Tipper), Thompson River (Campbell and Tipper, 1966), and southernmost Rocky Mountain Trench (Leech, 1966). These occurrences indicate that the major drainage routes were established by the Miocene and preceded by a long interval of erosion.

The flat-lying Plateau Basalts of central British Columbia and southern Yukon Territory have yielded K–Ar dates of 10, 12, and 13 m.y. They overlie and interfinger with clastic sediments containing late Miocene flora and locally include diatomite deposits. The Plateau basalts and olivine basalts and their equivalents, the andesites and basalts of the Endako Group, were erupted from fissures and shield volcanoes onto a gently undulating erosion surface that had a relief of 1,500 to 2,000 feet. In the east, the lavas apparently flowed as wide sheets into the valleys bordering Columbia Mountains (Campbell and Tipper, 1966). In the west, however, in the Coast Mountains, the basalts, in response to differential uplift in the Pliocene now lie at elevations of as much as 8,000 feet. Volcanic rocks in northwestern British Columbia equivalent to the Plateau Basalts are the earliest flows of the Mount Edziza volcanic cone (Fig. VIII-53), the basalt flows of the Level Mountain Formation, and the interfingering rhyolitic and dacitic pyroclastics of the Heart Peaks Formation that have yielded a K–Ar date of 15 m.y.

Near Alsek River in St. Elias Mountains, volcanic rocks possibly of Miocene age are varicoloured agglomerate, breccia, tuff, rhyolite, dacite, andesite, and basalt about 3,500 feet thick. These rocks are believed to overlie Paleocene and older strata unconformably and to have been extruded onto a surface having more than 2,000 feet relief. Farther northwest in Kluane Ranges, about 1,600 feet of massive and columnar basalt and andesite are

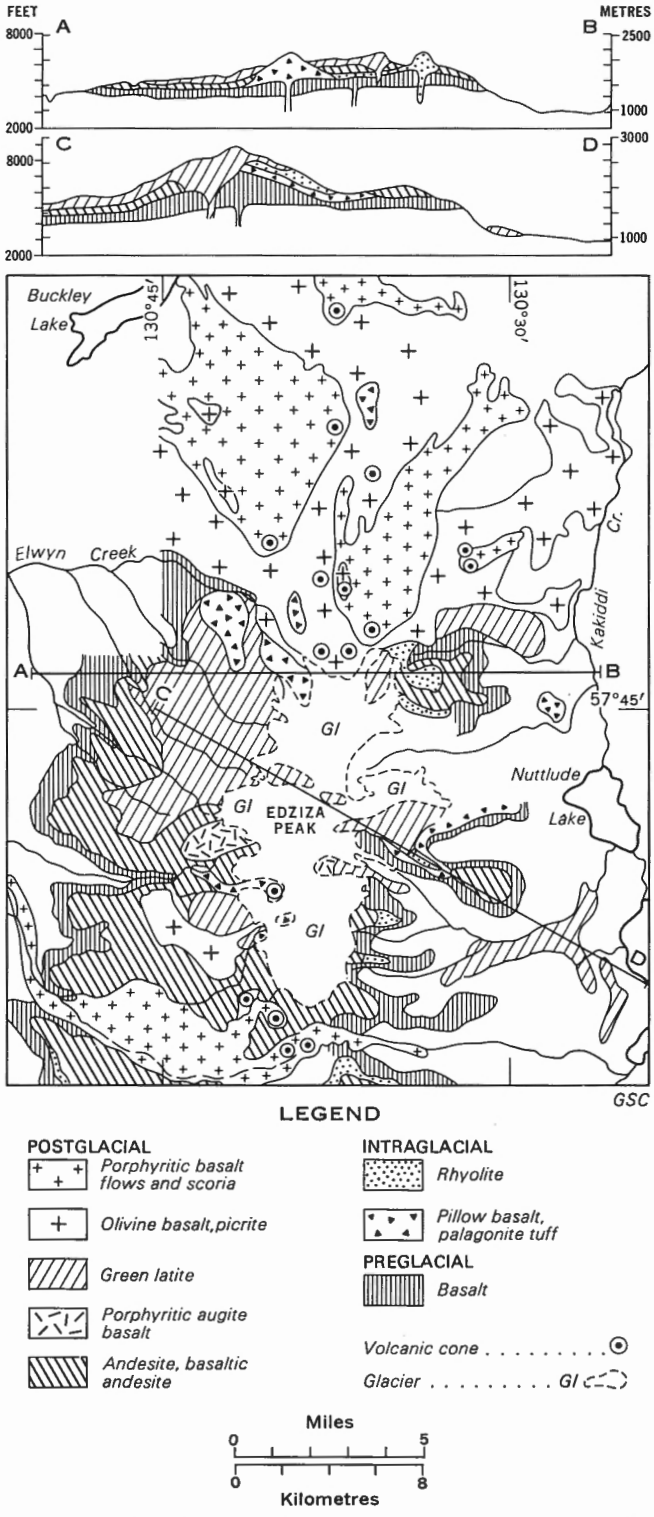


FIGURE VIII-53. Mount Edziza volcanic complex, northwestern British Columbia (Souther, 1966).

intercalated with sand, shale, and coal at the base, and contain intercalated layers of tuff and ash near the top. They are overlain by 350 feet of coarse volcanic conglomerate.



PLATE VIII-20. Cinder cone, 500 feet high, on post-glacial basalt of north flank of Mount Edziza volcanic complex, British Columbia. Partly eroded cones in middle distance.

Lavas of the Selkirk Group and related rocks in southwestern Yukon Territory and northern British Columbia range from Pliocene to Recent. They were extruded onto an erosion surface similar to the present one, and many flows were confined to the present valleys. The Tuya Formation in northern British Columbia includes a great deal of volcanic ash that is intercalated with flows near the centres of eruption. In several places volcanic cones consisting predominantly of ash are capped by flat-lying flows thought to have been emplaced at a time when the region was covered by ice. The "Valley" olivine basalts of the southern Interior Plateau occur as benches and in valleys carved into the plateau surface. Many of the volcanic cones or centres are shown on Figure VIII-45.

Rocky Mountains. In upper Flathead River valley, the Kishenehn Formation of latest Eocene or earliest Oligocene age was deposited in a structural basin along the down-

thrown side of the Flathead normal fault. Two main facies are recognized by Price (1966): conglomerates and breccias adjacent to the fault derived locally from the Paleozoic and Precambrian rocks of the upthrown side, and a basinal facies of fine clastic material, marl, and lignite. Sedimentation took place simultaneously with displacement on the fault as the sequence of rock types that occur as clasts is the inverse of their normal stratigraphic order in the hanging-wall succession. About 6,600 feet of sediments are present. The Kishenehn lies with angular unconformity on strata deformed by the Laramide Orogeny, and establishes an upper limit for dating the underlying structures of the southern Rocky Mountains.

Interior Plains. Gravels derived from Precambrian and Cambrian rocks of the western and main ranges of southern Rocky Mountains are included in the late Eocene Swift Current Creek beds and the early Oligocene Cypress Hills Formation of southwestern Saskatchewan and southeastern Alberta. The latter, capping the Cypress Hills, consists of varying amounts of sands and gravels and has a maximum thickness of 300 feet. The pebbles are mainly quartzite and chert with some trachyte and other porphyritic rocks. The formation has yielded a large vertebrate fauna, especially mammals. Other remnants of fluvial deposits representing erosion of the Rocky Mountains are the Miocene Wood Mountain gravels of southern Saskatchewan and the Miocene or Pliocene conglomerates and gravels capping Swan Hills, Hand Hills, and Saddle Hills in central Alberta.

Quaternary

Almost all the Cordilleran Region and Interior Plains were glaciated during the Pleistocene (*see* Chapter XII and Map 1253A). In the unglaciated part of Yukon Plateau, the gold-bearing stream deposits of the Klondike including the "White Channel" gravels and related deposits were preserved. They may be latest Tertiary. Elsewhere, a great variety of glacial and fluvioglacial deposits are present and in some of the pre-glacial valleys attain great thicknesses and may include earliest Pleistocene and latest Tertiary sediments. Large snowfields and numerous active glaciers are present in the Coast, St. Elias, Selkirk, and Rocky Mountains. Some other ranges have a few small glaciers, the largest of which are in southeastern Selwyn Mountains. Several small volcanoes near Mount Edziza, on lower Iskut River, and north of Terrace have been active within the last 2,000 years, erupting since the last retreat of the ice (Pl. VIII-20).

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IX. Economic Minerals of Western Canada

Introduction 490

Mineral Provinces 491

Metallic Deposits 494

Non-metallic Deposits 518

Fuels 522

Selected References 544



INTRODUCTION

The economic minerals of Western Canada are numerous and diverse (Table IX-1) as the region embraces a great many geological environments that range from those of the complexly deformed rocks in the mountains of the Western Cordillera to those of less disturbed and flat sedimentary sequences of the Eastern Cordillera and Interior Plains (*see* Chap. VIII). In 1966 the region yielded 93.6 per cent of all fuels produced in Canada, valued at $1,080 million, some 30 per cent of the industrial minerals and materials valued at $249 million, and about 15 per cent of the metals valued at $302 million. Total value of the economic minerals and materials produced in 1966 was $1,631 million representing 41 per cent of that for Canada as a whole. Most important commodities produced, in order of their total cumulative values, are oil, coal, natural gas, lead, gold, copper, silver, and zinc. Other products small in value yet not produced elsewhere in Canada are tin, mercury, and potash.

Mineral production, which began in 1836 with the mining of coal on Vancouver Island, was small until the discovery of placer gold on Pend-d'Oreille and Fraser Rivers from 1855 to 1860. Placer gold was foremost in production until 1884 when it was again surpassed by coal. In the late 19th and early 20th centuries the numerous lode gold and base metal mines developed in the Cordilleran Region dominated production for many years. Natural gas was first discovered at Medicine Hat in 1890, natural gas and naphtha at Turner Valley in 1913, and crude oil at Norman Wells in 1920. During 1953 the value of crude oil produced in Canada, largely from Western Canada, exceeded for the first time that of any other single commodity; nickel, gold, and copper were the leaders in previous years.

The Cordilleran Region, in comparison with other metal-bearing parts of Canada, has produced about 90 per cent of the lead, more than half the zinc, nearly half the silver, and nearly one fifth the copper produced to date. With recent development of several large copper deposits the proportion of Cordilleran production may well rise. Conversely, with only two major gold mines in operation the value of Cordilleran gold, which formerly represented more than a fifth of Canadian production, is decreasing whereas that from the Canadian Shield in Ontario and Quebec is increasing relatively. The Cordilleran Region provides the greater part of the antimony, bismuth, cadmium, and tungsten produced in Canada, and all the tin and mercury. Platinum, magnesium, palladium, molybdenite, chromite, manganese, cobalt, and selenium have also been produced in the past. Iron has become important recently through development of contact metamorphic deposits of magnetite. Since 1958, nickel valued at more than a million dollars annually has been produced.

The fuels—petroleum, natural gas, and coal—are found in considerable abundance in Western Canada, the region yielding some 98 per cent of the oil and gas, and 44 per cent of the coal produced in Canada during 1965.

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1 Authorship responsibilities for sections of this chapter are: Mineral provinces and metallic deposits by H. W. Little; Coal by B. A. Latour; Paleozoic and Mesozoic oil and gas fields by H. R. Belyea and D. F. Stott, respectively; Foothills oil and gas fields by R. J. W. Douglas. The accounts of non-metallic minerals were prepared by H. W. Little, D. F. Stott, and H. R. Belyea, or taken from previous editions.



IX

Economic Minerals of Western Canada

H.W. Little, H.R. Belyea, D.F. Stott, B.A. Latour, and R.J.W. Douglas[1]

Ironsides pit of Phoenix Copper mine, British Columbia.

The oil and gas occur mainly in the southern parts of the Interior Plains, the Devonian, Cretaceous, and Mississippian systems being most prolific, and from the thrust sheets of the Foothills of the Eastern Cordillera. Coal, of Cretaceous and Tertiary age, occurs at many places through the Cordillera and Plains, but the largest deposits and most mines are in the Eastern Cordillera and southern Plains.

Of the industrial minerals, the most valuable are asbestos, potash, sulphur, salt, and gypsum, whereas those of lesser importance are barite, fluorite, bentonite, sodium and magnesium sulphate, and diatomite. Large quantities of structural material such as cement, lime, sand and gravel, crushed rock, building stone, firebrick, common brick, and other clay products have been produced.

MINERAL PROVINCES

The economic minerals of the Cordilleran Region and Interior Plains include many of the metallic and industrial minerals, natural gas, petroleum, and coal. The region is divisible into two geologically distinct provinces—the Eastern and Western Mineral Provinces. Their boundary in the south is Rocky Mountain Trench and in the north follows close to the western boundary of the Eastern Cordilleran system (*see* Fig. IX-1 and Mineral Deposits of Canada, Map 1252A). The Eastern Province includes the Interior Plains, the Rocky Mountains and Foothills, Mackenzie Mountains, and much of Selwyn and Ogilvie Mountains; the Western Province embraces the remainder of the Canadian Cordillera.

Although similarities in stratigraphy and structure persist in places for some distance across the border between these provinces and similar types of mineral deposits are known in both, the differences between them are impressive. The Eastern Province is characterized by carbonate, sandstone, shale, and evaporites, unmetamorphosed or of very low grade metamorphism. In the eastern part, beneath the Interior Plains, the strata overlying the Precambrian crystalline basement are virtually undisturbed, comprising remnants of several depositional basins. In the smaller, western part, which forms the Eastern Cordillera, the sediments are folded and faulted and a few small intrusive bodies occur.

In contrast, the Western Province had a most complex history. The Proterozoic and early Paleozoic rocks which are exposed almost entirely in the eastern part are mainly miogeosynclinal facies. The late Paleozoic and early Mesozoic rocks are, however, of typical eugeosynclinal facies. In Coast and Omineca geanticlines the rocks are for the most part intensely folded and highly metamorphosed. Elsewhere intense deformation and high grade metamorphism are local, but the regional grade of metamorphism, although low, is generally much higher than that of the Eastern Province.

Orogenic movements occurred throughout and plutonic activity is recorded in the late Proterozoic, mid-Paleozoic, throughout the Mesozoic, and the early Tertiary. In the Mesozoic, linear basins of deposition were formed, some of which, towards the close were intermontane and continental, followed in the Tertiary by the extrusion of plateau basalt and the formation of volcanic cones.

TABLE IX-1 *Mineral production of Western Canada in 1966, and cumulative totals*

	Western Mineral Province		Eastern Mineral Province	
	1966 (thousands)	Cumulative totals (thousands)	1966 (thousands)	Cumulative totals (thousands)
FUELS				
Coal	31 tons	83,690 tons	6,607 tons	440,707 tons
	$ 300	$ 337,442	$ 22,131	$ 1,543,952
Liquid petroleum gases			51,190 bbl.	*
			$ 99,908	$ 418,790
Natural gas			1,326,195 Mcf	9,299,045 Mcf
			$ 171,598	$ 1,110,663
Petroleum crude			319,266 bbl.	3,158,806 bbl.
			$ 787,256	$ 7,311,831
METALS				
Antimony	1,406 lb.	48,592 lb.		
	$ 745	$ 13,982		
Bismuth	47 lb.	6,203 lb.		
	$ 199	$ 11,192		
Cadmium	1,257 lb.	37,549 lb.	125 lb.	377 lb.
	$ 3,222	$ 63,285	$ 301	$ 937
Chromite	—	1 ton		
	—	$ 32		
Cobalt	—	2 lb.		
	—	$ 0.4		
Copper	105,006 lb.	3,560,054 lb.		
	$ 56,014	$ 702,428		
Gold, lode	119 oz.	16,559 oz.	—	0.06 oz.
	$ 4,476	$ 486,295	—	$ 2
Gold, placer	42 oz.	16,262 oz.	0.18 oz.	41 oz.
	$ 1,554	$ 357,287	$ 7	$ 983
Iron concentrates	2,152 tons	18,102 tons		
	$ 20,779	$ 159,330		
Lead	226,438 lb.	15,335,202 lb.	211,687 lb.	492,394 lb.
	$ 36,656	$ 1,259,190	$ 31,640	$ 66,904
Magnesium	—	205 lb.		
	—	$ 88		
Manganese	—	2 tons		
	—	$ 33		
Mercury	—	4,171 lb.		
	—	$ 10,445		
Molybdenum	17,306 lb.	24,655 lb.		
	$ 28,072	$ 40,570		
Nickel	3,622 lb.	28,231 lb.		
	$ 3,104	$ 22,427		
Platinum and palladium	—	2 oz.		
	—	$ 167		
Selenium	—	1 lb.		
	—	$ 1		
Silver	9,712 oz.	606,768 oz.	22 oz.	940 oz.
	$ 13,554	$ 423,980	$ 31	$ 620
Tin	711 lb.	16,837 lb.		
	$ 917	$ 13,974		

TABLE IX-1 *Mineral production of Western Canada in 1966, and cumulative totals (conc.)*

	Western Mineral Province		Eastern Mineral Province	
	1966 (thousands)	Cumulative totals (thousands)	1966 (thousands)	Cumulative totals (thousands)
Tungsten (WO_3)	4,185 lb.	20,236 lb.		
	$ 5,000	$ 43,691		
Zinc	314,395 lb.	13,354,409 lb.	380,513 lb.	748,646 lb.
	$ 49,065	$ 1,232,880	$ 57,459	$ 102,666
INDUSTRIAL MINERALS				
Arsenious oxide	—	22,019 lb.		
	—	$ 273		
Asbestos	89 tons	591 tons		
	$ 15,071	$ 104,772		
Barite	22 tons	252 tons		
	$ 176	$ 2,942		
Diatomite	0.07 ton	4 tons		
	$ 4	$ 144		
Fluorspar	0.15 ton	36 tons		
	$ 5	$ 792		
Gypsum and anhydrite	—	1,247 tons		
	—	$ 6,323		
Hydro-magnesite	—	2 tons		
	—	$ 28		
Iron oxide and ochre	—	18 tons		
	—	$ 155		
Jade	12 lb.	238 lb.		
	$ 13	$ 109		
Mica	—	12,822 lb.		
	—	$ 186		
Perlite	—	1 ton		
	—	$ 11		
Sodium carbonate and magnesium sulphate	—	24 tons		
	—	$ 373		
Talc	—	2 tons		
	—	$ 35		
			Phosphate rock	— 4 tons
				— $ 17
			Potash	1,990 tons 5,029 tons
				$ 62,665 $ 173,917
			Salt	235 tons 2,752 tons
				$ 3,877 54,448
			Sodium sulphate	405 tons 5,411 tons
				$ 6,472 $ 73,909
			Sulphur, elemental	2,041 tons *
				$ 40,238 $ 332,226
			Gypsum and anhydrite	340 tons 6,559 tons
				$ 1,004 $ 28,884
STRUCTURAL MATERIALS				
Cement	708 tons	10,190 tons		
	$ 15,959	$ 159,543		
Clay products	$ 4,100	$ 60,093		
Lime and limestone	1,484 tons	*		
	$ 2,696	$ 41,105		
Sand and gravel	24,320 tons	*		
	$ 21,960	$ 164,267		
Stone	77 tons	1,154 tons		
	$ 215	$ 9,080		
			Cement	1,545 tons 25,780 tons
				$ 30,870 $ 425,498
			Clay products	$ 5,305 $ 134,002
			Lime and limestone	129 tons 3,154 tons
				$ 2,256 $ 41,082
			Sand and gravel	30,876 tons 522,526 tons
				$ 21,195 $ 320,778
			Stone	2,167 tons 17,909 tons
				$ 3,962 $ 35,676

Sources: Dominion Bureau of Statistics, British Columbia Department of Mines, and Mineral Resources Branch of Department Energy, Mines and Resources.

— No production

* Quantity not available

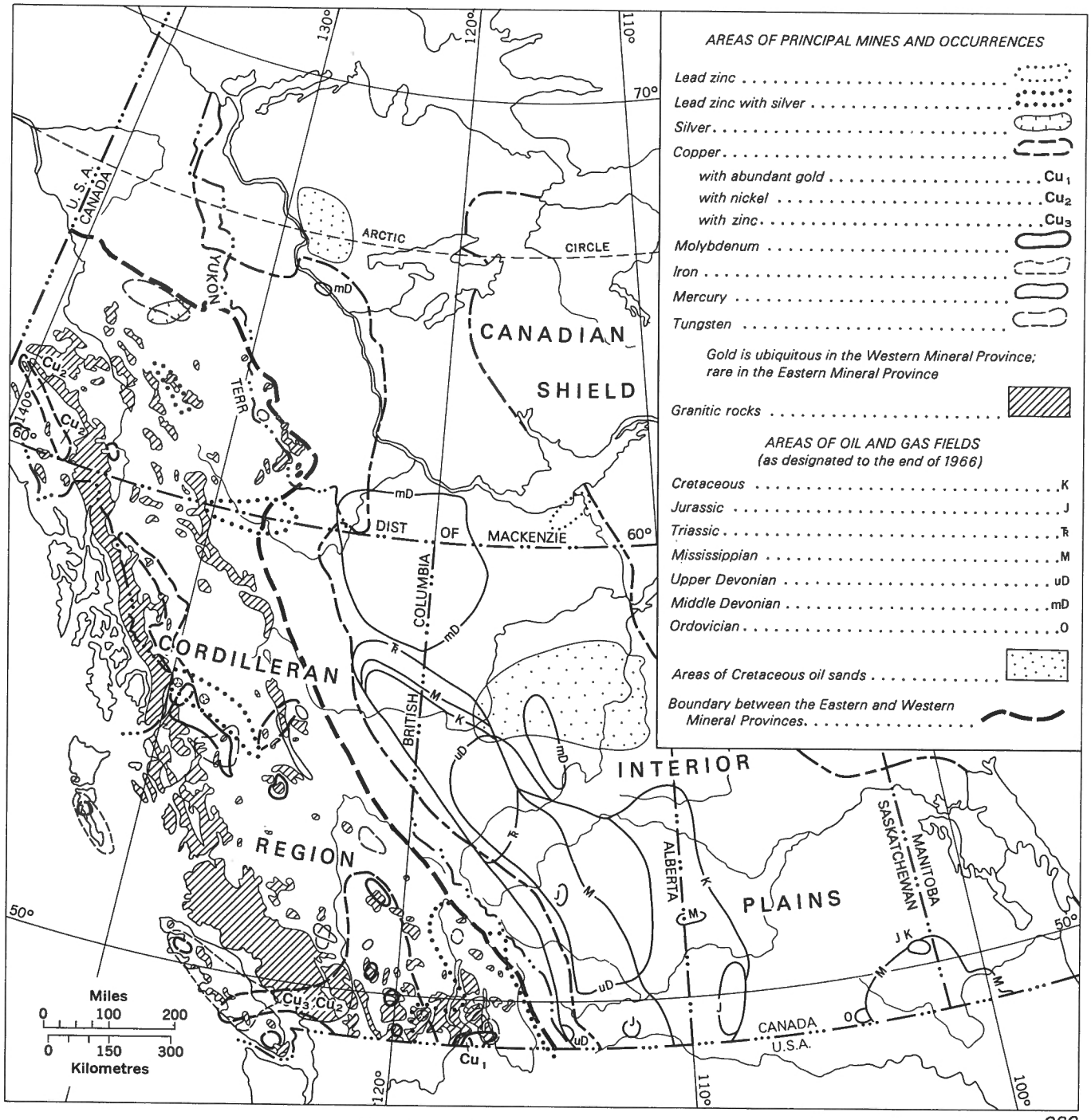


FIGURE IX-1. Mineral provinces of Western Canada. GSC

METALLIC DEPOSITS

Eastern Mineral Province

Deposits of this province are typical of an environment reflecting an epeirogenic tectonic history, simple deformation, and only local metamorphism and plutonism. Contact metamorphic, magmatic segregation, and hypothermal deposits are unknown, and mesothermal deposits are rare. Non-ferrous metallic deposits are epithermal veins or low temperature replacements in limestone or shale, and ferrous deposits are of sedimentary origin. The great bulk of the economic minerals (Table IX-1) are

petroleum, natural gas, and coal, and non-metallic deposits such as evaporites, saline lake deposits, or structural materials such as building and crushed stone, sand, gravel, cement, and bentonite.

Metallic mineral deposits with the important exception of Pine Point lie in the western part of the province. They are few in number, widely separated, and are not assigned to specific metallogenic provinces or epochs. Mineral properties and their production up to and including 1966 are given in Table IX-2.

Lode Deposits

Lead–Zinc–Silver

Lead–zinc–silver deposits (Fig. IX-2), some of which contain minor cadmium and gold, are classified as concordant or transgressive according to their relationships with the country rock.

Concordant Deposits

These are represented by Pine Point (10)[1], Monarch and Kicking Horse (98), and the Kootenay King (Fig. IX-2) mines.

The *Pine Point* region on the south shore of Great Slave Lake, District of Mackenzie, has a total of twenty-eight known orebodies ranging from a few thousand to 3 million tons and occurring in a belt 2 to 4 miles wide and 22 miles long (Fig. IX-3). To date, all production has come from two orebodies, each about 1,600 by 600 feet and 100 feet or more thick; a third is stripped in preparation for mining. Production began late in 1964, and by the end of 1965 more than 379 thousand tons of crude ore had been shipped to Trail, British Columbia and Kellogg, Idaho. Crude ore averaged nearly 50 per cent lead and zinc and the milling ore graded nearly 12 per cent.

The region is underlain by about 1,000 feet of Middle Devonian carbonates that rest on Precambrian crystalline basement and are succeeded unconformably by green, calcareous Upper Devonian shale. The carbonates grade northward into black shales and southward into evaporites. Regional dip is westerly at about 20 feet per mile (Campbell, 1957). Locally the beds have been folded into gentle flexures whose limbs are a few hundred to a few thousand feet long and dip up to 125 feet per mile. The fold axes plunge roughly S65°W, subparallel to the trace of faults in the Precambrian basement evident on aeromagnetic maps. Although post-Devonian movement on the faults cannot be proven, it is thought some small displacement may be responsible for the folds.

The ore occurs in the Presqu'ile Formation, in saccharoidal, vuggy dolomite that may have replaced reefal limestone. Some ore also occurs in a fine-grained granular dolomite that may underlie the Presqu'ile Formation but is mainly its stratigraphic equivalent. Other partly equivalent facies such as gypsiferous Nyarling Formation to the southeast and the limestones of the Sulphur Point Formation to the northwest are not mineralized. Primary ore at Pine Point (Schiller and Hornbrook, 1964) consists of disseminations, banded encrustations, and vug fillings of sphalerite and galena with minor marcasite, native sulphur, and bitumen. Secondary minerals are cerussite, smithsonite, and limonite (Bell, 1931) and occur to a depth of 100 feet. Bell suggests that the ore occupies zones of jointing along anticlinal crests, but Campbell (1957) considers the data to be too sparse to draw any firm conclusions as to structural control. In general, the orebodies lie in a narrow zone parallel to the fold axes, but primary control is the porosity of the dolomites of the Presqu'ile Formation.

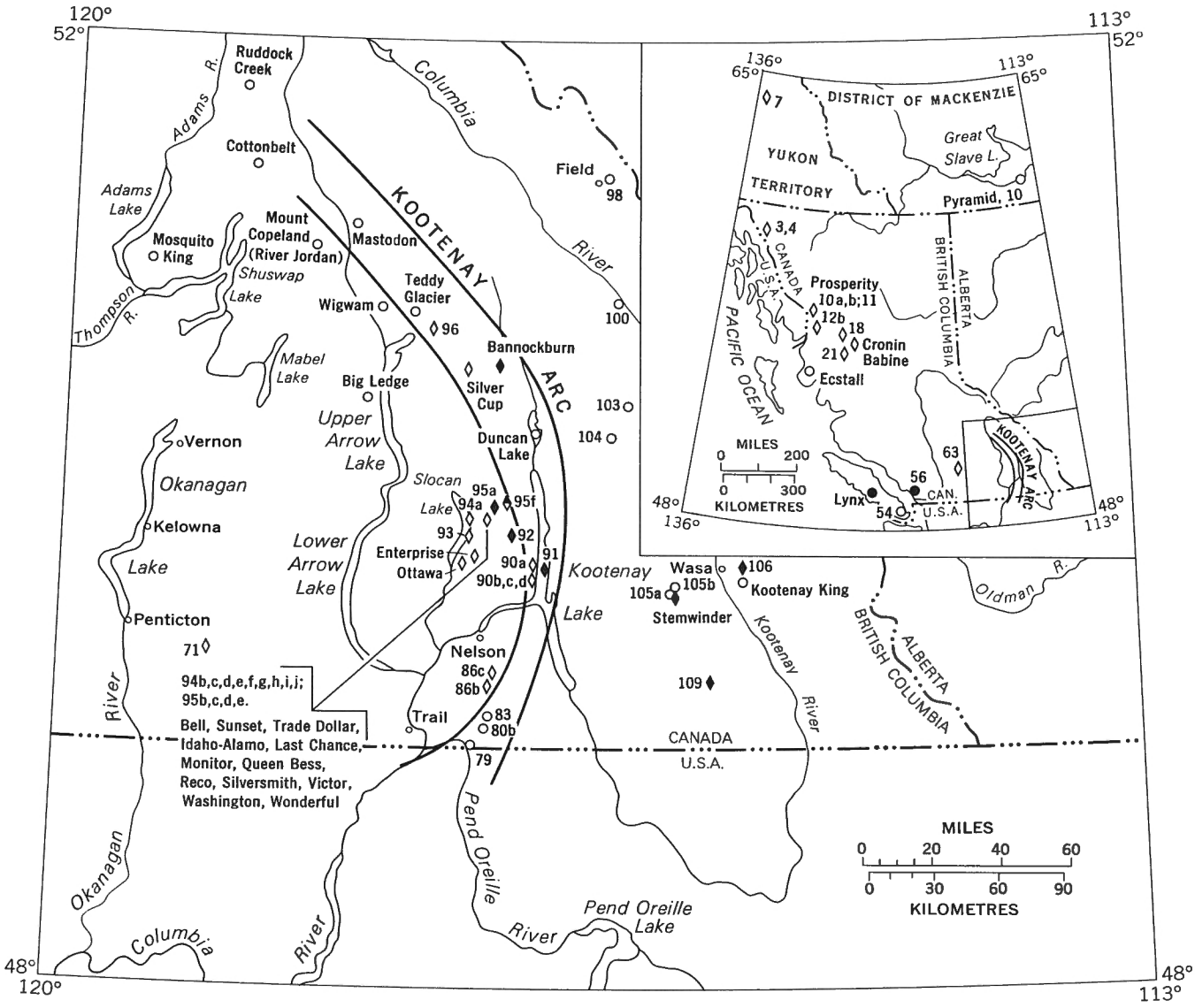
Four processes for the formation of the ore have been suggested by Bell: chemical precipitation during sedimentation; lateral secretion from cold meteoric waters; leaching of deep-seated rocks by artesian waters; and deposition from hydrothermal solutions of magmatic sources. The syngenetic origin is unlikely as lead isotope ages obtained by Folinsbee indicate ages for the ore of 200 to 300 million years—much younger than the Middle Devonian age of the Presqu'ile Formation. No igneous rocks are known to intrude the Paleozoic strata within many miles of the deposits. Bell and Campbell favour a hydrothermal origin, but Campbell indicates that some solution and redeposition by meteoric waters has occurred.

Monarch and Kicking Horse deposits (98), which occur near Field in the Rocky Mountains, contain zinc, lead, silver, and cadmium. They are concordant bodies within dolomite of the Middle Cambrian Cathedral Formation (Ney, 1957), but whether the dolomite is primary or related to the mineralization is not known. Ore minerals occur most commonly disseminated in the matrix of grey dolomite breccia and also in veinlets cutting white and black dolomite. Mineralization is confined between two horizons 200 feet apart in the lower part of the formation, and mainly within gentle anticlinal folds. The ore is apparently unrelated to intrusive rocks, the nearest of which comprise the Ice River complex some 12 miles to the south and furthermore are of mid-Paleozoic age, much older than the ore which seemingly is related to folds produced during Laramide Orogeny. Similar mineral occurrences are reported at about the same stratigraphic horizon to the east on Mount Eisenhower and Copper Mountain (J. D. Aitken).

Transgressive Deposits

The *Estella* mine (106) is near Wasa east of Rocky Mountain Trench. The ore occurs in replacement veins and fissure fillings and contains zinc, lead, silver, cadmium, gold, and some cobalt. The Estella and the Kootenay King, a nearby small concordant deposit, occur in the Aldridge Formation in the lower part of the Helikian Purcell succession. This region also contains a num-

[1] Number is that assigned the deposit on Mineral Deposits of Canada, Map 1252A.



INDEX TO NUMBERED DEPOSITS

10. Pine Point
7. United Keno
3. Tulsequah Chief
4. Big Bull
10a. Premier Border
10b. Silbak Premier
11. Dunwell
12b. Toric
18. Silver Standard
21. Duthie
54. Tyee
56. Britannia
63. Nicola (Planet)
71. Highland-Bell
79. Reeves-MacDonald
80b. Jersey
83. H.B.
86b. Yankee Girl
86c. Ymir
90a. Florence (Ainsmore)
90b. Highland
90c. Number One
90d. Yale (Highlander)
91. Bluebell
92. Cork-Province
93. Galena Farm
94a. Bosun
94b. Hewitt
94c. Ivanhoe
94d. Mammoth
94e. Richmond-Eureka
94f. Ruth-Hope
94g. Silversmith and Slocan Star
94h. Standard
94i. Van Roi
94j. Violamac (Victor)
95a. Lucky Jim
95b. Noble Five
95c. Payne
95d. Rambler-Cariboo
95e. Surprise
95f. Whitewater
96. Spider
98. Monarch and Kicking Horse
100. Silver Giant
103. Paradise
104. Mineral King
105a. North Star
105b. Sullivan
106. Estella
109. St. Eugene

Concordant dissemination . . . . . ○ Non-concordant dissemination . . . . . ● Transgressive (veins) . . . . . ◊
Transgressive (veins with considerable replacement) . . . . . ♦

Notes: Numbers are those assigned to deposits shown on Map 1252A; others are named on figure.
Properties from which lead and zinc are the principal product are underlined. GSC

FIGURE IX-2. Lead–zinc deposits of Western Canada, classified according to type (by H. W. Little).

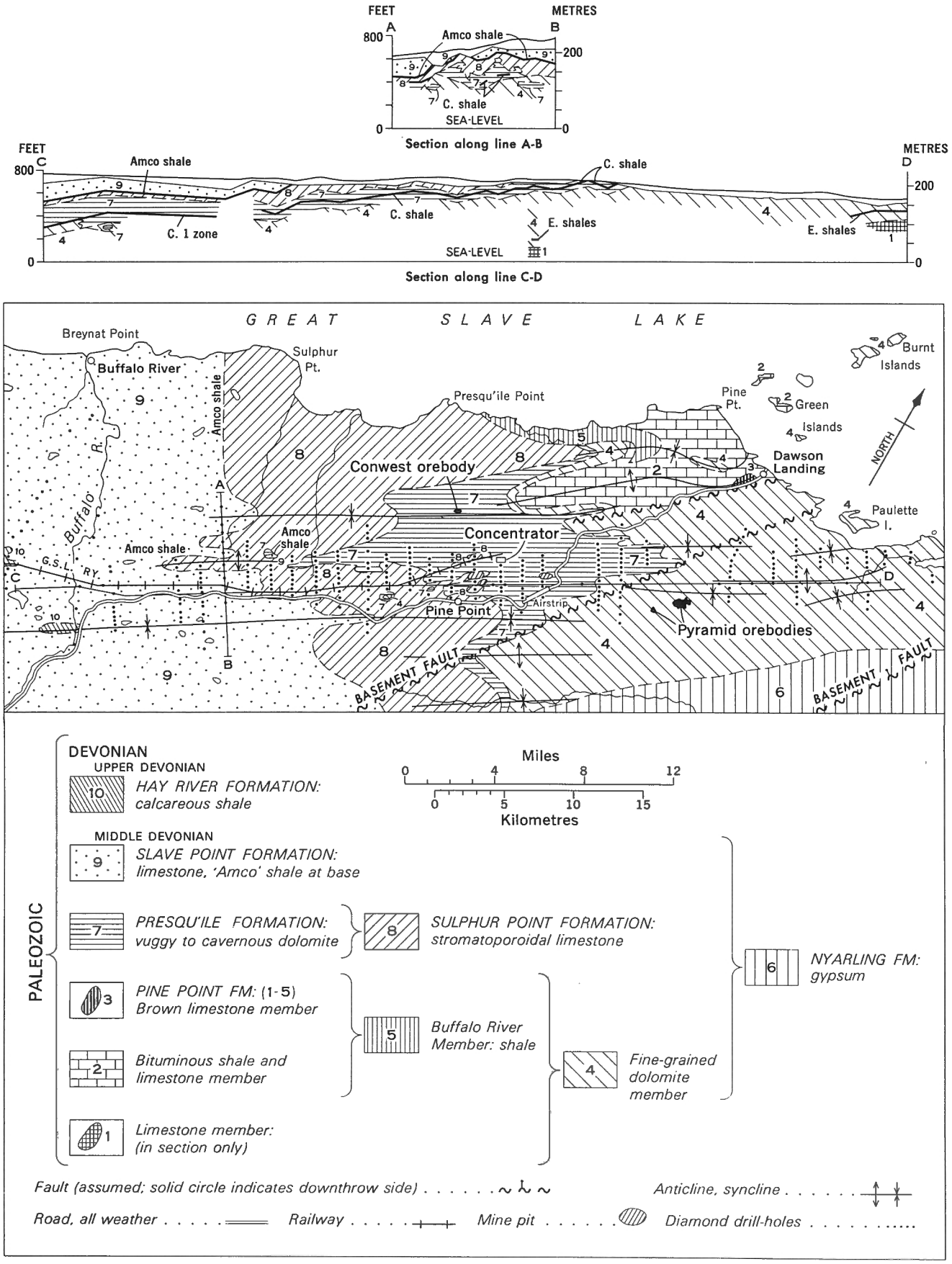


FIGURE IX-3. Geology in the vicinity of Pine Point lead–zinc deposits, District of Mackenzie (after A. W. Norris). GSC

ber of sparsely mineralized quartz veins. The age of the base metal deposits is not known, but some quartz veins occur in late Lower Cretaceous, or possibly younger, syenitic rocks.

Copper

Concordant Deposits

The *Redstone* deposit (Fig. IX-9) in Mackenzie Mountains (Green and Godwin, 1964) is a bedded copper deposit. It occurs (J. A. Coates) mainly in three persistent green carbonate-rich beds within a sequence of Hadrynian mudstone and siltstone. The sequence, which strikes north and dips steeply west, is disrupted by two major west-dipping thrust faults and numerous normal faults of small displacement. In decreasing order of abundance, ore minerals consist of pyrite, chalcopyrite, bornite, chalcocite, covellite, tennantite, and galena. Supergene malachite, azurite, native copper, and limonite are present in various amounts. The mineralization has been traced laterally for at least 4 miles. Grade averages 2.30 per cent copper across 6.5 feet at the north end and 3.74 per cent across 5.5 feet at the south end, but less than 1 per cent in the central part.

Sedimentary Deposits

Iron

The *Clear Hills* deposits are mainly oölitic hematitic sandstone and shale occurring in the marine Upper Cretaceous Kaskapau Formation at Clear Hills, Alberta. The largest deposit is estimated at more than 1,000 million tons with an average grade of 31 per cent iron; another deposit of slightly higher grade exceeds 100 million tons. Other deposits are smaller and of lower grade. Tests are being made to determine if a marketable concentrate can be produced by beneficiation. The main deposit, near Swift Creek, is 3 to 30 feet thick, underlies an area about 13 by 3 miles, and is covered by up to 200 feet of overburden (Kidd, 1959). The ferruginous sandstone contains oölites formed of concentric shells of goethite and quartz in a matrix of quartz, goethite, siderite, and chamosite (?), with a little apatite and sericite.

The *Snake River* deposit in northern Mackenzie Mountains (Green and Godwin, 1964) occurs in the Hadrynian Rapitan Formation. The iron-formation, which is 137 feet thick, consists of jasper and hematite, with thin intercalations of mudstone. Ore is estimated at 20 billion tons of which about 5 billion tons can be mined by open-pit methods. The grade averages 46 per cent Fe, not allowing for dilution by waste interbeds. The phosphorus content of 0.35 per cent is due to the presence of apatite.

Surficial Deposits

Gold

Placer gold has been produced on South Saskatchewan, Peace, McLeod, Athabasca, Bow, Oldman, and other rivers, but North Saskatchewan has been the chief source. It was at first presumed to be derived from the Precambrian Shield to the east through concentration from glacial till. In 1887, however, Tyrrell proposed a western source, the Selkirk Mountains, for the Cretaceous sediments of western Alberta, and implied that the gold placers resulted from reconcentration during erosion of the Cretaceous rocks. Placer mining in Alberta began in 1859 or 1860, reached peak production in 1896, and is still operating on a minor scale. Total recorded gold production is about 19,000 ounces.

Western Mineral Province

Deposits of the Western Mineral Province are diverse, as is characteristic of areas of great geological complexity (Table IX-1). Metallic deposits containing a great variety of metals, several of which may occur in a single deposit, are widespread. Many types are represented, including pyrometasomatic, vein, magmatic segregation, and replacement deposits ranging from hypothermal to epithermal. In the descriptions of properties only one or two examples of each class and type are given because of space limitations. Pegmatites containing beryl, muscovite, or lepidolite are known, but of these only mica has been produced. Deposits of gypsum and bentonite are rare. Production data are shown in Table IX-2.

Although basic contributions to the concept of metallogenic provinces and epochs were made more than 40 years ago by Schofield (1919) and Brock (Brock and Schofield, 1928), much remains to be done before a valid and detailed metallogenic picture can be drawn. Only recently has the complexity of the tectonic history of the Canadian Cordillera been fathomed. Dating many of the mineral deposits and relating each to some phase of this complex history remains to be done.

Metallogenic maps of Canada for uranium, beryllium, molybdenum, iron, manganese, and tin have been published, but compilation of other metals has been published only in preliminary form at small scale by Lang (1961). Lang concluded that other than a division into eastern and western Cordilleran provinces no subdivision was as yet justified. He pointed out that in the western province gold is widely distributed, copper nearly so, and lead and zinc occur in many scattered areas, although mainly in the southeast. Minor nickel, cobalt, and iron occurrences are fairly numerous in southern British Columbia, and iron is abundant on Vancouver Island.

More recent studies by R. Mulligan indicate that in the eastern part of the Western Mineral Province

TABLE IX–2 | *Total production from main lode metal mines and industrial mineral properties of Western Canada, to 1966 inclusive*

No.[1]	Property[2]	Area or Camp	Location[3]	Years of Production[4]	Ore or Mineral — Thousand tons	Gold — Thousand oz.	Silver — Million oz.	Copper — Million lb.	Lead — Million lb.	Zinc — Million lb.	Other metals or minerals — Million lb. (mlb) or million tons (mt)
	District of Mackenzie										
1	Canada Tungsten	Flat River, N.W.T.	61°128°NE	1962-66	327	—	—	a	—	—	WO_3 8.29 mlb
10	Pine Point	Great Slave Lake	60°114°NE	1964-66	2,197	—	a	—	382	676	Cd 1.26 mlb
	Yukon Territory										
7	United Keno	Mayo, Yukon	63°135°NE	1913-66	2,748[5]	—	146.7	—	464	274	Cd 3.59 mlb
9, 10	Whitehorse Copper Belt (Pueblo, etc.)	Whitehorse, Yukon	60°135°NW	1897-1930	300	—	—	13	—	—	
	British Columbia										
2	Polaris-Taku	Taku River	58°133°NW	1938-51	753	231	a	a	—	—	
3, 4	Tulsequah Chief and Big Bull	Taku River	58°133°NW	1951-57	1,029	94	3.4	27	26	125	Cd 0.45 mlb
5	Cassiar Asbestos	McDame Creek	59°129°SW	1952-66	6,033	—	—	—	—	—	Asbestos 0.59 mt
9	Big Missouri	Portland Canal	56°130°SE	1927-42	848	58	0.1	—	a	a	
10a	Premier Border	Portland Canal	56°130°SE	1950-53	42	3	0.1	—	4	4	
10b	Silbak-Premier	Portland Canal	56°130°SE	1918-66	4,664	1,813	40.9	4	59	16	Cd 0.18 mlb
11	Dunwell	Portland Canal	55°129°NW	1926-41	50	10	0.3	a	2	2	
12a	Dolly Varden	Portland Canal	55°129°NW	1919-40	37	a	1.4	a	a	—	
12b	Toric	Portland Canal	55°129°NW	1928-59	1,379	a	18.6	—	10	1	
13	Outsider	Portland Canal	55°130°NE	1906-28	139	a	a	5	—	—	
14	Bonanza	Portland Canal	55°129°SW	1928-35	724	3	0.3	32	—	—	
15a	Golskeish	Portland Canal	55°129°SW	1918-29	51	5	a	—	—	—	
15b	Granby Point	Portland Canal	55°129°SW	1917-38	62	6	0.2	a	a	—	
15c	Hidden Creek	Portland Canal	55°129°SW	1914-36	23,948	121	6.6	709	—	—	
18	Silver Standard	Hazelton	55°127°SW	1913-65	216	15	7.4	a	17	26	Cd 0.33 mlb
19	Rocher Déboulé	Hazelton	55°127°SW	1915-52	53	4	a	6	a	a	
20	Red Rose	Hazelton	55°127°SW	1942-54	118	a	a	a	—	—	WO_3 2.21 mlb
21	Duthie	Smithers	54°127°NE	1923-53	65	3	1.6	a	6	5	Cd 0.01 mlb
22	Granisle	Babine Lake	54°126°NE	1966	206	1	a	2	—	—	
23	Pinchi Lake	Fort St. James	54°124°NE	1940-44	692	—	—	—	—	—	Hg 4.0 mlb
24	Surf Point	Porcher Island	54°130°SW	1919-39	68	20	a	a	—	—	

TABLE IX–2 *Total production from main lode metal mines and industrial mineral properties of Western Canada, to 1966 inclusive (cont.)*

No.[1]	Property[2]	Area or Camp	Location[3]	Years of Production[4]	Ore or Mineral Thousand tons	Gold Thousand oz.	Silver Million oz.	Copper Million lb.	Lead Million lb.	Zinc Million lb.	Other metals or minerals Million lb. (mlb) or million tons (mt)
26	Jedway	Moresby Island	52°131°SE	1962-66	3,296	—	—	—	—	—	Fe conc. 1.80 mt
27	Surf Inlet	Princess Royal Island	53°128°SW	1902-43	1,012	389	0.2	6	—	—	
28	Endako	Endako	54°125°SE	1965-66	7,848						Mo 5.15 mlb
31	Aurum (Island Mountain)	Wells, Cariboo	53°121°SW	1934-66	1,056	486	0.7	—	—	—	
32	Cariboo Gold Quartz	Wells, Cariboo	53°121°SW	1933-58	1,855	705	0.6	—	—	—	
33	Boss Mountain	100 Mile House	52°120°SW	1965-66	658	—	—	—	—	—	Mo 3.53 mlb
34	Minto	Bridge River	50°122°NW	1934-40	89	18	0.1	a	a	—	
35	Wayside	Bridge River	50°122°NW	1915-37	43	5	a	—	—	—	
36	Bralorne-Pioneer	Bridge River	50°122°NW	1899-1966	7,558	3,954	0.9	—	—	—	Minor WO_3
37	Iron Mike	Sayward, V.Is.	50°125°SW	1966	150	—	—	—	—	—	Fe conc. 0.09 mt
38	Coast Copper (Old Sport)	Benson River, V.Is.	50°127°SE	1962-66	1,275	83	0.1	41	—	—	Fe conc. 0.16 mt
39	Yreka	Quatsino, V.Is.	50°127°SW	1902-66	79	1	0.1	5	—	—	
40	Empire	Benson River, V.Is.	50°127°SE	1957-66	3,611	—	—	—	—	—	Fe conc. 1.76 mt
41	Nimpkish	Nimpkish Lake, V.Is.	50°126°SW	1959-63	2,392	—	—	—	—	—	Fe conc. 1.0 mt
42	Zeballos Iron (F.L.)	Zeballos, V.Is.	50°126°SW	1962-66	1,227	—	—	—	—	—	Fe conc. 0.94 mt
43a	Mount Zeballos	Zeballos, V.Is.	50°126°SW	1933-44	82	57	a	a	a	a	
43b	Privateer	Zeballos, V.Is.	50°126°SW	1934-53	314	172	a	a	—	—	
44a	Central Zeballos	Zeballos, V.Is.	50°126°SW	1938-47	58	42	a	a	a	a	
44b	Spud Valley	Zeballos, V.Is.	50°126°SW	1936-51	210	54	a	a	a	a	
45	Hualpai	Nootka Sound, V.Is.	49°126°NE	1959	62	—	—	—	—	—	Fe conc. 0.02 mt
46	Argonaut (Iron Hill)	Campbell River, V.Is.	49°125°NW	1951-57	4,109	—	—	—	—	—	Fe conc. 2.16 mt
47	Domineer (Mount Washington)	Courtenay, V.Is.	49°125°NE	1961-66	373	4	0.2	7	—	—	
48a	Cornell	Texada Island	49°124°NW	1897-1919	45	15	a	3	—	—	
48b	Marble Bay	Texada Island	49°124°NW	1899-1929	314	50	0.4	15	—	—	
48c	Vananda (Little Billie)	Texada Island	49°124°NW	1896-1952	70	12	a	2	—	—	
49	Texada	Texada Island	49°124°NW	1885-1966	12,181	14	0.2	19	—	—	Fe conc. 6.21 mt
51	Indian Chief	Clayoquot, V.Is.	49°126°SE	1904-38	81	1	0.1	2	—	—	
52	Brynnor	Kennedy Lake, V.Is.	49°125°SE	1962-66	3,915	—	—	—	—	—	Fe conc. 2.94 mt
53	Blue Grouse	Cowichan Lake, V.Is.	48°124°NE	1917-60	275	a	a	15	—	—	
54	Tyee	Mount Sicker, V.Is.	48°123°NE	1901-51	221	27	0.5	14	a	4	Cd 0.01 mlb
55	Sunro	Jordan River, V.Is.	48°124°SE	1962-66	520	3	a	15	—	—	
56	Britannia	Vancouver	49°123°NE	1905-66	48,218	484	5.2	1,032	34	271	Cd 0.91 mlb

57	Craigmont	Merritt	50°120°SW	1961-66	8,634	—	a	266	—	—	
58	Bethlehem	Merritt	50°120°SW	1963-66	7,242	9	0.4	95	—	—	Mo 0.04 mlb
59	Vidette	Clinton	51°120°SW	1933-40	54	30	a	a	a	—	
60	Windpass	Kamloops	51°120°SE	1916-44	103	34	a	a	—	—	
61	Iron Mask	Kamloops	50°120°NE	1901-28	182	4	a	5	—	—	
62	Falkland Gypsum	Kamloops	50°119°SW	1925-56	1,247	—	—	—	—	—	Gypsum 1.25 mt
63	Nicola (Planet)	Merritt	50°120°SE	1926	79	8	0.2	a	2	1	Minor Bi
64	Pride of Emory	Hope	49°121°SW	1958-66	2,370	a	a	14	—	—	Ni 31.45 mlb
65	Copper Mountain	Princeton	49°120°SW	1917-62	34,775	188	4.4	613	—	—	
66a	Hedley Mascot	Hedley	49°120°SE	1936-49	682	223	a	2	—	—	
66b	Nickel Plate	Hedley	49°120°SE	1904-58	3,315	1,359	0.1	2	—	—	
67	French	Hedley	49°120°SE	1954-61	62	33	a	—	—	—	
69	Dividend-Lakeview	Osoyoos	49°119°SE	1907-49	123	16	a	a	a	a	
70	Fairview (Morning Star)	Fairview	49°119°SW	1933-41 1947-61	122	14	0.2	a	a	a	0.37 mt silica flux containing low values in Au
71	Highland-Bell	Beaverdell	49°119°SE	1901-66	387	11	27.0	—	19	23	Cd 0.08 mlb
72	Cariboo-Amelia	McKinney	49°119°SE	1894-1962	137	82	a	—	a	a	
73a	Mother Lode	Greenwood	49°118°SW	1900-62	4,317	173	0.7	77	—	—	
73b	Sunset	Greenwood	49°118°SW	1900-18	120	5	a	2	—	—	
74a	B.C.	Greenwood	49°118°SW	1900-19	103	1	a	9	—	—	
74b, f, g, h	Phoenix Copper (Knob Hill-Ironsides, Idaho, Brooklyn-Stemwinder, Snowshoe, Rawhide)	Greenwood	49°118°SW	1900-66	19,855	830	4.9	404	a	—	
74c	Dentonia	Greenwood	49°118°SW	1900-48	133	38	0.2	—	a	a	
74d	Emma	Greenwood	49°118°SW	1901-21	256	7	a	5	—	—	
74e	Oro Denero	Greenwood	49°118°SW	1903-17	136	3	a	4	—	—	
74i	Winnipeg	Greenwood	49°118°SW	1900-40	59	12	a	a	—	—	
75	Rock Candy	Grand Forks	49°118°NW	1918-42	36	—	—	—	—	—	Fluorspar 0.04 mt
76	Union	Franklin	49°118°NE	1913-46	189	55	1.4	a	a	1	
77	Velvet	Rossland	49°117°SW	1901-64	98	20	a	3	a	a	
78a	Centre Star group (Centre Star, War Eagle, Le Roi, Josie)	Rossland	49°117°SW	1893-1942	5,915	2,706	3.3	118	—	—	
78b	Red Mountain (Coxey)	Rossland	49°117°SW	1966	79	—	—	—	—	—	Mo 0.53 mlb
79	Reeves-Macdonald	Salmo	49°117°SE	1949-66	5,365	—	0.4	—	108	376	Cd 2.21 mlb
80a	Emerald, Dodger, and Feeney	Salmo	49°117°SE	1944-58	1,067	—	—	—	—	—	WO_3 13.74 mlb
80b	Jersey	Salmo	49°117°SE	1949-66	6,246	—	0.6	—	231	490	Cd 3.73 mlb
81	Bayonne	Kootenay Lake	49°116°SW	1935-51	89	42	0.1	—	a	a	
82a	Gold Belt	Salmo (Sheep Creek)	49°117°SE	1934-51	257	80	a	—	—	—	

TABLE IX–2 *Total production from main lode metal mines and industrial mineral properties of Western Canada, to 1966 inclusive (conc.)*

No.[1]	Property[2]	Area or Camp	Location[3]	Years of Production[4]	Ore or Mineral Thousand tons	Gold Thousand oz.	Silver Million oz.	Copper Million lb.	Lead Million lb.	Zinc Million lb.	Other metals or minerals Million lb. (mlb) or million tons (mt)
82b	Kootenay Belle	Salmo (Sheep Creek)	49°117°SE	1904-50 1961-65	333	110	a	—	—	—	Silica flux 0.03 mt
82c	Reno	Salmo (Sheep Creek)	49°117°SE	1906-50	430	232	0.2	—	—	—	
82d	Sheep Creek	Salmo (Sheep Creek)	49°117°SE	1902-50 1961-62	719	304	0.1	—	—	—	Silica flux 0.04 mt
83	H.B.	Salmo	49°117°SE	1912-66	5,249	a	0.8	—	86	456	Cd 3.76 mlb
84a	Arlington	Salmo	49°117°SE	1900-66	60	53	0.1	—	2	1	
84b	Second Relief	Salmo	49°117°SE	1902-59	228	99	a	a	a	a	
85	Hunter V and Double Standard	Ymir	49°117°SE	1902-29	63	1	0.3	—	—	—	
86a	Centre Star (Wesko)	Ymir	49°117°SE	1936-50	57	12	0.1	—	2	1	
86b	Yankee Girl	Ymir	49°117°SE	1907-51	408	124	0.7	—	14	14	Cd 1.88 mlb
86c	Ymir	Ymir	49°117°SE	1899-1950	367	110	0.5	—	11	2	
87a	Athabasca	Nelson	49°117°SE	1898-1943	46	20	a	a	a	a	
87b	Granite Poorman	Nelson	49°117°SE	1890-1954	200	65	a	a	a	a	
87c	Silver King	Nelson	49°117°SE	1889-1958	222	a	4.4	15	a	a	
88	Queen Victoria	Nelson	49°117°SE	1907-61	49	a	a	2	—	—	
89	Molly Gibson	Nelson	49°117°NE	1899-1950	62	a	1.0	—	5	a	
90a	Florence (Ainsmore)	Ainsworth	49°116°NW	1899-1960	223	a	0.4	—	25	6	Cd 0.02 mlb
90b	Highland	Ainsworth	49°116°NW	1890-1954	98	a	0.3	—	21	1	Minor Cd
90c	Number One	Ainsworth	49°116°NW	1889-1929	40	a	2.0	—	a	—	
90d	Yale (Highlander)	Ainsworth	49°116°NW	1889-1964	436	a	0.9	—	46	11	Minor Cd & Sb
91	Bluebell	Kootenay Lake	49°116°NW	1895-1966	4,078	—	5.5	4	400	419	Cd 1.92 mlb
92	Cork-Province	Kaslo	49°117°NE	1900-66	211	a	0.5	—	13	20	Cd 0.15 mlb
93	Galena Farm	Slocan	49°117°NE	1900-66	90	a	0.6	—	6	10	Minor Cd
94a	Bosun	Slocan	49°117°NE	1898-1962	70	a	1.9	—	11	7	Minor Cd
94b	Hewitt	Slocan	49°117°NE	1900-66	117	a	1.8	—	4	5	Minor Cd
94c	Ivanhoe	Slocan	49°117°NE	1895-1935	44	a	0.5	—	5	1	
94d	Mammoth	Slocan	49°117°NE	1925-66	94	a	1.0	—	7	10	Cd 0.04 mlb
94e	Richmond-Eureka	Slocan	49°117°NE	1896-1961	40	a	0.8	—	5	2	Minor Cd
94f	Ruth-Hope	Slocan	49°117°NE	1895-1962	66	a	2.5	—	22	4	Minor Cd
94g	Silversmith and Slocan Star	Slocan	49°117°NE	1893-1965	391	1	7.3	—	71	26	Cd 0.04 mlb
94h	Standard	Slocan	49°117°NE	1905-58	806	a	8.7	—	83	108	Cd 0.13 mlb

94i	Van Roi	Slocan	49°117°NE	1893-1958	314	a	2.8	—	18	17	Cd 0.04 mlb
94j	Violamac (Victor)	Slocan	49°117°NE	1923-66	163	2	4.1	—	47	31	Cd 0.18 mlb
95a	Lucky Jim	New Denver	50°117°SE	1893-1959	1,175	a	0.6	—	8	176	Cd 0.43 mlb
95b	Noble Five	Slocan	49°117°NE	1893-1956	44	a	0.5	—	5	4	Minor Cd
95c	Payne	Slocan	50°117°SE	1893-1939	122	—	3.7	—	38	2	
95d	Rambler-Cariboo	Slocan	50°117°SE	1895-1951	209	a	3.5	a	24	6	Cd 0.01 mlb
95e	Surprise	Slocan	49°117°NE	1893-1929	49	a	1.9	—	12	8	
95f	Whitewater	Kaslo	50°117°SE	1892-1956	501	2	3.5	—	31	51	Cd 0.08 mlb
96	Spider	Lardeau	50°117°NW	1911-58	140	12	1.7	a	24	25	Cd 0.13 mlb Sb 0.02 mlb
97	Meridian	Lardeau	50°117°NW	1903-41	98	17	a	—	—	—	
98	Monarch and Kicking Horse	Field	51°116°SE	1888-1953	813	—	0.8	—	102	157	Cd 0.06 mlb
99	Parson	Spillimacheen	51°116°SW	1940-65	199	—	—	—	—	—	Barite 0.20 mt
100	Silver Giant	Spillimacheen	50°116°NE	1908-57 1959-66	927	a	0.6	a	65	7	Cd 0.02 mlb Sb 0.04 mlb Barite 0.04 mt
101	Brisco	Spillimacheen	50°116°NE	1945-66	—	—	—	—	—	—	Barite 0.09 mt
102	Western Gypsum	Windermere	50°115°SW	1950-66	1,951	—	—	—	—	—	Gypsum 1.95 mt
103	Paradise	Toby Creek	50°116°SE	1901-52	71	a	0.7	—	16	8	Minor Cd
104	Mineral King	Toby Creek	50°116°SE	1954-66	2,202	—	1.7	1	77	184	Cd 0.63 mlb Barite 0.02 mt
105a	North Star	Kimberley	49°115°NW	1895-1929	68	—	1.3	—	48	a	
105b	Sullivan	Kimberley	49°115°NW	1900-66	99,371	4	217.2	6	13,035	11,278	Sn 17.41 mlb Cd 1.72 mlb Iron Sinter 0.55 mt
106	Estella	Cranbrook	49°115°NW	1951-66	80	a	0.1	—	8	16	Cd 0.01 mlb
108	Mayook	Cranbrook	49°115°SW	1926-54	111	—	—	—	—	—	Gypsum 0.11 mt
109	St. Eugene	Cranbrook	49°115°SW	1899-1929	1,062	3	5.9	—	249	32	

Most data have been supplied by the British Columbia Department of Mines and Petroleum Resources

[1] Number of property is from Map 1252A, Mineral Deposits of Canada (in folio)

[2] Production includes all known production by all owners, past and present

[3] Latitude and longitude of southeast corner of one-degree quadrangle and quarter of quadrangle in which property is situated

[4] Years of earliest and latest recorded production

[5] Tonnage prior to 1947 is not included

a Less than one-half of smallest unit shown in column

— No production

and in the Eastern Mineral Province zinc is almost everywhere associated with lead. In the central and western parts of the Western Mineral Province, however, zinc is much less abundant, lead even less so, and in some deposits zinc is affiliated with copper rather than lead. Copper, on the other hand, is minor in the eastern lead–zinc province, but widespread and abundant in the volcanic, eugeosynclinal region.

Mulligan (1966) has outlined a tin–beryllium–tungsten–molybdenum metallogenic province of lithophile affinity that occupies a belt lying west of Rocky Mountain Trench between the International Boundary and the Yukon. In the northwest tin and tungsten are also present, associated with small granitic intrusions that extend many miles eastward from Clear Creek, Yukon (Bostock *in* Little, 1959).

In central and western British Columbia the association of large, low-grade molybdenum deposits with late, leucocratic biotitic quartz monzonite facies of granitic intrusive complexes is apparent. This association is also evident for porphyry copper deposits. The age of the older facies of these complexes ranges from Jurassic to Upper Cretaceous, but that of the late leucocratic facies has not yet been established except in Tulsequah area where it is related to flows and pyroclastics of the late Upper Cretaceous and Early Tertiary Sloko Group (J. G. Souther).

Lode Deposits

Lead–Zinc–Silver

A great number of lead–zinc deposits that occur in southeastern British Columbia (Fyles, 1966) form part of a productive area also embracing Idaho and western Montana (Fig. IX-2). In the Purcell anticlinorium the Sullivan (105b), North Star (105a), and St. Eugene (109) deposits are of Precambrian age. Others considered to be Mesozoic occur along the Kootenay Arc from Salmo area in the south to the Big Bend of Columbia River in the north. Lead–isotope ratios indicate they have been contaminated by material derived from the Precambrian deposits (Sinclair, 1966). Muraro (1966) has also shown that lead–zinc deposits of the Kootenay Arc were formed before folding and metamorphism occurred, the folding having affected their shape in areas where flowage took place, and the metamorphism altering their texture. Deposits on the eastern flank of Kootenay Arc are for the most part concordant replacement, whereas those in the inner or concave part are mainly vein deposits most of which, in Slocan district, carry a much higher content of silver than the concordant types.

North of the Kootenay Arc from the Big Bend of Columbia River to northern British Columbia lead–zinc deposits are sparse. Veins with lead and zinc occur at Rancheria and Logjam Creek near the Yukon border. Replacement deposits occur farther north at Ketza River and Ross River, and at Keno Hill the veins, rich in silver, mark the north end of the lead–zinc belt.

Lead and zinc also occur in western British Columbia in complex vein deposits that yield mainly silver, the Toric (12b), Silver Standard (18), and Highland–Bell (71) mines, or with gold, silver, and copper as at Silbak Premier (10b). Zinc, with relatively little lead, is affiliated with copper in a few deposits—Britannia (56), Tyee (54), Tulsequah Chief (3), and Big Bull (4).

Fyles (1966) has classified lead–zinc deposits into concordant and transgressive types, but pointed out that both types may be represented in the same deposit, as in the Whitewater mine of the Slocan area.

Concordant Deposits

Except for Pine Point and a few deposits in Aiken Lake area, concordant lead–zinc deposits occur in the southeastern metallogenic province. Fyles subdivided the concordant deposits into Sullivan, Metaline, Salmo, and Shuswap subtypes which are, respectively, hypothermal replacements of argillaceous beds; mesothermal replacements of dolomitized limestone affected only by broad folding and faulting; mesothermal replacements of dolomitized limestone isoclinally folded and sheared; and replacements of marble and schist in complexly folded regions of high grade regional metamorphism.

The *Sullivan subtype* is unique, represented only by the *Sullivan* (105b) mine, the largest lead–zinc mine in the Cordillera (Table IX-2), which alone has produced nearly one half the metal wealth of British Columbia. It is of Precambrian age and comprises high temperature replacement of thin-bedded argillite of the Helikian Aldridge Formation. The orebody lies on the east side of the Purcell anticlinorium, whose limbs are characterized by gentle dips and minor warps (Freeze, 1966). The hanging-wall sediments are albitized and locally chloritized and tourmalinized. The footwall sediments contain a few conglomerate lenses and are heavily tourmalinized (Fig. IX-4). The ore preferentially replaces certain argillaceous members, but in places exhibits marked cross-cutting relationships. The orebody lies between the Kimberley and Hidden Hand faults which strike easterly, dip north, and have normal displacement, the former having a stratigraphic throw of about 10,000 feet. Steep north-trending normal faults displace the orebody a few tens of feet.

Based on the Houtermans lead-evolution model, the age of Sullivan lead was estimated to be 1,250 m.y. by Leech and Wanless (1962) and 1,340 m.y. by Sinclair (1966). As the lead–isotope ratios are virtually constant throughout the orebody a single period of mineralization from a single source is indicated, although some later redistribution may have occurred. Like Swanson and Gunning (1945), Leech and Wanless, and Sinclair favour an epigenetic source related to the source magma of Moyie Intrusions. Freeze (1966) believes that evidence indicates that the Sullivan is hydrothermal rather than syngenetic reconstituted by regional metamorphism. Although high

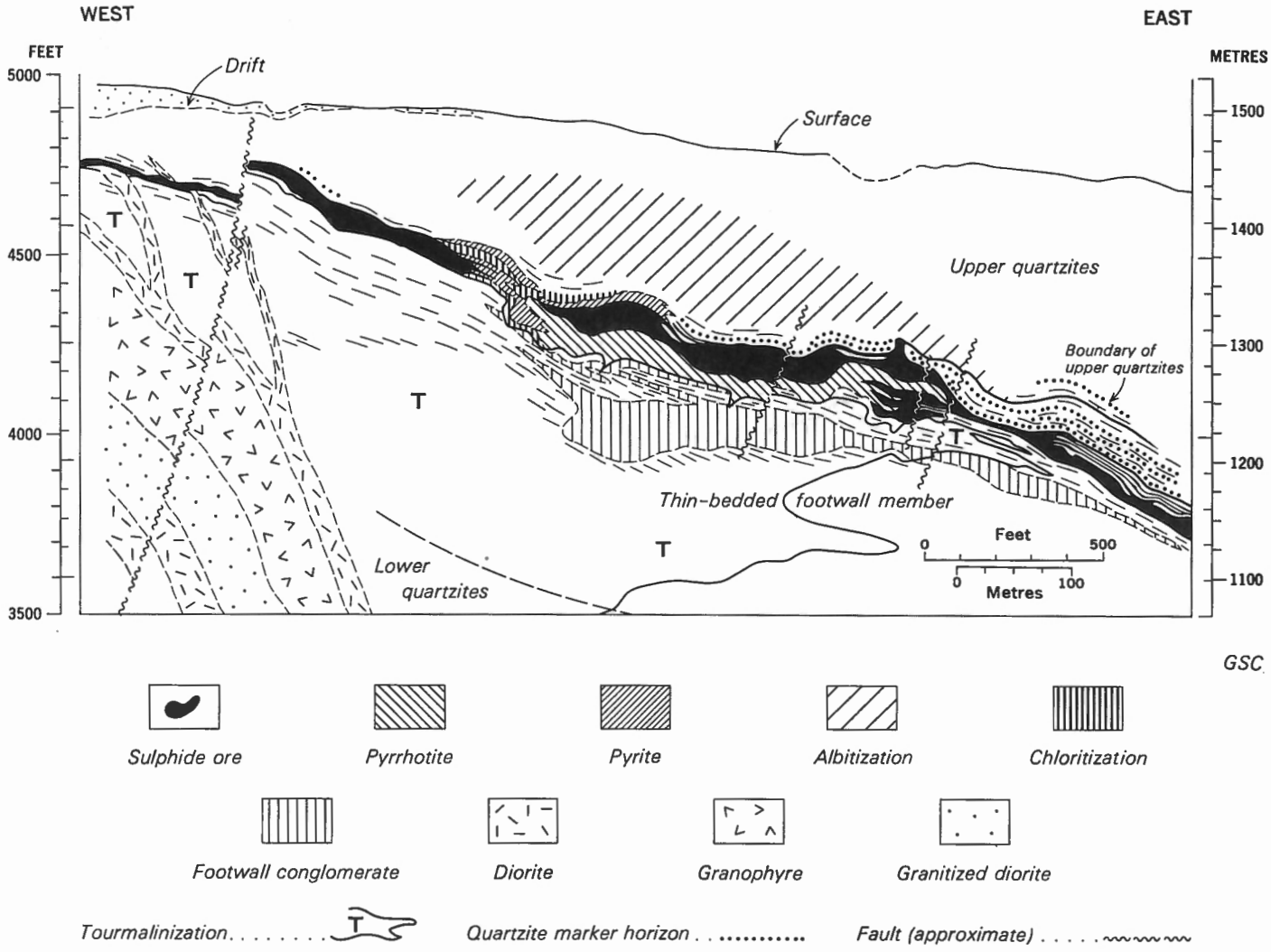


FIGURE IX-4. Cross-section showing conformable nature of Sullivan orebody, British Columbia (after A. C. Freeze).

temperature minerals such as garnet, tremolite, scapolite, and cordierite occur in the orebody and minimum temperatures of formation of 460°C for deposition of the sphalerite have been determined, the sediments barely a few feet away have reached only the chlorite–albite stage of regional metamorphism. Tourmalinization preceding mineralization is apparently younger than the Moyie sills. A lamprophyre dyke dated as 765 m.y. old intruded the ore zone during the rather lengthy period of mineralization.

The central part of the ore zone consists of a mass, roughly rectangular, dominantly of pyrite in the west and pyrrhotite in the east. Around this "iron core" the orebody is for the most part comparatively rich in lead, silver, tin, and arsenic, but towards the periphery zinc and antimony predominate. Boulangerite and other minor antimony minerals are disseminated mainly in the main ore zone but also in intersecting open fissures. Antimony is a substantial byproduct of the Sullivan mine. Cassiterite occurs, with iron sulphides and minor galena and sphalerite, in a vein that strikes northerly and dips 80 to 85°E, and intersects the main ore zone. However, cassiterite is found also at the base of the main orebody about the "iron core." The contact relationships between the vein and the orebody are vague. About 600 tons per year of tin concentrate is produced.

Metaline subtype is named after Metaline District, Washington. Deposits, which are concordant, occur in unmetamorphosed carbonate rocks either gently folded and faulted or brecciated. Examples are the Monarch and Kicking Horse (98), and Pine Point (10) mines in the Eastern Province.

Salmo subtype occurs in an environment of isoclinal folding with superimposed folding and shearing, and low greenschist facies grade of regional metamorphism. An example is the *Jersey* (80b). The host rock is Lower Cambrian limestone in a miogeosynclinal succession of

pelitic, arenaceous, and calcareous rocks. The orebodies are encased in dolomitic envelopes within the limestone and commonly skarn forms the footwall. The limestone and adjacent sediments are recumbently folded and disrupted by thrust faults and small normal faults (Whishaw, 1954). The orebodies comprise sphalerite and galena with pyrite and some pyrrhotite, the iron sulphides being a little less abundant than the ore minerals. They form irregular pods (Fig. IX-5) the position of which is con-

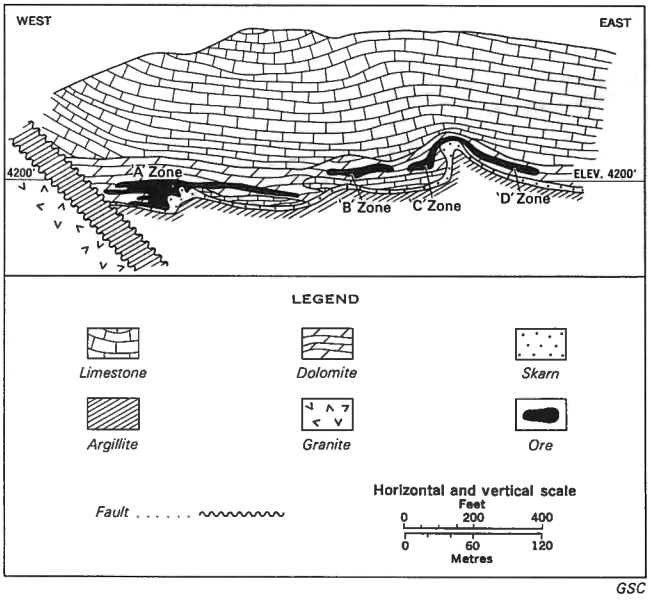


FIGURE IX-5. Diagrammatic cross-section of Jersey lead–zinc deposit, Salmo, British Columbia (after Q. G. Whishaw).

trolled by subsidiary coaxial folds on the limbs of the recumbent folds, and occur mainly in the anticlinal crests of the subsidiary folds and only on the upright limb of the fold. Trackless mining has been used successfully because of the generally flat lying nature of the orebodies.

Other deposits of this type are the H.B. (83), Reeves–MacDonald (79), and Duncan Lake, all occurring along Kootenay Arc in dolomitized limestone of Lower Cambrian age.

Shuswap subtype. The Big Ledge, Wigwam, Cottonbelt, Ruddock Creek, and Mount Copeland deposits all occur in mainly calcareous layers in the schist, marble, and paragneiss sequences of the Shuswap Complex that mantle complexly folded domes with cores of granite gneiss and have undergone high grade regional metamorphism. The age of the host rock is generally unknown but may be Lower Cambrian in part. All deposits are large, low grade, and as yet not exploited.

Mount Copeland (Fig. IX-2) deposit was formerly known as River Jordan (Riley, 1961). Two bedded lodes comprising galena, sphalerite, pyrite, pyrrhotite, and minor chalcopyrite, with quartz, calcite, and some barite, occur on both limbs of a rudely conical synform that is overturned to the southwest and plunges southeast. The lodes are thought to occur in the same bed of drag-folded and brecciated impure limestone. They are estimated to contain nearly 3 million tons of ore grading more than 5 per cent each of lead and zinc and 1 ounce silver per ton.

Transgressive Deposits

Transgressive deposits are scattered throughout the Cordillera, ranging in type from replacements in or near shear and fracture zones to veins with very little wall-rock replacement. These relationships are largely dependent on the susceptibility to replacement of the host rock. Transgressive deposits commonly contain copper minerals or sulphosalts and in general have a higher silver content than concordant deposits.

The *Bluebell* mine (91) on Kootenay Lake is a replacement of limestone in isoclinally folded schists (Irvine, 1957). Three ore zones, each consisting of clusters of orebodies, are spaced at 1,500-foot intervals along the strike of Lower Cambrian limestone. The orebodies occur along narrow cross-fractures in the limestone bed.

Silver

Most silver produced from the Cordilleran Region is obtained as a byproduct from lead–zinc ores, the Sullivan mine being the greatest source (Table IX-2). From several mines, however, the primary product is silver, among them United Keno (7), Dolly Varden (12a), Toric (12b), Silver Standard (18), Duthie (21), Hunter V and Double Standard (85), Number One (90c), Molly Gibson (89), Silver King (87c), and several mines in the Slocan camp. Most produce lead and zinc and a few, gold. The Silver King is unique in that it yielded only copper and silver, the latter occurring largely as the mineral stromeyerite. Silbak Premier mine (10b) was also a noteworthy producer of silver, but the complex ore also yielded much lead, zinc, copper, cadmium, and gold, only the gold being greater in value than the silver.

In the *Keno Hill–Galena Hill* area of the Yukon, United Keno (7) is the largest silver producing mine in Canada. The area (Fig. IX-6) is underlain by metamorphosed sedimentary rocks that have been intruded by pods of basic rock, now altered to greenstone (McTaggart, 1960; Boyle, 1965). The strata trend easterly and dip gently to moderately southward, the apparent succession consisting of "Lower Schist", "Central Quartzite", and "Upper Schist" units. On Keno Hill, the "Lower Schist" unit contains a band of quartzite, the "No. 9", lithologically similar to the "Central Quartzite." Recent work indicates that the "Lower Schist" and "Central Quartzite" are of Mesozoic age whereas the age of the "Upper Schist" unit and its relationship to the other two remain questionable.

Because rock exposures are few, the structure in the Keno and Galena Hills area is not fully understood.

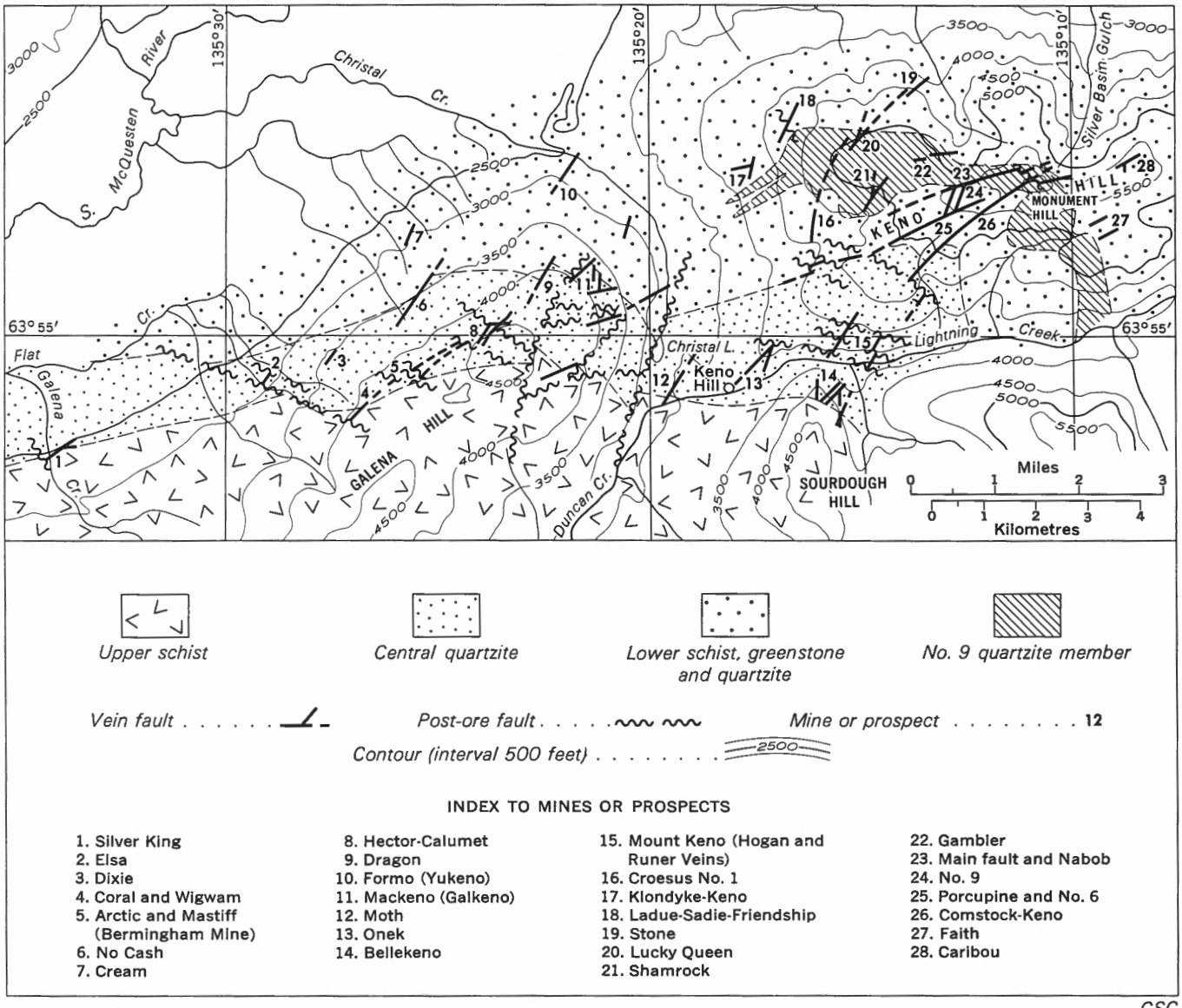


FIGURE IX-6. Geology of Keno Hill–Galena Hill, Yukon Territory, showing location of silver mines and prospects (after R. W. Boyle).

McTaggart regarded it as highly complex with an early stage of overthrusting to the north, producing bedding faults and tight, recumbent folds, followed later by northwesterly open folding, then vein-faults, and still later by low-angle bedding and cross-faults. Most vein-faults strike north to N70°E and dip steeply east to southeast. Boyle concluded that the ore shoots are localized in brecciated zones of early quartz, which was precipitated in the vein-faults with pyrite, arsenopyrite, and minor sulphosalts prior to galena, freibergite, and associated sulphides and carbonates, where the vein-faults traverse competent quartzite members of the "Lower Schist" and "Central Quartzite" or the greenstone bodies. Later cross-faults, which strike northwesterly and dip about 60°SW displace the vein-faults horizontally from a few feet to 500 feet. Primary ore minerals are galena, sphalerite, pyrite, and freibergite in a gangue of quartz and siderite. Supergene enrichment of lead and silver occurs in the upper part of the veins.

Gold

Gold occurrences are scattered throughout the Western Province. The larger producers, however, are more restricted; those that have produced more than 18,000 ounces are shown in Figure IX-7. Rising costs have forced many mines to close as the price of gold has remained fixed. In recent years only two major gold mines have been operating, the Cariboo Gold Quartz (32), which ceased operations in 1967, and the Bralorne–Pioneer (36). These mines yielded 53 per cent of the 1966 Cordilleran lode gold production, the balance being derived as a byproduct mainly of copper mines (Table IX-2).



INDEX TO NUMBERED DEPOSITS

2. Polaris-Taku
3. Tulsequah Chief
4. Big Bull
9. Big Missouri
10b. Silbak Premier
15c. Hidden Creek
27. Surf Inlet
31. Aurum (Island Mountain)
32. Cariboo Gold Quartz
34. Minto
36. Bralorne-Pioneer
43a. Mount Zeballos
43b. Privateer
44a. Central Zeballos
44b. Spud Valley
48b. Marble Bay
54. Tyee
56. Britannia
59. Vidette
60. Windpass
65. Copper Mountain
66a. Hedley Mascot
66b. Nickel Plate
67. French
72. Cariboo Amelia (McKinney)
73. Mother Lode

PHOENIX COPPER

74b. Brooklyn-Stemwinder
74f. Knob Hill-Ironsides
74g. Rawhide
74h. Snowshoe
— Idaho

74c. Dentonia
76. Union
77. Velvet

78a. CENTRE STAR GROUP

Centre Star
Le Roi
War Eagle
Josie

— I.X.L.
81. Bayonne
82a. Gold Belt
82b. Kootenay Belle
82c. Reno
82d. Sheep Creek
84a. Arlington
84b. Second Relief
86a. Centre Star (Wesko)
86b. Yankee Girl
86c. Ymir
87a. Athabasca
87b. Granite Poorman

Deposit . . . ○

Notes: Numbers are those assigned to deposits shown on Map 1252A, other deposits are named on figure. Properties from which gold is the principal product are underlined.

GSC

FIGURE IX-7. Lode gold deposits of the Cordilleran Region (by H. W. Little).

Nearly all deposits are mesothermal quartz veins with carbonate. Metallic minerals consist of native gold, tellurides, pyrite, arsenopyrite, pyrrhotite, sphalerite, and galena. Gold is often associated with the galena, particularly in high grade ores. Host rocks are commonly quartzites or granitic rocks, perhaps more susceptible to veining because of their competency.

The *Bralorne–Pioneer* mine (36), the most productive gold mine in the Canadian Cordillera, was formed in 1959 by the amalgamation of the Bralorne and Pioneer mines. The geology has been described by C. E. Cairnes, Joubin (1948), and Poole (1955). Impure cherts of the Fergusson Group, possibly of Permian age, are overlain successively by argillaceous sediments of the Noel, the Pioneer greenstone, and argillaceous beds of the Hurley Formations, all possibly of Upper Triassic age. These rocks trend west-northwesterly and were apparently folded into two anticlines separated by the Cadwallader shear zone. The southern limit of the northern anticline has been injected by an ultrabasic to acidic intrusive complex, the intrusions being less abundant south of the shear zone.

The veins occur between the Cadwallader shear zone and the Fergusson reverse fault to the north (Fig. IX-8). Two vein systems have been identified, the King, which occurs in the King section of the mine, and the Pioneer, which occurs in the Crown, Empire, and Pioneer sections. All veins are of fractured quartz and some ribboned, within which are disseminated native gold, gold telluride, several sulphides, and scheelite. The most westerly veins of the mine comprise the King vein system, from which the early production was obtained. Veins of the King system trend westerly, dip north, and are offset dextrally more than 200 feet by each of two steep north-trending faults, in the more westerly of which a mineralized vein occurs. The veins and ore shoots occur within and near the western end of an augite diorite body.

In the Pioneer vein system the ore-bearing quartz veins occur mainly in greenstone, according to Joubin, but Poole regards diorite as the most important host-rock. Ore shoots occur in all rock types except serpentine, but the richest ore occurs near serpentine, perhaps due to damming of the mineralizing solutions. The longest and most productive veins strike west to west-northwest and dip from 65°N to vertical. Two are several thousand feet long, and branches up to several hundred feet long are also ore bearing. There is only one major fault, which is between the Crown and Empire sections of the mine, and that strikes N20°W and dips 54°W. It is an oblique slip fault; the dip-slip is 370 feet, the hanging-wall having moved up, and the strike slip 240 feet. Displacement of the veins is also dextral.

Cariboo Gold Quartz mine (32) as now constituted comprises the old Cariboo Gold Quartz property, which has been inactive since 1959, and the nearby Aurum, formerly Island Mountain, mine (31), which was acquired in 1954 and from which all the ore supplying the concentrator has been derived in recent years (Guiguet, 1961). Impure quartzite, phyllite, slate, argillite, and minor grey limestone of the Midas and Snowshoe Formations of the Cariboo Group, the lower part of which is of Lower Cambrian age, form the northeast limb of an anticlinorium that trends northwest and is overturned to the southwest (Sutherland Brown, 1957). Most faults trend northerly and dip steeply east, the hanging-wall side having moved

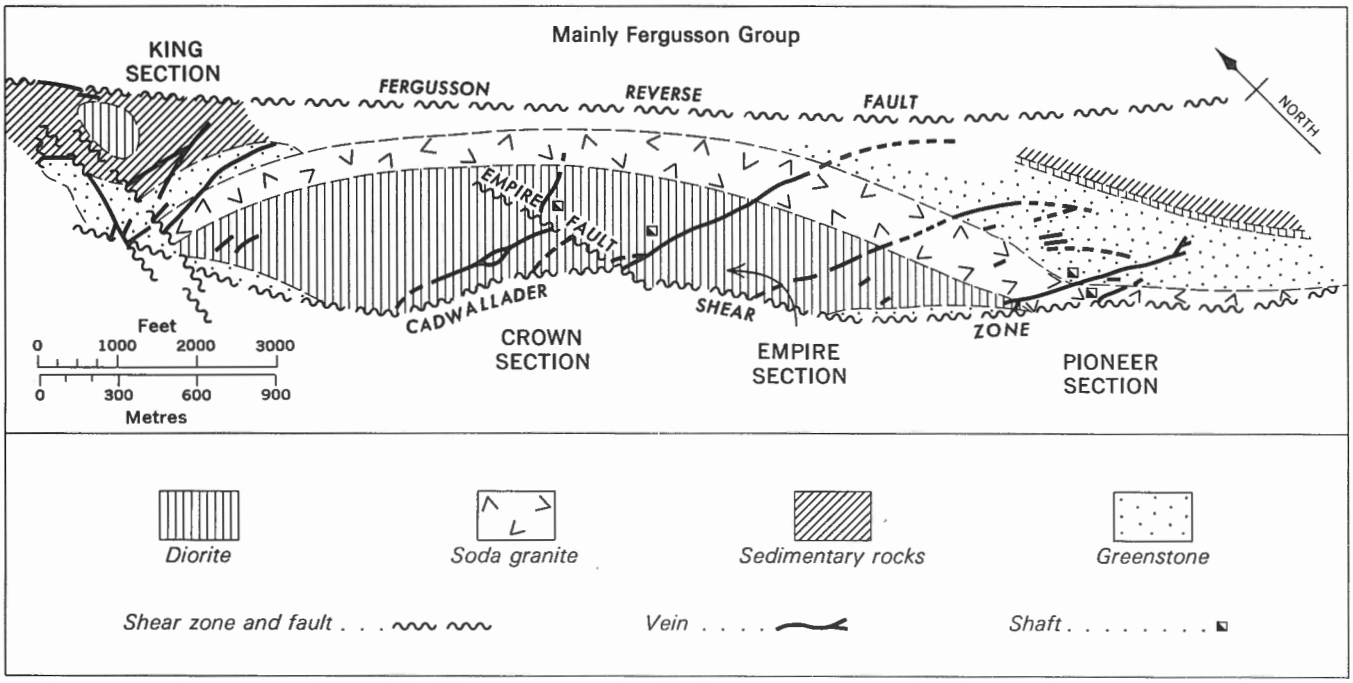


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FIGURE IX-8. Bralorne–Pioneer gold mine, British Columbia; plan at 3,200-foot elevation (modified from F. R. Joubin).

southward and downward. The largest fault has horizontal and dip-slip components of about 1,400 feet each.

Four types of veins are recognized: transverse, diagonal, northerly, and strike veins. Transverse veins occupy *ac* joints whereas the diagonal and northerly veins occupy conjugate faults that also were developed during folding. The strike veins are the longest, but are commonly barren or sparsely mineralized. They occupy faults apparently unrelated to the foregoing structural pattern. Only one strike vein has been of mineable grade. The transverse veins strike N30° to 55°E and dip mainly 70 to 90°SE. They are small but so numerous that they provided most of the ore from the Cariboo Gold Quartz section. The diagonal veins, which strike N70°E to east and dip steeply south, are up to 5 feet wide and average about 150 feet long. Many tributary branches are of the transverse type. The northerly veins strike N20°E to north and dip 50 to 70°E.

In recent years, however, most ore has come from replacement bodies in limestone beds. On the limbs of folds they are tabular bodies parallel to the bedding, but on the crests of the folds they are cigar-shaped, the long axes being parallel to the regional plunge, 20°NW, of folds on the limb of the anticlinorium. Metallic minerals consist of pyrite and associated free gold with minor galena, sphalerite, cosalite, bismuthinite, scheelite, pyrrhotite, arsenopyrite, and chalcopyrite. Gangue minerals are quartz, ankerite, and muscovite in the veins, and ankerite with some quartz in the replacement bodies.

Hedley Camp, one of the larger gold camps on the Cordillera, contains the *Hedley Mascot* (66a), and *Nickel Plate* (66b) mines. These are unusual in that the ore occurs in skarn near the contact with unaltered Upper Triassic limestone. Early workers believed the ore to be related to the gabbro sills which produced the skarn, but Billingsley and Hume (1941) showed that the ore was spacially related to a single sheet, the "Midway" dyke, and to small folds subsidiary to the major ones, to zones of brecciation in skarn near the "Midway" dyke, and to transverse dykes which effected a damming of mineralizing solutions. Because the transverse dykes were emplaced after the formation of skarn but before mineralization, the ore is not derived from the "Midway" dyke but is related to a later period of mineralization, possibly from nearby granodiorite intrusions.

Copper

Copper deposits are found in four principal regions —south–central, southwestern, and northwestern British Columbia, and southwestern Yukon (Fig. IX-9). They are closely related to basic volcanic rocks, particularly those of Triassic age where invaded by granitic rocks. Copper ranks fourth in value of metal produced in the Western Mineral Province (Table IX-1); more than 3½ billion pounds valued at more than $600 million has been produced to date.

The largest copper deposits are replacement bodies which are subdivided according to the type of host rock as contact metamorphic, shear zone replacement, and porphyry deposits, the last being the replacement or fracture filling of brecciated intrusive rocks. Vein deposits of copper, such as those at Rossland, commonly show replacement of wall-rock adjacent to the veins, but at Rossland, although important amounts of copper were produced, the value of gold was greater.

Contact metamorphic deposits. These are represented by Pueblo (9), Big Chief–Little Chief (10), Craigmont (57), Phoenix (74 b, f, g, h), Mother Lode (73a), Sunset (73b), Marble Bay (48b).

The *Phoenix* deposits are the most productive of the contact metamorphic type. Chert and greenstone comprising the Knob Hill Formation of Carboniferous or Permian age is overlain unconformably by chert sharpstone 'conglomerate', a peculiar mainly bedded rock composed largely of chert fragments that was formerly regarded as jasperoid, but later shown by Seraphim (1956) to be a sedimentary breccia. The sharpstone conglomerate is overlain by Middle Triassic Brooklyn limestone and an upper sharpstone conglomerate which may be younger or the same repeated by folding or low-angle faulting. The strata are warped into a basin-like structure (Fig. IX-10) much of the eastern part of which is covered by Lower Tertiary sedimentary and volcanic rocks dipping east and terminated by a normal fault.

The limestone is converted to skarn generally near the base, but sometimes well above it, although no large intrusive bodies are known nearby. The closest is a Cretaceous granodiorite more than 2 miles distant. Around the periphery the structure seems simple with moderate to steep inward dips, but in the central part the limestone beds dip vertically and are apparently tightly folded being cut off on the north by a south-dipping reverse fault. Most orebodies lie within the skarn around the southern and western parts of the basin. The Stemwinder, however, occurs in a lens of limestone breccia or conglomerate that may form the core of the synclinal basin and may be Norian.

The largest orebodies are exposed in the Knob Hill–Ironsides pit (*Frontispiece*). The skarn, together with interbedded green argillite, lenses of sharpstone conglomerate, and a little quartzite, has been silicified and impregnated with chalcopyrite, pyrite, specular hematite, and some magnetite and pyrrhotite. This forms the 'lower ore band' which trends northerly and dips gently eastward. Above, also trending northerly but dipping more steeply eastward, is the 'upper ore band' which lies below the upper sharpstone conglomerate. Host rock for this orebody is mainly marble, with minor sharpstone conglomerate, skarn, and green argillite.

Shear zone replacement deposits. These are massive sulphide deposits and include the Britannia (56), Sunro (55), Hidden Creek (15c), Granduc, Tyee (54) mines and others.



INDEX TO NUMBERED DEPOSITS

9. Pueblo
10. Big Chief- Little Chief
3. Tulsequah Chief
4. Big Bull
10b. Silbak Premier
13. Outsider
14 Bonanza
15c. Hidden Creek
19. Rocher Déboulé
22. Granisle
27. Surf Inlet
48a. Cornell
48b. Marble Bay
48c. Vananda (Little Billie)
51. Indian Chief
53. Blue Grouse
54. Tyee
55. Sunro
56. Britannia
57. Craigmont
58. Bethlehem
61. Iron Mask
64. Pride of Emory
65. Copper Mountain
66a. Hedley Mascot
66b. Nickel Plate
73a. Mother Lode
73b. Sunset
74a. B.C.
PHOENIX COPPER
74b. Brooklyn-Stemwinder
74f. Knob Hill-Ironsides
74g. Rawhide
74h. Snowshoe
Idaho
74d. Emma
74e. Oro Denero
77. Velvet
78a. CENTRE STAR GROUP
Centre Star
Le Roi
War Eagle
Josie
87c. Silver King
88. Queen Victoria

Pyrometasomatic . . . ○ Replacement . . . ● 'Porphyry' and stockwork . . . □ Vein . . . ◊ Unclassified . . . ×

Notes: Numbers are those assigned to deposits shown on Map 1252A, other deposits are named on figure.
Properties from which copper is the principal product are underlined.

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FIGURE IX-9. Copper deposits of the Cordilleran Region, classified according to type (by H. W. Little).

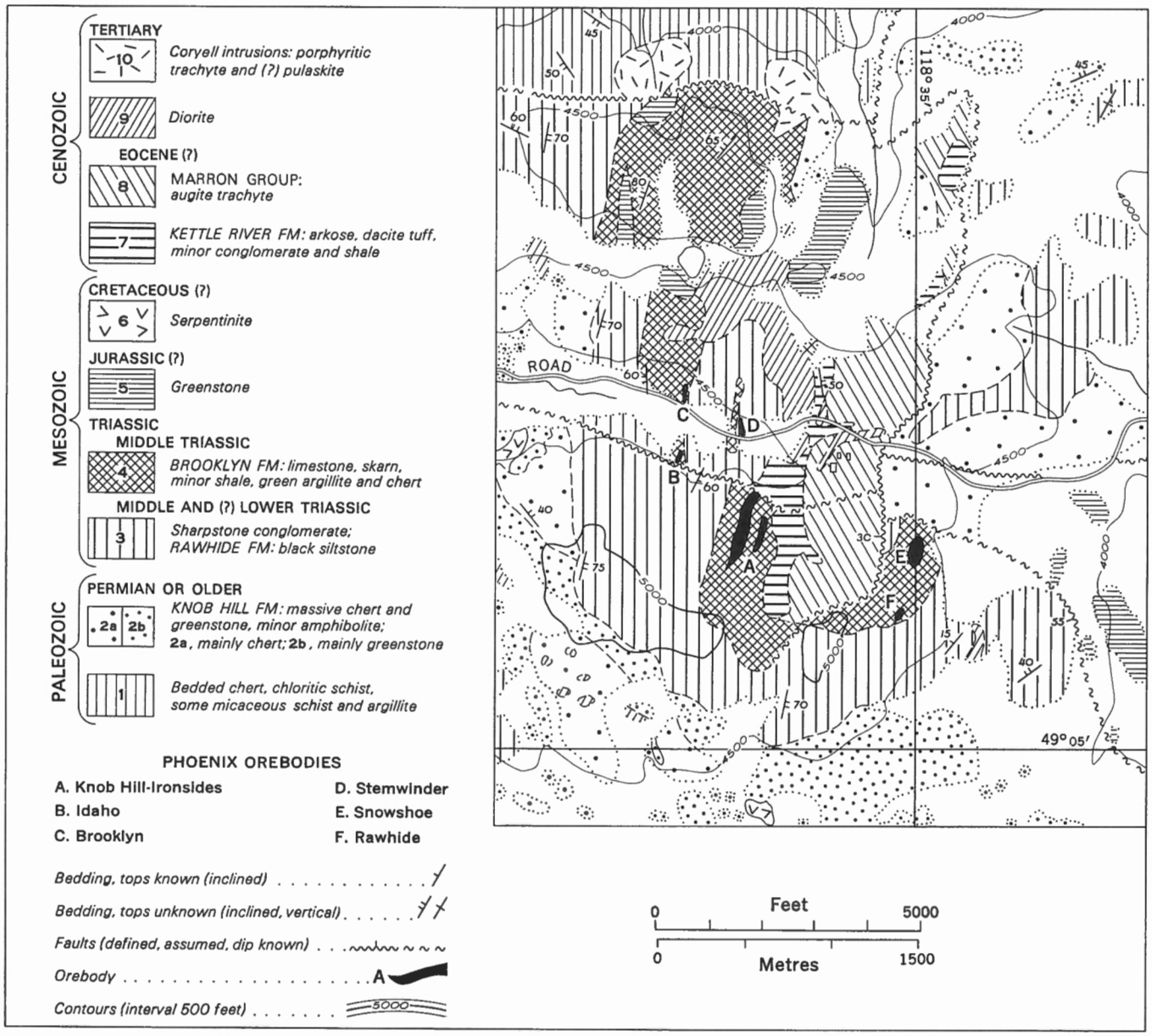


FIGURE IX-10. Geology in the vicinity of Phoenix copper deposits, British Columbia (by H. W. Little).

GSC

Britannia, the largest copper deposit mined to date in the Canadian Cordillera, occurs in a large roof pendant in the Coast batholith 20 miles north of Vancouver. The host rocks are steeply dipping quartzites, slates, cherts, and green tuffaceous sediments and volcanic clastics of the Britannia Formation, the lower part of the Gambier Group, of probable Upper Jurassic and Lower Cretaceous age. The rocks are intensely sheared along a zone 2,000 feet wide striking northwest and dipping steeply southwest into fissile sericite schist and, locally, talc or chlorite. The orebodies occur in the hanging-wall of a thrust fault near the footwall of the major shear zone (Irvine, 1948). Eight orebodies are known, which occur at intervals for 1½ miles of the 5-mile-long shear zone. They comprise huge lenticular to sheet-like masses formed of quartz, pyrite, and, locally, sphalerite, that were subsequently brecciated and then mineralized by chalcopyrite. Mining is by underground methods.

Porphyry copper deposits. Copper Mountain (65), Bethlehem (58), Catface, Granisle (22), and Galore Creek.

The *Bethlehem* mine (58) has recently come into production (White, Thompson, and McTaggart, 1957; Carr, 1960). The deposit occurs within the Guichon quartz diorite batholith of Early Jurassic age. The older quartz diorite on the Bethlehem property is intruded by a complex of granite, quartz diorite, dacite porphyry, and

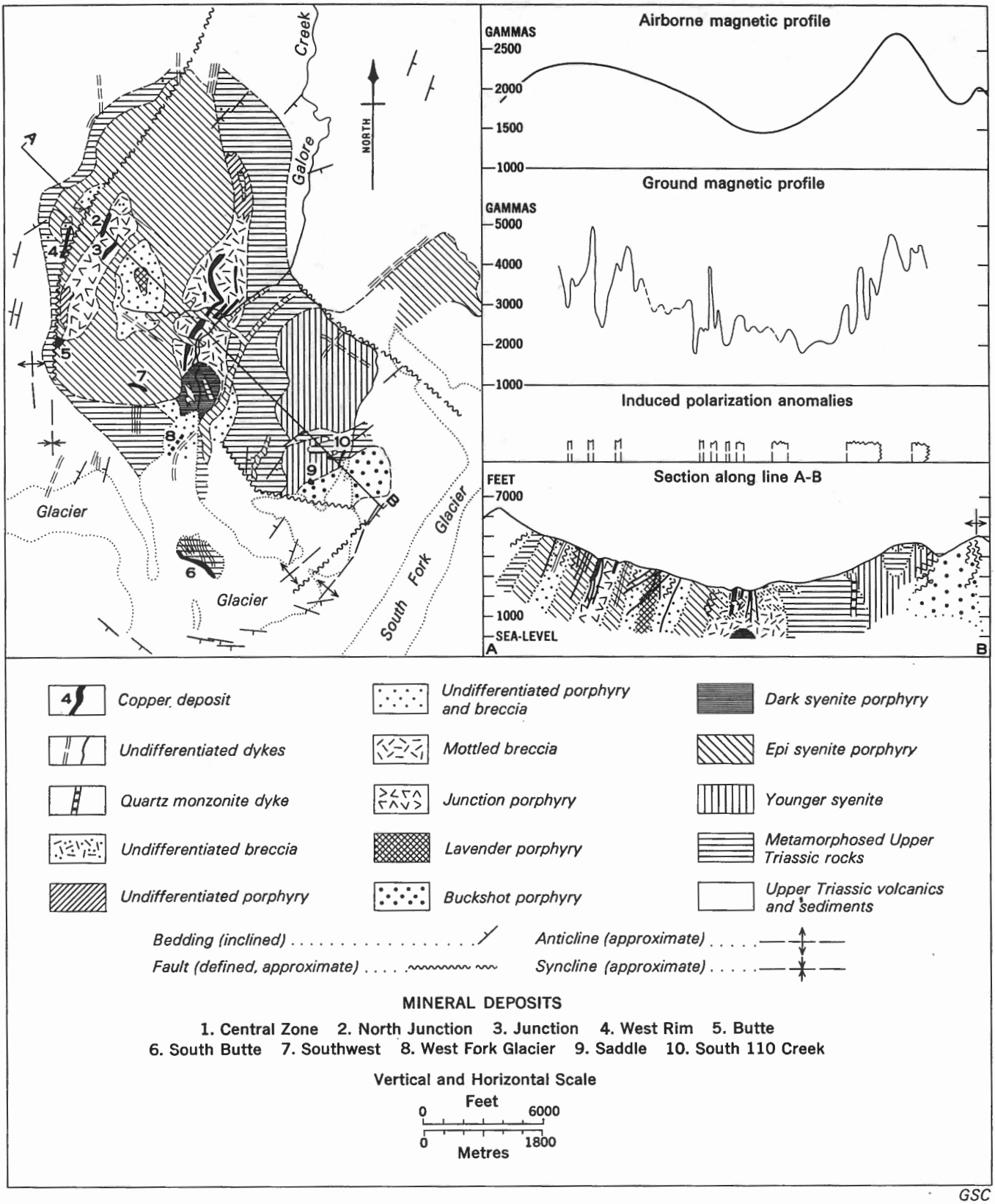


FIGURE IX-11. Geology and cross-section of Galore Creek copper deposits, British Columbia (after D. A. Barr).

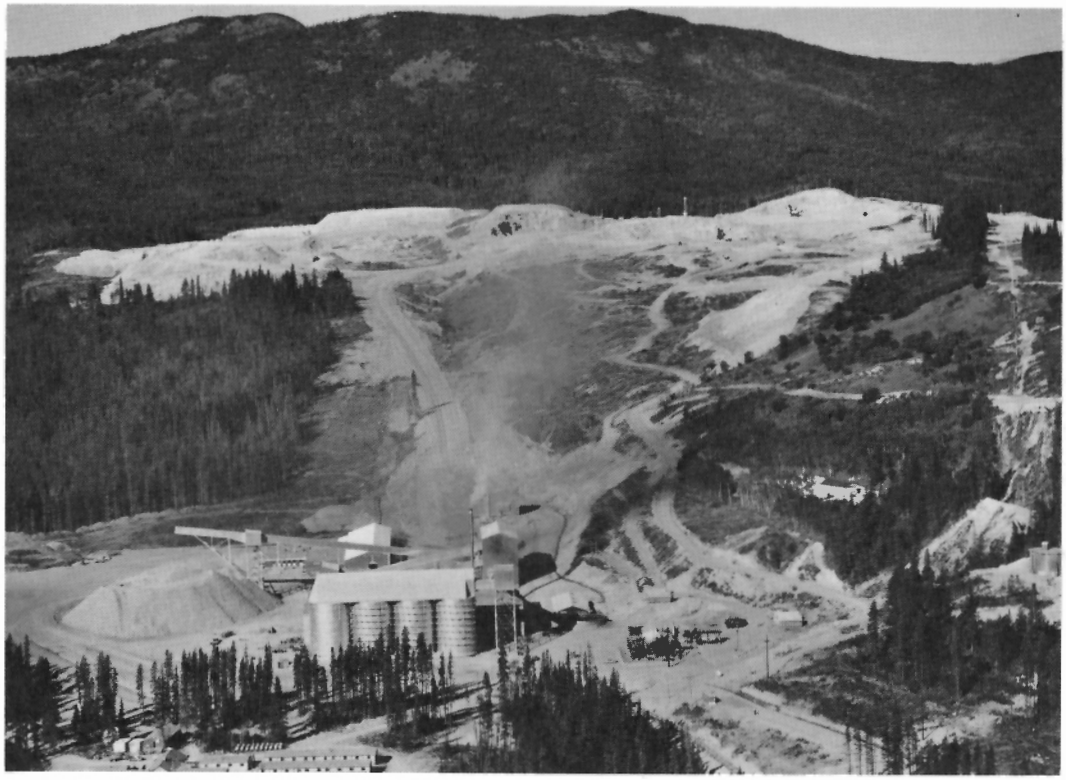


PLATE IX-1

Open pit and mill of Endako molybdenum mine, British Columbia.

quartz diorite porphyry. Explosion breccia is associated with the porphyry bodies. The breccia, host to the ore, contains fragments mainly of porphyry in abundant fine matrix. The breccia is silicified, sericitized, and has undergone argillic alteration. It also contains tourmaline, amphiboles, biotite, albite, epidote, chlorite, calcite, and zeolites. Three principal bodies of breccia, known as the Iona, Jersey, and East Jersey, are mineralized, mainly in their more altered parts, by chalcopyrite, bornite, and minor pyrite and molybdenite. Initial mining is by open pit.

Galore Creek copper deposits (Fig. IX-9) lie within a mountainous area 2 by 3½ miles east of Stikine River on the south flank of Stikine Arch (Barr, 1966), and though characteristically porphyry copper type they are in part pyrometasomatic. The area is underlain by volcanic and sedimentary rocks of probable Upper Triassic age which are rudely domed about a complex body of epidotized syenite porphyry (Fig. IX-11). Within the complex are zones of cataclastic breccia that occur mainly along facies contacts and which are hydrothermally altered and mineralized with chalcopyrite, bornite, pyrite, and magnetite in a gangue of biotite, potash feldspar, gypsum, garnet, chlorite, and carbonate. Secondary minerals, such as chalcocite, cuprite, native copper, malachite, and azurite occur as much as 100 feet from the surface.

Ten deposits are known of which the largest, the 'Central Zone', is 6,500 feet long and trends north-northeast. The copper minerals are disseminated throughout, but several sections of higher grade material are formed by swarms of veinlets and replacement masses.

Molybdenum

To the end of 1966 production of molybdenum in the Western Mineral Province was 24,655 pounds, valued at $40,570, and derived from a few small, high grade deposits. Recent exploration has established large deposits of low grade ore on the Endako (28), Boss Mountain (33), Red Mountain (78b), Lime Creek (16), and Glacier Gulch properties. The first three were in production in 1966; the others were under development.

The distribution of molybdenum deposits is similar to that of copper deposits; indeed they are commonly associated. Vokes (1963) has shown that molybdenum deposits are spacially and genetically related to the Nelson Intrusions in southeastern British Columbia, and to the Guichon batholith and Topley Intrusions in central British Columbia. The Glacier Gulch deposit on Hudson Bay Mountain and the Lime Creek deposit on Alice Arm appear to be related to leucocratic Tertiary acidic intrusions.

The *Endako* (28), in central British Columbia near Fort Fraser, is the largest molybdenum mine in Canada. Mining by open pit (Pl. IX-1) was begun in 1965, and 22,000 tons is now milled per day. Ore reserves are

stated to exceed 90 million tons grading 0.20 per cent MoS_2.

The region is underlain by a batholith of the Topley Intrusions, considered to be Lower Jurassic although some parts may be younger. South of the mine, on the shore of François Lake, altered lavas probably of the Takla Group underlie a small area, but contacts with Topley Intrusions are not exposed. A few miles to the west, the batholith is overlain unconformably by the late Tertiary Ootsa Lake Group.

Carr (1966) established fourteen distinct facies, nine of which he referred to as quartz monzonite, one as granodiorite, and four as quartz diorite. Because bedrock exposures are poor the relationships of only a few units have been established—the others have to be deduced by indirect evidence. The orebody lies within the Endako quartz monzonite, an early phase of the Topley Intrusions. It is pinkish grey, medium grained with biotite and minor hornblende and contains numerous small phenocrysts of red orthoclase and a few of plagioclase and quartz and, in many places, clusters of more basic biotite-rich inclusions. To the north lies the generally finer grained leucocratic Casey biotite quartz monzonite, an elongate, sigmoid body oriented mainly in an east–west direction. The relative age of the two intrusions has not been established.

At the surface the orebody is more than 5,000 feet long and 1,000 feet wide, trends west-northwesterly, and appears to plunge gently west. The orebody is more than 500 feet deep in the central part and grades outward into sub-ore. Along a small part of its southern boundary, however, the orebody terminates against an east-striking fault that dips steeply north. The host rock has undergone, to various intensity, silicification, sericitization, and potash metasomatism. Ilmenite, leucoxene, and sphene are unusually abundant. The altered quartz monzonite has been intruded by numerous pre-ore dykes up to 50 feet wide of aplite and feldspar porphyry, and narrower post-ore dykes of lamprophyre. All rocks except the lamprophyre are traversed by predominantly gently dipping sinuous quartz veins from a fraction of an inch to 5 feet wide and as much as 400 feet or more along strike and down dip. The larger veins, however, in the western part of the orebody more commonly trend east-northeast and dip south. A few major faults trend east, northeast, and northwest, and numerous small lineaments trend north. The ages of these faults have not been determined, but as no mineralized quartz veins have been seen in them, they are probably post-ore. However, some molybdenite does occur in the gouge.

Molybdenite occurs mainly in the quartz veins and in fractures in aplite; much of it is fine grained, but coarse flakes do occur. According to C. W. Ball there appear to be three stages of mineralization. The first consists of molybdenite in a few early quartz veins and in adjacent altered wall-rock. The main mineralization is fine-grained molybdenite along the margins of narrow quartz stringers. It is accompanied by pyrite more extensive than the molybdenite orebody. The pyrite comprises less than 1 per cent of the orebody and occurs both in the quartz veins and disseminated in the wall-rocks. The final mineralization is coarse-grained molybdenite associated with calcite. Minor magnetite and specularite are present; chalcopyrite is rare.

Iron

The productive iron deposits of the Western Mineral Province are entirely on the coastal islands of British Columbia. They consist of pyrometasomatic magnetite deposits adjacent to diorite or granodiorite intrusions formed in skarn at the base of Upper Triassic limestone overlying basaltic pillow lavas. The Tasu (25) and Jedway (26) mines are on Moresby Island, the Empire (40), Nimpkish (41), Zeballos Iron (42), Iron Mike (37), Brynnor (52), and Argonaut (46) are on Vancouver Island, and the Texada (49) is on Texada Island. The Tasu and Texada deposits also contain considerable copper ore. It is conjectured (Eastwood, 1965) that the basalts, which contain as much as 14 per cent iron, may be the source of iron in these deposits since chemical conditions for an ion exchange from silicate to oxide are favourable at the contact with limestone. About 2 million tons per year of iron ore concentrate is produced and exported to Japan.

About 100,000 tons of pig iron per year is obtained at Trail and Kimberley from iron sulphide tailings produced in the milling of lead–zinc ores.

Tungsten

Tungsten deposits of the Western Mineral Province are somewhat scattered but appear to form three distinct provinces. The first, a beryllium–tin–tungsten–molybdenum province, extends along Omineca geanticline, mainly on the west side (Mulligan, 1966); the second is within the multi-metallogenic province centred about Hazelton; the third is related to granitic intrusions that form an east–west belt centred at Keno Hill, Yukon (Bostock *in* Little, 1959). Two types of lode deposits are of economic importance: quartz fissure veins and contact metamorphic.

The quartz fissure veins commonly carry pyrrhotite, chalcopyrite, and molybdenite, with wolframite or ferberite more common than scheelite. The *Red Rose* (20) is in a tungsten-bearing vein cutting a diorite body which is probably a cupola of the granodiorite stock forming the core of Rocher Déboulé Mountain (Kindle *in* Little, 1959), but at depth the vein follows the contact between the diorite and metamorphosed Lower Cretaceous argillite, quartzite, and greywacke (Sutherland Brown, 1960). The

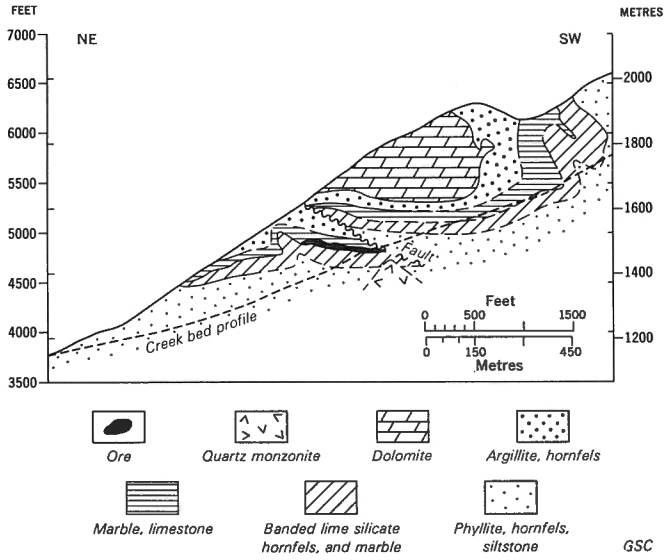


FIGURE IX-12. Cross-section through Canada Tungsten mine, Flat River, District of Mackenzie (by S. L. Blusson).

vein strikes N30°to 45°W and dips 40° to 75°SW. Within it, the two ore shoots rake steeply northwest, and consist of quartz with feldspar, biotite, hornblende, ankerite, tourmaline, apatite, scheelite, ferberite, molybdenite, and chalcopyrite. The Red Rose produced tungsten concentrate during World War II.

Contact metamorphic deposits occur as irregular replacement bodies in pyroxene–garnet skarn and to a lesser degree in adjacent limestone and granite. Such bodies are usually cut by tungsten-bearing quartz veins. Scheelite is the most abundant ore mineral and is usually accompanied by abundant pyrrhotite. Former producers are the Emerald (80a), Dodger, and Feeney mines near Salmo, British Columbia. In 1965 at the *Canada Tungsten* mine (1) at Flat River, District of Mackenzie, 108,000 tons of ore averaging 2.53 per cent WO_3 was milled. Ore reserves, prior to mining, were reported to be 1,320,000 tons grading 2.51 per cent WO_3 and 0.45 per cent copper (Green and Godwin, 1963). The scheelite (S. L. Blusson) is disseminated in a dark green pyroxene and garnet skarn forming a tabular replacement body at the base of a massive Lower Cambrian limestone (Fig. IX-12). The skarn occurs only on the lower, flat-lying limb of an overturned syncline which is intruded by quartz monzonite and cut by pre-ore faults. These faults and an intrusive cupola about 100 feet below the deposit are thought to localize the ore within the limestone host rock. The ore consists of scheelite, pyrrhotite, and minor chalcopyrite, enriched locally by coarse-grained scheelite–microcline–quartz veins and later veinlets of fine-grained biotite, quartz, scheelite, and sulphide.

Mercury

Mercury deposits of the Western Mineral Province occur in two subprovinces (Stevenson, 1940). The more southerly extends from Kamloops westward to Bridge River and the other is along Pinchi fault. An isolated occurrence is known at Sechart, Vancouver Island. In both subprovinces mercury occurs as cinnabar replacing carbonate rocks adjacent to faults. The carbonate rocks are either dolomitized limestone or carbonated serpentinite. A few minor deposits occur in altered volcanic or sedimentary rocks. The host rocks range in age from Permian to early Tertiary.

Of the two deposits along the Pinchi fault, the *Pinchi Lake* mine (23) was the main source of mercury for the allies during World War II, when 4.2 million pounds of mercury valued at more than $10 million was recovered at the two properties. Cinnabar, minor pyrite, and minor stibnite occur as veinlets and blebs in dolomite of Permian age, although one orebody is in carbonatized brecciated serpentine. Quartz and calcite are gangue introduced with the cinnabar into cracks and cavities near the Pinchi fault zone.

Nickel

The *Pride of Emory* mine (64) is to the present the only source of nickel ore in the Cordilleran Region. Although discovered in 1923 it did not reach major production until 1958. The host rocks comprise a complex of peridotite, pyroxenite, and hornblendite which, except for a zone of schist on the south, is surrounded by diorite and minor norite of the Coast Intrusions of post-Paleozoic, pre-Eocene age. Twelve orebodies are known (James and Eastwood, 1964) ranging from 50 to 150 feet in diameter and from 250 to 1,000 feet long. They plunge steeply north or west, most within or near a plane that strikes N75°W. It is uncertain why they are so localized. The ore shoots are disseminated and massive sulphides consisting mainly of nickeliferous pyrrhotite, pentlandite, and chalcopyrite, with small amounts of other sulphides of iron, nickel, cobalt, and copper. It has not been clearly established whether magmatic segregation or hydrothermal action was the primary process by which the ore was emplaced.

Minor Metals

Other metals produced in the Western Mineral Province are cadmium, antimony, tin, and bismuth, each valued at more than $10 million, platinum at more than $100,000, and manganese, chromium, palladium, and cobalt valued at less than $100,000 each (Table IX-1). Uranium is reported in several deposits. Cadmium, antimony, tin, and bismuth are obtained as byproducts in lead–zinc mines. Antimony is also obtained from stibnite-bearing veins.

Sedimentary Deposits

Unlike the Eastern Mineral Province, there are no metallic deposits of sedimentary origin that are currently of economic value in the Western Mineral Province. Of interest are germanium-bearing coaly beds in sediments of Eocene age near Powell River and extensive magnetite-bearing Lower Cambrian quartzite up to 8 feet thick in Salmo area.

Surficial Deposits

Metallic minerals occurring in surficial deposits are placer and chemical precipitates. Of the latter the bog iron deposits are the most important, but none has yet been exploited. On a tributary of Zymoetz River east of Terrace more than half a million tons of ore assaying more than 50 per cent iron rests on a pyrite-bearing porphyritic rock. Similar deposits occur at the head of Taseko Creek northwest of Lillooet.

Gold

Gold placer mining superseded the fur trade as the important spur to industrial development of the Canadian Cordillera. This impetus occurred in the southern part in 1858 and in the Yukon fifty years later. As the placer fields became depleted the attention of prospectors and miners turned to the discovery and development of lode mines. In addition to gold and an appreciable amount of silver (average fineness 861 in British Columbia and about 700 in the Klondike), placer mining has yielded minor amounts of tin and tungsten minerals and some platinum metals. Early operations consisted almost entirely of working shallow deposits by hand, but as early as 1880 hydraulic methods were introduced. Later, bucket-line and suction dredges were used, and were in operation until 1966 in the Yukon where permafrost requires thawing of the gravels.

More than 5 million ounces of placer gold worth nearly $100 million has been produced in British Columbia, whereas Yukon placers have yielded nearly 11 million ounces valued in excess of $250 million. In the past decade nearly all placer gold in Canada was from operations in Klondike area, Yukon.

The alluvial deposits formed for the most part during the Tertiary. Deep decay of rocks under moderate to semitropical conditions occurred in the early Tertiary. Uplift in the late Tertiary led to concentration of gold in stream gravels along with such minerals as magnetite, pyrite, platinum, osmiridium, cassiterite, scheelite, rutile, and monazite. Although glaciation destroyed many deposits, those in deep, narrow canyons escaped scouring by ice. Western Yukon, however, was not glaciated partly due to the aridity of the climate during the Pleistocene, with the result that the Klondike district was one of the richest placer fields in the world.

The *Klondike* district (3) near Dawson was the source of more than half the placer gold of Western Canada. Bedrock is mainly the Klondike schist of sedimentary origin, containing albite, quartz, chlorite, and calcite. Bodies of sheared quartz-feldspar porphyry may also be present. Some rocks resemble those of late Proterozoic or Cambrian age that lie some distance to the northeast, but the age is not known.

Numerous quartz veins, mostly small, occur in the Klondike schist and contain pyrite, a little chalcopyrite, galena, and free gold. Most gold is concentrated by erosion in the pre-Pleistocene creek channels, known as the 'white channel.' Subsequent stream action has reworked these auriferous gravels to form new placers in present stream beds, at various levels as much as 300 feet below the 'white channel' paystreak (McConnell, 1905).

The *Cariboo* district (30) was at one time the most productive placer gold field in British Columbia. As records are incomplete for the years prior to 1874 total production is not known. It is estimated to be more than 2.5 million ounces, valued at more than $50 million. Williams and Lightning Creeks, the two most celebrated, are reported to have yielded gold to the value of some $20 million.

The region is one of moderate relief deeply dissected by stream erosion underlain by phyllite, quartzite, argillite, and limestone of the early Paleozoic Cariboo Group. Within the main productive belt these rocks form two anticlinoria and an intervening synclinorium striking northwest (Sutherland Brown, 1957). Decomposition by weathering of auriferous quartz veins during the late Cretaceous or early Tertiary probably caused concentration and enrichment of the gold by alternate solution and deposition (Johnston and Uglow, 1926). Uplift in the late Tertiary resulted in rejuvenation of streams and concentration of the gold in stream gravels. Subsequent glaciation apparently was limited in destructive effect because of stagnation of the ice in the narrow valleys so that some Tertiary placers were preserved. Elsewhere reconcentration of gold from the glacial till was accomplished by recent stream action.

Other placer fields, each of which has yielded $1 million to $40 million in gold are Atlin, Fraser River, Omineca, Dease Creek, and Tulameen in British Columbia, and Sixtymile River, Haggart, Livingstone, Burwash, Henderson, and Clear Creeks in Yukon.

Hydrothermal and Metamorphic Deposits

Asbestos

Many asbestos deposits are known in the Western Mineral Province, but of these only that at Cassiar is in production.

At *Clinton Creek,* near Dawson, Yukon, cross-fibre asbestos up to one quarter inch long occurs in a small ultrabasic body enclosed by phyllites, argillites, and quartzites. Production from this property commenced in late 1967. *Cassiar Asbestos* (5) is on Mount McDame about 3 miles north of Cassiar. It lies within a 500-by-1,700-foot ultrabasic body of Mississippian (?) age. The asbestos occurs as cross-fibre veins in dark green serpentine, commonly one half to one inch wide and up to 15 feet long. Gabrielse (1963) concluded that the asbestos was formed long after serpentinization occurred, during the deformation accompanying emplacement of Cassiar Batholith during the Columbian Orogeny under metamorphic conditions represented by the upper part of the greenschist facies.

Barite, Witherite, Fluorite

Barite is produced only in upper Columbia Valley south of Golden. That from the Silver Giant mine (100) near Spillimacheen is recovered from tailings whereas that from the Mineral King mine (104) near Invermere is a byproduct of lead–zinc mining. Barite veins are mined at Brisco (101), Parson (99), and Larrabee. The mineral also occurs as gangue in the Britannia copper mine (56) but none is recovered.

Witherite occurs near the lower crossing of Liard River by Alaska Highway in British Columbia and is reported near the crossing of Pelly River by Canol Road in the Yukon. Both occurrences are replacements of limestone cut by narrow veins.

The largest known deposit of fluorite in the Cordillera is the *Rock Candy* (75), a lenticular stockwork of veins composed of fluorite, barite, quartz, and calcite in early Tertiary Coryell syenite. Near Birch Island, the Rexspar property, in recent years also a uranium prospect, contains over a million tons of material containing 20 per cent fluorite and 15 per cent celestite.

Magnesite

Extensive beds of magnesite, possibly of sedimentary rather than hydrothermal origin, are known in eastern Purcell Mountains. In the Lower Cambrian Cranbrook Formation near Cranbrook, the magnesite bed is 25 to 55 feet thick, and intermittently exposed over 18,000 feet. It is interbedded with quartzite into which it grades through zones 75 to 150 feet thick of alternating thin quartzite and carbonate strata. A similar but smaller deposit occurs in the same formation several miles farther west. Recently magnesite has been discovered in the Proterozoic Mount Nelson Formation near Brisco. The deposits are pod-like bodies and apparently were formed by hydrothermal alteration of dolomite. The largest deposits, reported in 1966, occur in the Middle Cambrian Cathedral Formation on the west flank of Mount Brussilof, near the junction of Cross and Mitchell Rivers. None of the magnesite deposits of the Cordillera has been exploited.

Perlite

Perlite of early Tertiary age intercalated with rhyolite occurs at François Lake in central British Columbia and was first noted by Dawson in 1876. About 1,112 tons of the volcanic glass was mined and shipped in 1953, but there has been no additional production.

Talc

Talc is common in the serpentinites of the Western Mineral Province, but few deposits have been worked. Production was mostly from southern Vancouver Island, and small amounts from the vicinity of Lillooet.

Jade

Within the last 10 years the collection of jade, which occurs as alluvial boulders (some up to 10 tons), has developed into a substantial minor industry. Jade has been reported in lower Fraser Valley near Hope and from many places in the interior of British Columbia from the International Boundary to Wheaton Creek. The most productive region is in the Fraser Canyon north of Hope and along the tributary Bridge and Yalakom Rivers. The jade is nephrite variety. Although the host-rock of jade is serpentinite, little jade has been found in place.

Evaporites

Gypsum

Gypsum is used in the manufacture of wallboard, plaster, and portland cement. Only one deposit, at *Falkland* (62), has been worked in the Western Mineral Province, more than a million tons having been quarried. Pods of anhydrite and gypsum occur in argillaceous sediments of the Permian Cache Creek Group, closely associated with faults. The deposits have been regarded as hydrothermal, but recently Baird (1964) suggested that they are sedimentary evaporites squeezed upwards along faults and emplaced by plastic flow.

Devonian gypsum was quarried from 1945 to 1958 at Mayook (108) in Rocky Mountain Trench east of Cranbrook. Large deposits occur in the Eastern Mineral Province in the Middle Devonian Burnais Formation, currently exploited at two quarries near Windermere. Elsewhere in the Eastern Mineral Province gypsum has been reported at many places, the largest deposits being in Manitoba. It is mined at Amaranth (17) from a bed 40 feet thick in the Jurassic Amaranth Formation occurring 100 feet below the surface, and at Silver Plains (22) from a bed 30 to 50 feet thick lying at 150 feet depth. At Gypsumville (15), Manitoba, it occurs as isolated ridges arising above swamp and although situated within the outcrop belt of Silurian rocks, its age is not definitely known.

Other occurrences are known in southern Yukon in the Rainy Hollow area of northwestern British Columbia, and in the Triassic Whitehorse Formation at Mowitch Creek in Jasper Park. It occurs also on the banks of Slave and Salt Rivers north and west of Fort Fitzgerald.

Sodium and Magnesium Sulphate

Sodium sulphate (Glauber's salt) is found in many closed drainage basins of southern Saskatchewan in the form of brines and intermittent or permanent crystal beds. A large industry has been built for its recovery; reserves are estimated at 200 million tons, over 4 million tons having been produced. The sodium sulphate is mined from dried or frozen lake deposits but recently the bulk is obtained by evaporation of brine from the alkali lakes. About 95 per cent of production is used in the pulp and paper industries. Plants are in operation at Palo, Chaplin Lake, Bishopric, Ormiston, and Gladmar. Production in 1965 totalled 323,000 tons valued at more than $5 million.

Sodium carbonate, sodium sulphate, and magnesium sulphate (epsom salt) are obtained from several small basins in the semi-arid Interior Plateau of British Columbia where the salts are precipitated in undrained lakes during times of abundant evaporation or cold weather. The precipitates form bowl-like masses of crystals embedded in mud. Total production is about 14,000 tons of magnesium sulphate and more than 10,000 tons of sodium carbonate.

Salt

Salt (halite) occurs in both the Middle and Upper Devonian strata in Western Canada. It is produced commercially from the Prairie Formation of the Middle Devonian Elk Point Group. This salt horizon occurs on a west-dipping homocline that extends from northeastern Alberta to Manitoba at depths varying from 600 feet at McMurray to over 5,000 feet between Edmonton and southern Saskatchewan. Thin layers of anhydrite and clay are common in the salt sequence, and the pure salt grades laterally to anhydrite and clastics. At Lindbergh salt for domestic and industrial consumption is produced from brine from a section 700 feet thick at a depth of 3,600 feet, and from a section 400 feet thick at a about 3,700 feet depth near Unity, Saskatchewan. At Duvernay brine is used to make caustic soda, chlorine, and hydrochloric acid. At Neepawa, Manitoba, brine is obtained from a porous zone in the Upper Devonian Souris River Formation at 1,160 feet and from the Winnipegosis Formation at 1,453 feet depth. The brine contains 170,000 to 180,000 ppm of dissolved salts of which 85 per cent is sodium chloride.

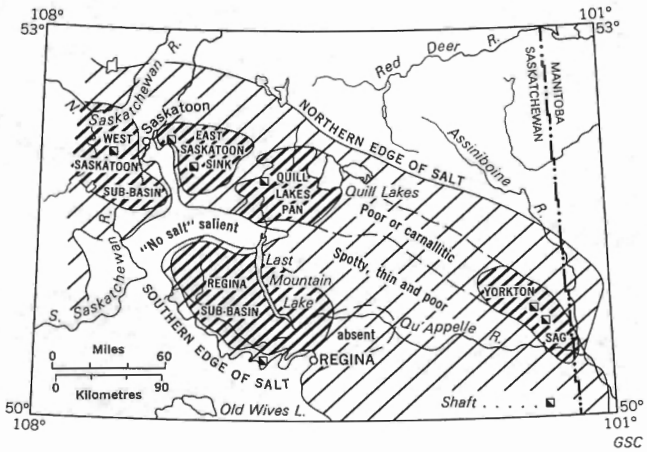


FIGURE IX-13. Potash occurrences in the Prairie evaporite basin, Saskatchewan (after A. M. Klingspor).

Potash

Potash occurs in the upper 300 feet of the Prairie Formation of the Middle Devonian Elk Point Group of Saskatchewan (Fig. IX-13). The potash-rich units (Klingspor, 1966), designated K-1, K-2, and K-3 zones, from base to top, are probably continuous across the basin but areas high in sylvinite and carnallite are separated by areas thin and poor in these minerals and with a high insoluble mineral content. The latter areas seem to correspond with high reef growth in the Winnipegosis Formation. Each zone consists of a variable succession of indistinctly bedded units of halite and lenticular unbedded masses of sylvinite (sylvite and halite), carnallite, and halite. The sylvinite contains the highest proportion of potassium but occurs only as thin zones. The percentage of potash in each zone increases gradually upwards until terminated abruptly by a clastic layer, generally of red anhydritic or halitic clay. The K-1 layer is the most coarsely crystalline and purest ore. The crystals range from 1 inch to 2 inches in the K-1 zone, whereas they range from 0.3 inch to 1.3 inches in the K-2 zone, and from 0.1 to 0.6 inch in the K-3 zone. Individual zones may be as much as 20 feet thick.

It is estimated that potash can be mined economically to depths of 3,500 feet. Commercial grade potash

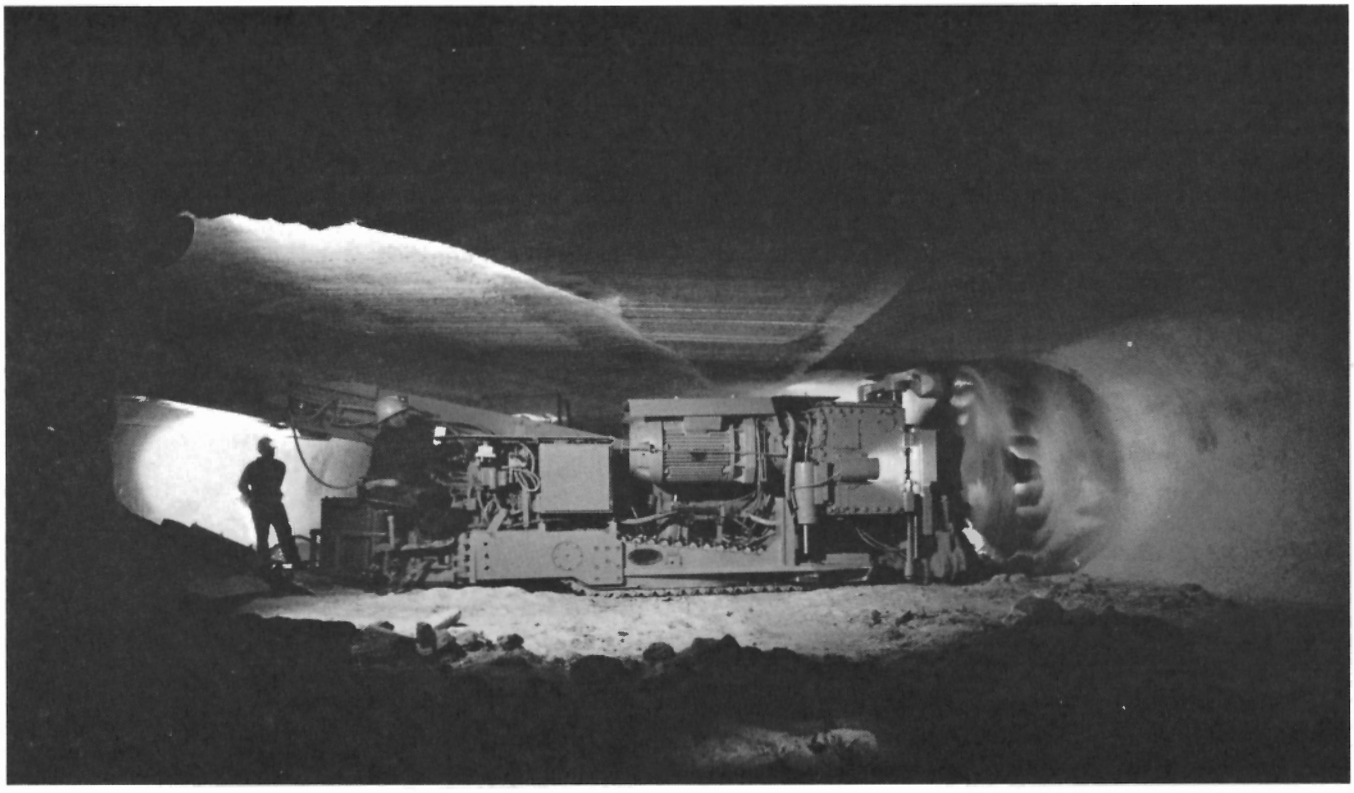


PLATE IX-2. Mining potash underground at Esterhazy, Saskatchewan.

is between 25 and 35 per cent K_2O. Recoverable reserves to a depth of 3,500 feet and grading 25 per cent or better with bed thicknesses of 5 to 10 feet have been estimated at 6.4 billion tons. Mines in operation by 1969 are at Patience Lake (12), Belle Plaine (13), Esterhazy (14), Lanigan, Viscount, Allan, and Vanscoy. At Belle Plaine the potash is obtained by solution of strata lying at over 5,000 feet depth. This plant is the only one in the world where potash is mined in this way. Salt is precipitated from the concentrated solutions and the residual solutions are fed to a crystallizer to precipitate the potash.

Other Sedimentary Deposits

Limestone, Dolomite

Limestone is one of the most valuable of the non-metallic minerals; it is used mainly in the production of cement. The Fort Whyte plant, Manitoba, uses limestone, 95 to 96 per cent $CaCO_3$, from the Devonian Elm Point Formation near Steep Rock and clay from surficial deposits nearby. High calcium limestone is obtained from the Devonian Point Wilkins Member of the Manitoba Group near Mafeking and shipped to the cement plant at Regina. In Alberta, the Upper Devonian Palliser limestone is quarried at Exshaw and Cadomin. At Bamberton on Vancouver Island, limestone is quarried from the Permian Sicker Group and shale is obtained from the Upper Cretaceous Nanaimo Group. The plant on Lulu Island near Vancouver obtains limestone from the Triassic Marble Bay Formation on Texada Island.

Lime is produced in Alberta from the Mississippian Loomis Member of the Mount Head Formation at Kananaskis, the plant formerly utilizing Cambrian limestone. The best bed is 70 feet thick, 95 to 99 per cent $CaCO_3$, and up to 2 per cent $MgCO_3$. Limestone from the Mississippian Livingstone Formation containing 97 per cent $CaCO_3$ and 2 per cent $MgCO_3$ is quarried for lime at Crowsnest Lake. This plant also produces crushed limestone for glass-making, poultry grit, stucco dash, coal-washing, and stock feed. The Devonian Elm Point Formation is quarried for lime at Spearhill, Manitoba. High magnesia lime and dolomite are quarried from the Ordovician Stonewall and Stony Mountain Formations and from the Silurian in Manitoba for crushed stone, rubble, curbstone, lime, and terrazzo. There are many excellent deposits of limestone throughout the Cordillera that are not at present being worked.

In Manitoba, a mottled dolomitic limestone containing large cephalopods and gastropods, known as Tyndall building stone, is obtained from the Ordovician Red River Formation, and a strikingly decorative mottled greenish and purplish grey dolomite building stone is obtained from the Stony Mountain Formation.

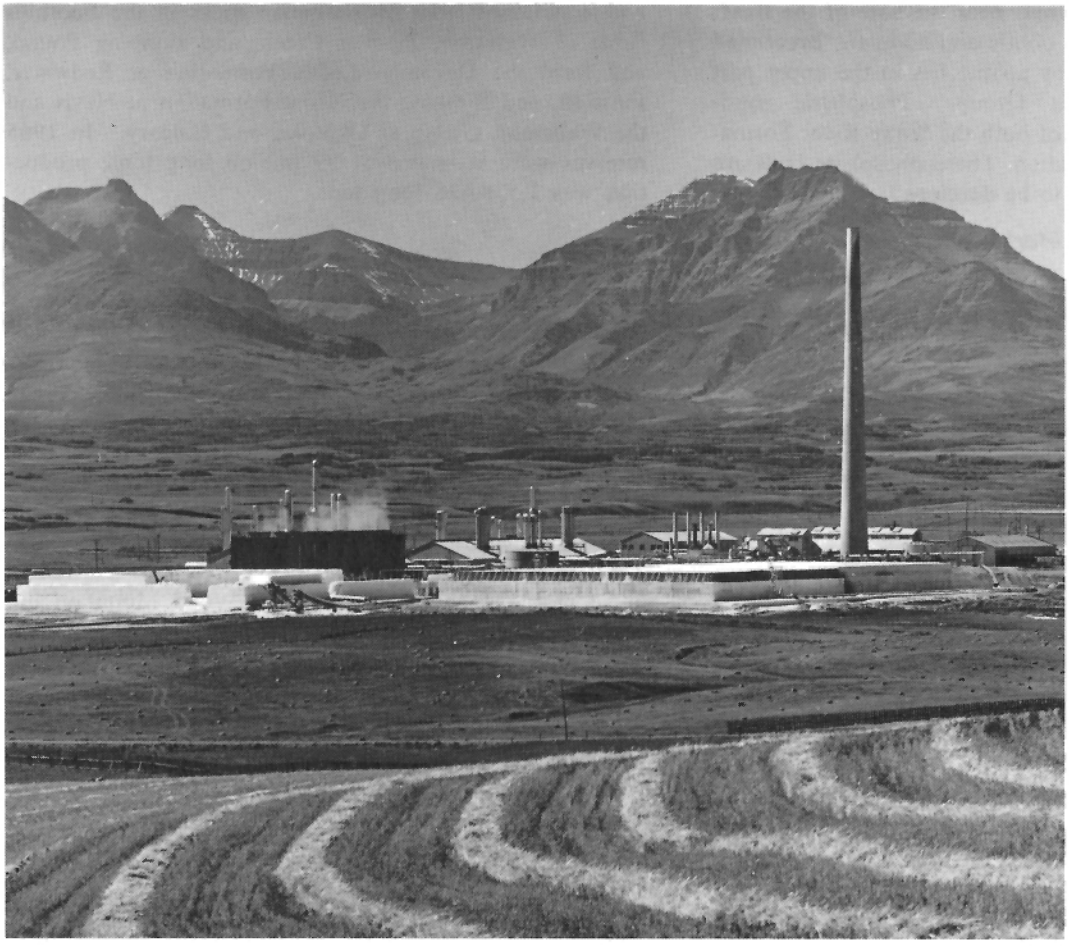


PLATE IX-3

Natural gas processing plant, Pincher Creek, Alberta.

Clay

In southern Saskatchewan clay for china, ball, fire, and stoneware products is obtained from the late Upper Cretaceous Whitemud Formation. Drain and sewer pipe, stone and household ware are manufactured from similar clay at Medicine Hat. Non-marine shales of Cretaceous and Tertiary age are used for local brick and tile industries. Shale from the Eocene Huntingdon Formation on Vedder Mountain in the lower Fraser Valley is the principal source of clay for the large tile, brick, pipe, and other clay products plants located nearby. Clay from surficial deposits has been used but is largely of inferior quality.

Bentonite

Bentonite is used for the production of well drilling mud, foundry moldings, and pellets. Non-swelling (calcium) bentonite is mined at Morden, Manitoba, from the Vermilion River Formation and is used in the purification of mineral and vegetable oils. Other deposits are known in the Battle Formation of southern Saskatchewan, in the Bearpaw and Edmonton Formations of south-central Alberta. The better types of swelling (sodium) bentonite are in the Edmonton and Bearpaw Formations of central Alberta and mining operations are located near Onoway and Rosalind. Semi-swelling bentonite is found in the Ravenscrag and Battle Formations in southern Saskatchewan. A little bentonite has been mined near Princeton, British Columbia, where it occurs in beds 3 and 11 feet thick separated by four inches of lignite, in the coal-bearing part of the Eocene Princeton Group. A seam of white bentonite 9 feet thick occurs in coaly shales of the same age near Quilchena.

Diatomite

Diatomite has been mined 8 miles north and 24 miles south of Quesnel, British Columbia. The diatomite is of Upper Miocene age, rests upon clay, sand, and gravel, and is in places overlain by basalt flows. Maximum thickness of the deposits is about 83 feet.

Phosphate

Phosphate rock of Mississippian to Jurassic age occurs in the southern Rocky Mountains near Fernie,

Michel, and Crowsnest. That near the base of the Banff and Exshaw Formations is oölitic and nodular. Brecciated dolomite partly cemented by apatite lies in the upper part of the Rocky Mountain Group. Phosphatic sandstone occurs near the base of both the Spray River Formation and the Fernie Formation. These phosphate beds are thin and of too low grade to be developed at present.

Marl

Many small lakes in British Columbia contain deposits of marl which has been used locally for agricultural purposes. Probably the largest of these deposits lies in Cheam Lake near Chilliwack, where the marl is 4 to 12 feet thick and covers about 4 acres. In recent years 25,000 to 40,000 tons per year has been obtained by dragline dredging.

Hydrocarbon Associates

Sulphur

Sulphur is obtained from the hydrogen sulphide in sour natural gas during the normal preparation of the gas for use as fuel. Most sulphur produced is from Alberta and is obtained from Mississippian rocks in the Foothills fields of Waterton, Pincher Creek, and Jumping Pound, and from the Devonian Leduc Formation at Redwater, Innisfail, and Rimbey, the Nisku Formation at Nevis and the Wabamun Group at Okotoks, and Calgary. In 1965 reserves were estimated at 99 million long tons; production was 1,538,428 long tons.

Helium and Nitrogen

Helium has been produced since 1963 from two wells near Swift Current, Saskatchewan, from the Upper Cambrian Deadwood Formation. The reservoir is finely porous, silty to sandy dolomite, dolomitic siltstone and sandstone that overlie knobs of Precambrian rhyolite rising several hundred feet above the general basement surface. The gas is over 90 per cent nitrogen and up to 2 per cent helium. Net pay thicknesses in the two wells are 108 and 168 feet with recoverable reserves being estimated at 2 billion cubic feet and an annual capacity of 12 million cubic feet. Helium also occurs at Wood Mountain but is not produced. There the Cambrian sands are thinner and helium content lower.

FUELS

Coal

Coal deposits are widely distributed in Western Canada (Fig. IX-14), the largest being in the Foothills and southern Rocky Mountains. Most coal mined comes from beds of Lower Cretaceous and Upper Cretaceous age and some from Tertiary strata. Coal production reached an all time high of 7.0 million tons in 1949, gradually declined to 5.1 million tons in 1963, but increased to 6.8 million tons in 1967. Of the last, 0.9 million ton was mined in the Alberta Foothills, 2.7 million tons from the Alberta Plains, 2.0 million tons from Saskatchewan, and 1.2 million tons from the Cordilleran Region of British Columbia and the Yukon. Coal production from Western Canada in 1967 accounted for 60 per cent of the total Canadian output of 11.4 million tons.

Western Cordilleran Region

On Vancouver Island the coal-bearing strata belong to the Nanaimo Group of Upper Cretaceous age. The Nanaimo is about 7,000 feet thick; the upper part is non-marine and the lower part consists of intertonguing marine and coal-bearing beds. The coal is of high volatile bituminous rank. The strata were deposited in basins on a deeply eroded, uneven surface of Mesozoic sedimentary, volcanic, and plutonic rocks. The seams are floored by sandstone which rarely contains the roots of vegetation from which the coal was derived. Accordingly, it has been suggested that the greater part of the coal was derived from driftwood deposited in lagoons. Mining commenced in the Nanaimo and Comox fields in 1852, and nearly all recoverable coal has been extracted. Production has dwindled rapidly with only 375 tons being produced in 1967. Coal deposits of the same age occur farther north at Suquash and on Graham Island, the most northerly of the Queen Charlotte Islands. Lignitic coal of Tertiary age is also found on Graham Island.

Coal seams of Upper Jurassic or Lower Cretaceous age occur in rocks included with the Hazelton Group in Bulkley River valley and in Kispiox area north of Hazelton. The coal-bearing strata are cut by minor igneous intrusions and in places lie near younger volcanic rocks. As a result, the coal ranges in rank from low grade bituminous to low grade anthracite, depending on proximity to igneous rocks. Seams are as much as 14 feet thick, but have been folded and crushed. In 1967 production from the Telkwa coalfield near Telkwa amounted to 16,000 tons. The strata of the remotely situated Groundhog coalfield at the headwaters of Skeena River are presumably of the same age. The seams are closely folded and faulted. The coal is sheared, of semi-anthracite rank, but very high in ash.

The principal coal mining area in Yukon is at Carmacks where about 1,900 tons was mined in 1967. Production is from a seam 9 feet thick in the Tantalus Formation of early Lower Cretaceous age. This formation occurs in several small, isolated areas between Carmacks and Whitehorse. A little Tertiary coal has been mined about 40 miles north of Dawson.

Coals of Tertiary age are widely distributed over the Western Cordillera. The principal producing areas have been in the south near Princeton and Merritt where production ceased in 1961 and 1963 respectively. At Hat Creek the seams are exceptionally thick, totalling 466 feet. The coal is very dirty but may be, in part, recoverable by stripping.

Rocky Mountains and Foothills

In the southern Rocky Mountains and Foothills, coal deposits occur mainly in the Kootenay Formation of Lower Cretaceous age. The coal is mostly bituminous and locally semi-anthracite. Much is sheared and the strata may be highly inclined, closely folded, and repeated by

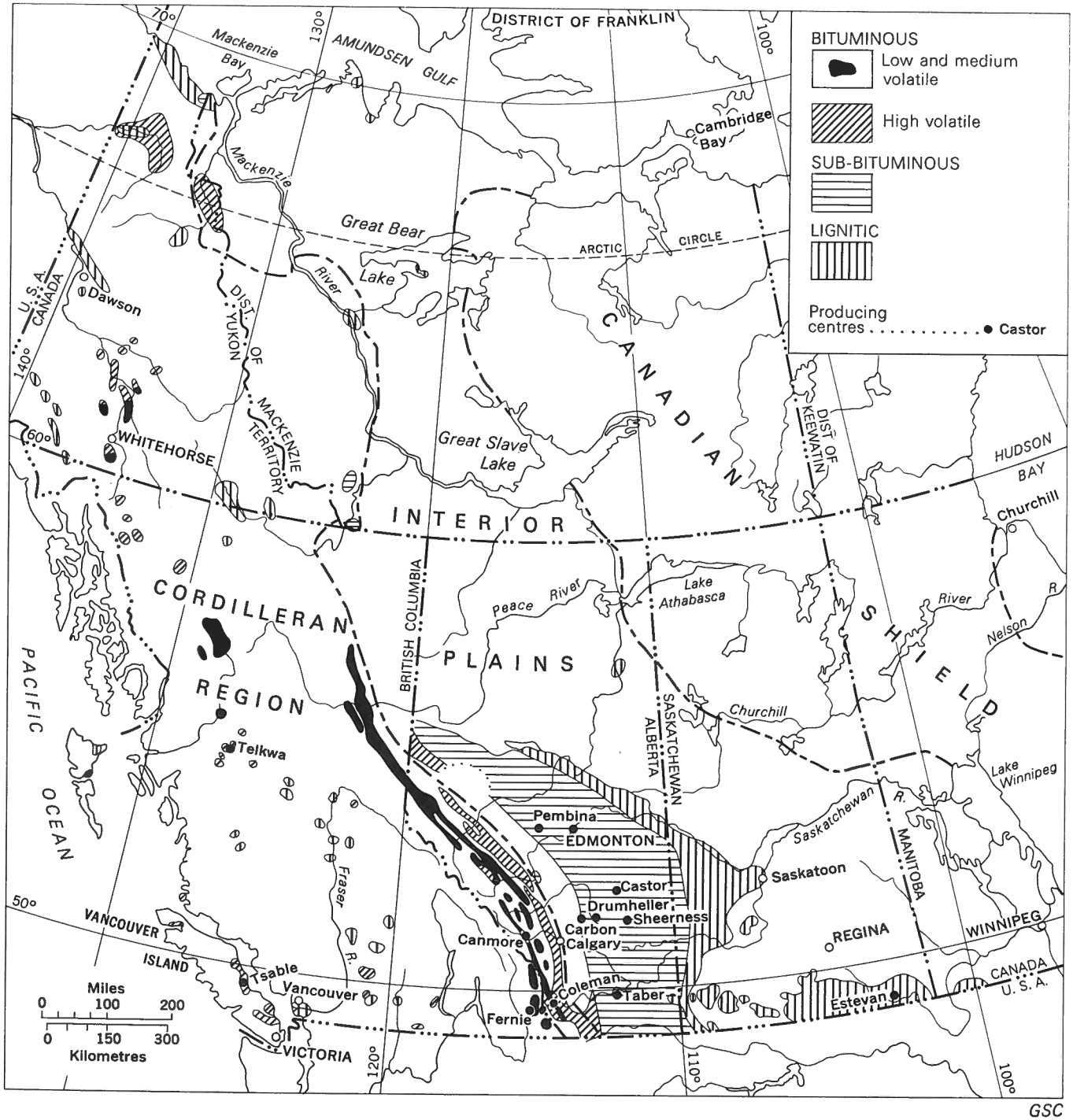


FIGURE IX-14. Coal-bearing areas of Western Canada, classified according to rank (by B. A. Latour).

thrust faults. Some is hard and blocky where the seams are less disturbed.

In southeastern British Columbia, Crowsnest coalfield comprises three areas, namely: the undeveloped upper Elk River area to the north; the central Crowsnest area in which there is large production at Michel; and the undeveloped Flathead River area to the south. In the Crowsnest area the Kootenay is about 3,500 feet thick and contains 22 seams with an aggregate thickness of 150 feet. At Michel, Corbin, and Tent Mountain the strata have been folded and faulted and the coal squeezed from the limbs into pockets on the crests and troughs of the folds where it is several times thicker than normal. Where such conditions occur near surface the coal can be economically recovered by strip mining. In 1967 the Crowsnest area produced 1.2 million tons and a cumulative production of 62 million tons. The coal is of medium volatile bituminous rank and has excellent coking qualities. Probable recoverable reserves of the Crowsnest coalfield are estimated at 5 billion tons.

In the Crowsnest coal area of Alberta the Kootenay Formation decreases in thickness from 800 feet at Coleman where five seams aggregate 47 feet, to 430 feet at Bellevue where three seams total 37 feet. The coal has been obtained from underground and strip mines, but at present only the Vicary mine near Coleman is producing. The coal is of medium and low volatile, bituminous rank and can be used for the production of coke. Production in 1967 totalled 621,000 tons; probable recoverable reserves are 3.0 billion tons.

Along Bow River valley in the vicinity of Canmore, coal from the Kootenay Formation has been mined continuously since 1888. The coal measures are about 1,000 feet thick, dip steeply west, and lie within the Paleozoic strata of the Front Ranges. The coal is of low volatile, bituminous rank with small pockets of semi-anthracite. Coal of coking quality is being mined at Canmore. Production in 1967 was 289,000 tons; probable recoverable reserves are 1.1 billion tons. Between Canmore and Crowsnest Pass much coal is known, principally along Kananaskis, Sheep, Highwood, and Oldman Rivers, and forms a large reserve estimated to be 3.3 billion tons. Some small operations have been carried out, but as the area can not be reached by rail development is in the prospecting stage.

In the central Foothills of Alberta, coal has been mined from the Luscar Formation of Lower Cretaceous age at Luscar, Mountain Park, Cadomin, and Nordegg. The Luscar Formation is about 1,000 feet thick and contains several seams. In its upper part two seams, 6 and 40 feet thick, have been mined. The coal is of low and medium volatile, bituminous rank. Mining ceased in 1956.

The basal beds of the Tertiary Paskapoo Formation are coal-bearing at Mercoal, Sterco, Coalspur, Hinton, and Saunders. Four seams lie on the flanks of broad folds or are contorted and thickened by minor folds and faults into large masses of sheared coal. The coal is of high volatile, bituminous rank. Production ceased in 1965; probable reserves of recoverable coal are 2.4 billion tons.

In the Peace River area, coal deposits are found in the Lower Cretaceous Gething Formation of the Bullhead Group. Considerable reserves are known nearby in the Carbon Creek basin and on Hasler Creek. Along Liard River at the south end of Mackenzie Mountains, and in the Mackenzie River valley, coal seams of Cretaceous and Tertiary ages occur but are undeveloped.

Interior Plains

Coal has been mined in many parts of the Interior Plains. Annual production from Saskatchewan (all lignite) has been constant at about 2 million tons since 1954; that from the plains area of Alberta has fluctuated between 2.0 and 2.6 million tons. In Saskatchewan coal is produced from the Tertiary Ravenscrag Formation near Estevan. Production in 1967 was 2,008,147 tons; the coal was used mainly for generation of electricity.

The coal deposits of the Alberta Plains occur mostly in Upper Cretaceous formations, but a few mineable beds are found in Paleocene formations. Most important Upper Cretaceous formations are the Foremost, Oldman, and Edmonton. Coal seams in the Foremost are mined at Taber and Milk River as well as at smaller mines as far east as Medicine Hat. At Lethbridge and Magrath the Oldman contains several coal seams. These are a good grade of high volatile, bituminous coal in contrast to the sub-bituminous rank of most other coal of this age on the plains. Beds equivalent to the Foremost and Oldman also contain mineable coal seams in the Peace River region. The Edmonton Formation is the main source of coal in central Alberta, the principal producing centres being Wabamun, Castor, and Drumheller. Five workable seams are known in the Drumheller area where production is mainly from underground mines. The coal is sub-bituminous, non-coking, and tends to be friable. Total production in 1967 from the Alberta Plains area was 2,664,000 tons.

Petroleum and Natural Gas

Ordovician Fields

Ordovician oil is produced from the Red River Formation from a single well near South Alma, Saskatchewan. Some 70,000 barrels have been recovered to date from a reservoir of bryozoan, non-skeletal calcarenite. It grades up-dip into dense dolomite and is capped by anhydrite and shale. A well at Hummingbird produced oil for a short time.

Devonian Fields

The major producing system in Western Canada is the Devonian, the province of Alberta yielding by far the

TABLE IX–3 *Proved recoverable reserves of crude oil in Western Canada to December 31, 1966, by province and geological age of reservoir (thousands of barrels of 34.97 gallons at 60°F).*

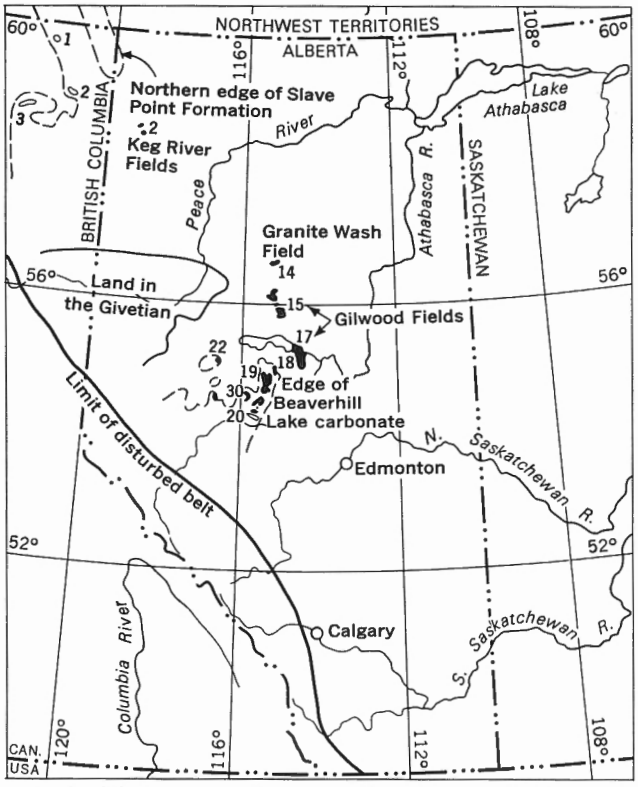
Reservoir		Alberta	Sask.	Man.	B.C.	N.W.T.	Total	%
Upper Cretaceous		1,460,096					1,460,096	17.9
Lower Cretaceous	Viking*	21,719	36,823				58,542	0.7
	Mannville*	129,462	38,285		8,496		176,243	2.2
Jurassic		33,620	233,779				267,399	3.3
Triassic		17,562			245,826		263,388	3.2
Permian					337		337	—
Mississippian		148,086	672,313	58,330	13,090		891,819	11.0
Upper Devonian	Wabamun*	8,167					8,167	0.1
	Nisku*	256,998					256,998	3.2
	Leduc*	1,760,181					1,760,181	21.7
	Beaverhill Lake*	2,105,064					2,105,064	25.9
Middle Devonian	Kee Scarp					47,125	47,125	0.6
	Slave Point	4,280					4,280	0.1
	Keg River	298,313					298,313	3.7
	Gilwood, Granite Wash	347,156					347,156	4.3
Ordovician			33				33	—
Unclassified		170,028			1,081		171,109	2.1
Totals		6,760,732	981,233	58,330	268,830	47,125	8,116,250	100.0
%		83.3	12.1	0.7	3.3	0.6	100.0	

* Includes stratigraphic equivalents

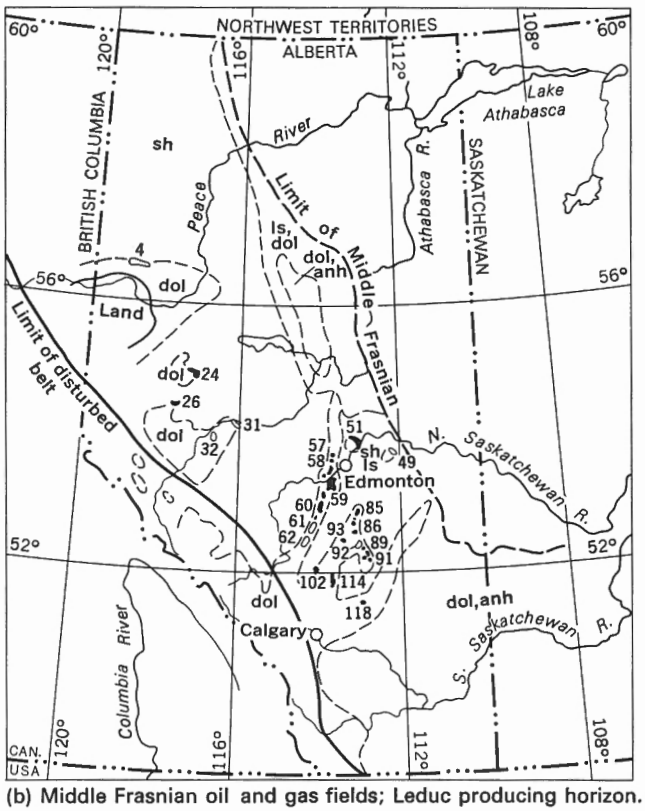
TABLE IX–4 *Proved recoverable reserves of natural gas in Western Canada to December 31, 1966, by province and geological age of reservoir (billions of cubic feet at 14.65 psi and 60°F).*

Reservoir		Alberta	Sask.	Man.	B.C.	N.W.T.	Total	%
Upper Cretaceous		3,056	304				3,360	7.3
Lower Cretaceous	Viking*	4,463	520		15		4,998	10.8
	Mannville*	5,539	89		375		6,003	13.0
Jurassic		497	33				530	1.1
Triassic		230			4,172		4,402	9.5
Permian		85			115		200	0.4
Mississippian		12,379	13		100		12,492	27.0
Upper Devonian	Wabamun*	2,966			67		3,033	6.6
	Nisku*	878					878	1.9
	Leduc*	5,345					5,345	11.6
	Beaverhill Lake*	1,384					1,384	3.0
	Other							
Middle Devonian		64			2,213	117	2,394	5.2
	Gilwood	260					260	0.6
Unclassified		967					967	2.0
Totals		38,113	959		7,057	117	46,246	100.0
%		82.4	2.1		15.3	0.2	100.0	

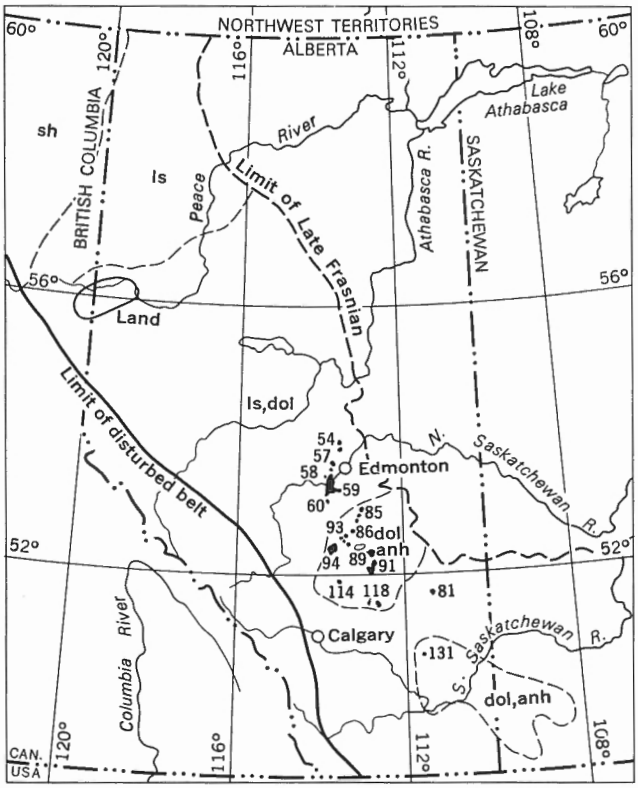
* Includes stratigraphic equivalents



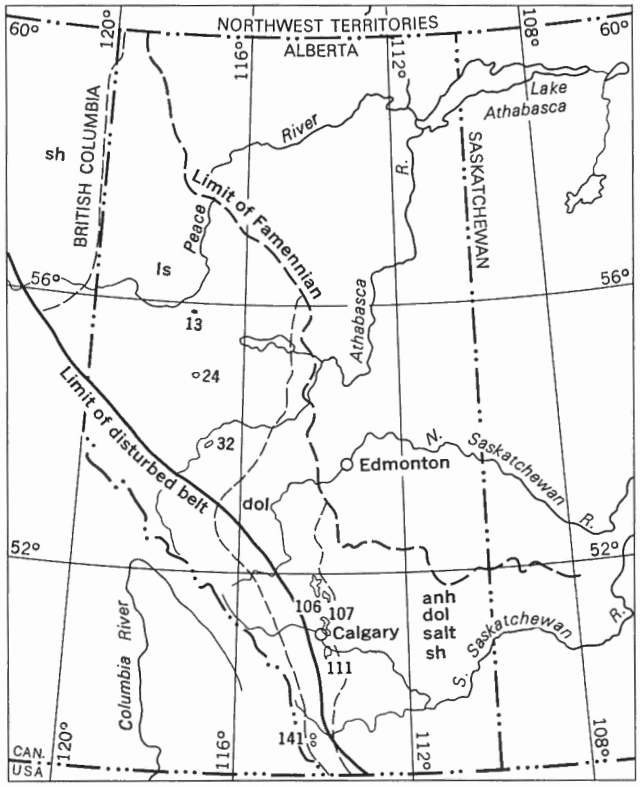
(a) Middle Devonian, Givetian and Upper Devonian, early Frasnian oil and gas fields.



(b) Middle Frasnian oil and gas fields; Leduc producing horizon.



(c) Late Frasnian oil and gas fields; Nisku and Camrose producing horizons.



(d) Famennian oil and gas fields; Wabamun producing horizon.

GSC

Scale: 1 inch to 190 miles or 305 kilometres

FIGURE IX-15. Devonian oil and gas fields of Western Canada. Oil fields shown as solid black areas, gas fields shown as open areas; fields numbered according to Map 1252A, data to end of 1965 (by B. MacLean).

most and containing the greatest reserves. Many different horizons are productive; some in the Middle Devonian and others, containing the bulk of the oil and gas, are in the Upper Devonian. The Devonian contains 59.6 per cent of the oil in Western Canada and 28.9 per cent of the gas (Tables IX-3 and 4). Production in 1966 was 140 million barrels of oil, 43 per cent of the total, and 510 million Mcf of gas, 32 per cent of the total for Western Canada.

Middle Devonian Fields

Granite wash. Basal Devonian sandstones commonly called 'granite wash' (Fig. IX-15a) fill irregularities on the Precambrian surface in northern Alberta. The sandstones are quartzose to arkosic with some dolomitic or anhydritic cement. Some are the near-shore equivalents of Elk Point carbonates and evaporites, but some may be older. In the Red Earth field (14) porosity is about 13 per cent, and the oil of 38 degrees A.P.I. is produced from 4,700 feet depth. In 1966 the granite wash yielded 2,759,159 barrels of oil and 120,375 Mcf of gas.

The *Keg River Formation* in northwestern Alberta was found productive of oil in 1965. The Rainbow (2), Rainbow South (3), Zama Lake (1), and Zama North areas include many separate fields and numerous discoveries, each in an apparently separate pool. The main producing horizon is the Rainbow Member of the Keg River Formation which forms pinnacle reefs or mounds that rest on the lower Keg River fine-grained crinoidal limestone platform. The reservoir commonly is a stromatoporoid and coral bioherm, overlain by laminites, presumably algal. The Keg River is commonly dolomitized to extremely coarsely crystalline white dolomite. Non-porous dolomite, anhydrite, and halite of the Muskeg Formation lie adjacent to and overlie the carbonate mounds. Production is from depths of 5,700 to 6,000 feet through 200 to 400 feet of strata at the Rainbow A, B, and F pools. Porosity ranges from 4 to 13 per cent, and horizontal permeability varies from 50 to 780 md. The oil has gravities of 40 to 42 degrees A.P.I. and 34 to 35 degrees A.P.I. in various pools.

Slave Point and Sulphur Point Formations. Gas occurs in the Slave Point Formation in the Petitot (1), Kotcho Lake (2), and Clarke Lake (3) fields of northeastern British Columbia (Figs. IX-15a, -16). The reservoir is a grey and white mottled, coarsely crystalline, vuggy dolomite resulting from several periods of dolomitization, that completely obscured the original lithology (Pl. IX-4b). The undolomitized deposits consist of a rubble of coarse stromatoporoid fragments set in a fine- to coarse-grained limestone matrix or cemented by sparry calcite. The reefs and associated porous rocks occur along the front of a carbonate platform flanked by greenish grey shales. Dips of carbonate beds at the front are 25 to 35 degrees. Porosity and permeability are greatest at

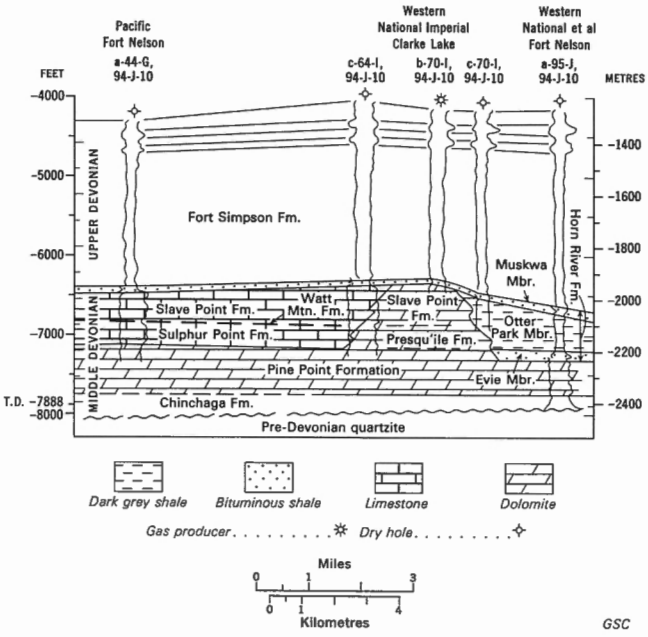


FIGURE IX-16. Cross-section through Middle Devonian Clarke Lake field, British Columbia (after Gray and Kassube).

the carbonate front and decrease with increasing distance from it. The porosity resulting from early and intermediate dolomitization is predominantly intercrystalline, averaging 10 to 12 per cent and reaching as high as 17 per cent; the permeability is less than 100 md. Later dolomitization increases porosity somewhat but also the voids may be filled by late precipitation of dolomite. The gas is trapped by a facies change to impermeable limestone. Maximum net gas pay is about 360 feet and depth to production is about 6,200. Production in 1966 was 42.6 million Mcf of gas.

In the Rabbit Lake gas field of southern District of Mackenzie the Sulphur Point reservoir is a stromatoporoid-coral limestone with good intergranular porosity. The accumulation occurs on the crest of an anticline on the northwest upthrown side of a small northeast-trending normal fault.

Gilwood Formation. The Mitsue (17) and Nipisi fields (15) produce oil from the Gilwood sandstones of the Elk Point Group (Fig. 15a). They are quartzose sandstones with scattered feldspar grains, poorly sorted, angular to subrounded, and interbedded with green pyritic shales. Porosity is 13 to 14 per cent. The oil is trapped by an up-dip pinchout of sand. Pay thickness varies from 10 to 40 feet at depths of about 4,900 feet. The oil at Mitsue is 42.9 degrees A.P.I. In 1966, 7,684,686 barrels of oil and 3.05 million Mcf of gas were produced.

Kee Scarp Formation. The Norman Wells field (1), discovered in 1920, produces from the Kee Scarp Formation on the homoclinal west flank of Norman Range of

Franklin Mountains. The producing horizon is the upper Kee Scarp reef, a light grey and buff limestone composed of abundant stromatoporoids and corals in a matrix composed largely of coarse, grain-supported, skeletal debris, much derived from stromatoporoids. The lower member of the Kee Scarp forms an extensive limestone platform overlying Hare Indian black shale. Bituminous shales of the Canol Formation cap the reef. The oil is 34.4 degrees A.P.I., 749,653 barrels being produced in 1965.

Upper Devonian Fields

Beaverhill Lake Formation. Oil of 41 to 45 degrees A.P.I. is produced from the Swan Hills Member of the Beaverhill Lake Formation in several fields south of Peace River Arch (Fig. IX-15a). Production in 1966 was 54,283,011 barrels of oil and 59.4 million Mcf of gas. The oil is trapped in reef-fringed banks covered and separated from one another by shale and shaly limestone. The banks, 400 to 500 feet thick, lie near the margins of the Beaverhill Lake carbonate platform. The facies developed in the Judy Creek (19) field (Murray, 1966) are a typical example of the reservoirs (Fig. IX-17). The platform carbonates are predominantly brown, fine-grained limestones, dolomitic limestones, coral beds and fine- to coarse-grained, largely skeletal, limestones carrying stromatoporoid and *Amphipora* fragments, and skeletal calcarenites. Porosity in the platform beds is largely restricted to lenses of grain-supported skeletal limestones. The banks resting on the platform are made of a central core of lime mudstone which contains patches of non-skeletal, pelletoid limestone with a little porosity. Dendroid stromatoporoids, *Amphipora,* ostracods, and gastropods are common organisms. Towards the margin of the banks the fine-grained limestones intertongue with stromatoporoid-algae reefs consisting of tabular, dendroid, and massive stromatoporoids in a skeletal calcarenite matrix that is largely derived from stromatoporoids and has excellent to good porosity. Coarse, porous, calcirudites and bedded calcarenites are associated with the reefs and overlie the major part of the bank. They have good intergranular porosity and together with the reef facies form the best producing zones of the Swan Hills Member. The banks grade laterally to and are overlain by impermeable shales and fine-grained shaly limestones of the Beaverhill Lake Formation.

The *Leduc Formation* of the Woodbend Group contains some of the major oil and gas fields of Alberta (Fig. IX-15b). The fields are too numerous to list. In 1966 they produced 52,788,795 barrels of oil and 285.4 million Mcf of gas. The oil and gas have accumulated in stratigraphic traps formed by reef complexes covered and separated by impervious greenish grey shales. The reefs developed on shoals composed predominantly of biostromal, stromatoporoid-rich beds and skeletal lime sands comprising the Cooking Lake platform. Commercial production from Cooking Lake has been obtained only at Skaro.

The reservoir of the Redwater (51) oil and gas field (Klovan, 1964) is one of the few that are undolomitized and provides an excellent example of the internal constitution of Devonian reef complexes (Figs. IX-18, -19). Other fields of this type are Golden Spike (59b) and Duhamel (85a). The upper, producing part of the reef complex is composed of organically bound massive stromatoporoid reef limestones, 3 to 30 feet thick, interbedded with coarse, skeletal calcarenites derived largely from stromatoporoid debris. Calcilutites occur sporadically on the northeast front of the Redwater complex

PLATE IX-4. PHOTOMICROGRAPHS OF TYPICAL DEVONIAN CARBONATE ROCKS:

a. Stromatoporoid growth enveloped by calcarenite (boundstone), granular calcite cement. Middle Devonian Sulphur Point Formation, Imperial Sun Arrowhead Aurora I-46, I-46-60-50-122-15, at 6,338 feet depth. Natural size. GSC 113782-C.

b. Dolomite pseudo-breccia caused by several stages of dolomitization; vuggy porosity. Middle Devonian Presqu'ile Formation, Clarke Lake gas field, B.C. Natural size. (Gray and Kassube, 1963).

c. Lime mudstone with more than 10 per cent comminuted skeletal debris (wackestone); gastropods, ostracods. Middle Devonian Slave Point Formation, California Standard Steen River 2-22, 2-22, 117-5 W 6, at 4,332 feet depth; x5. GSC 113781-E.

d. Non-skeletal calcarenite (grainstone), poorly sorted, bedded; a few algal-coated stromatoporoid and coral fragments; intergranular, intrafossil and small vug porosity. Middle Devonian Slave Point Formation, Pan-American Shell Kakisa F-35, F-35-61-00-117-15, at 1,718 feet depth. Natural size. GSC 113782-B.

e. *Amphipora* and non-skeletal grains in lime mud matrix (packstone). Upper Devonian Swan Hills Member (lagoonal facies), Beaverhill Lake Formation. Home Swan Hills A-10-20, 10-20-67-10 W 5, at 8,113 feet depth; x⅔. (R. W. Edie, 1961)

f. Skeletal calcirudite with lime mud matrix (skeletal lime packstone containing stromatoporoids, corals, ostracods, foraminifera). Upper Devonian Leduc Formation. Canadian Superior Holt 15-36, 15-36-44-22 W 4, at 4,875 feet depth; x5. GSC 113781-L.

g. Non-skeletal calcarenite (grainstone): grains and lumps of calcilutite, some coated; granular calcite cement. Upper Devonian Peechee Member, Southesk Formation. B.A. Hand Hills No. 1, 7-14-30-17 W 4, at 5,942 feet depth; x5. GSC 113781-G.

h. Massive stromatoporoid fragments in coarse-grained skeletal calcarenite matrix with intergranular porosity. Upper Devonian Leduc Formation, Imperial Golden Spike No. 11, 11-23-51-27 W 4, at 5,367 feet depth. Natural size. GSC 113782-A.

i. *Amphipora* limestone (wackestone). *Amphipora* in lime mud matrix with comminuted skeletal debris. Upper Devonian Leduc Formation, Imperial Redwater 56, 9-18-57-21 W 4, at 3,297 feet depth; x5. GSC 113781-K.

j. Crystalline dolomite; anhydrite partly fills intercrystalline voids; crossed nicols. Upper Devonian Nisku Formation, B.A. Pyrcz No. 1, 12-25-50-26 W 4, at 4,907 feet depth; x5. GSC 113781-F.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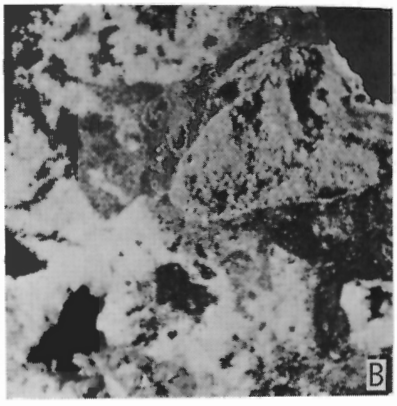
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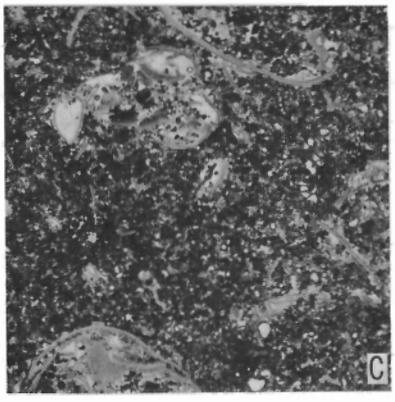
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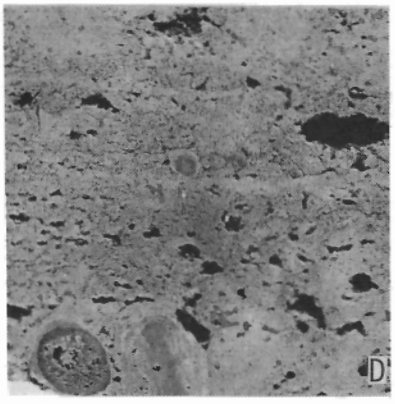
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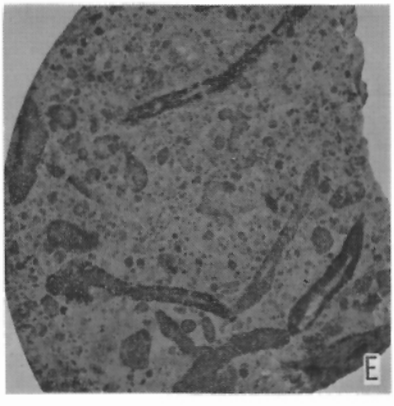
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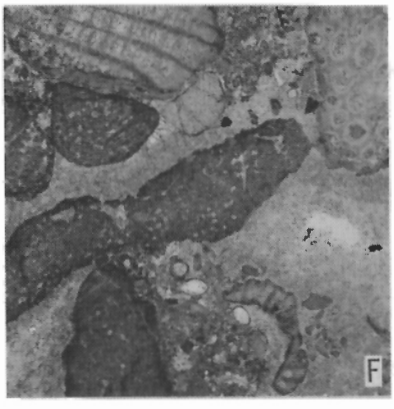
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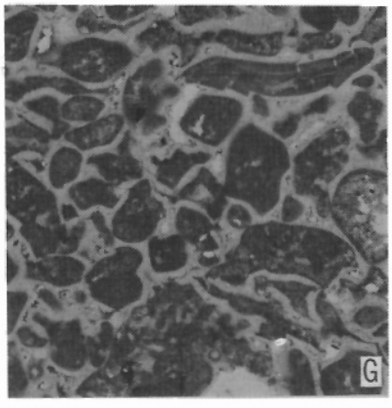
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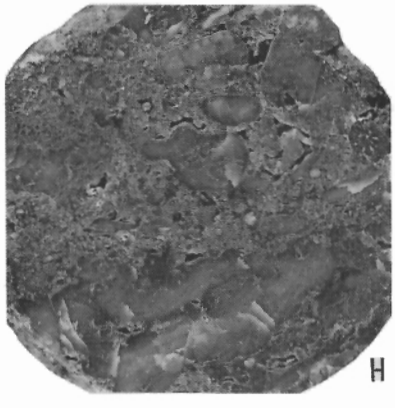
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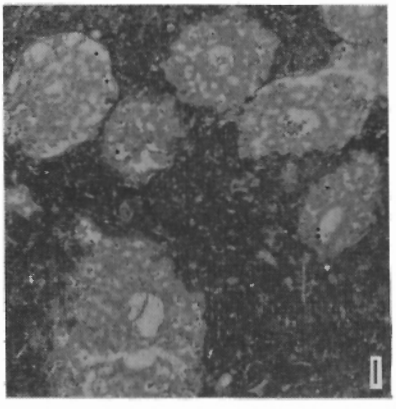
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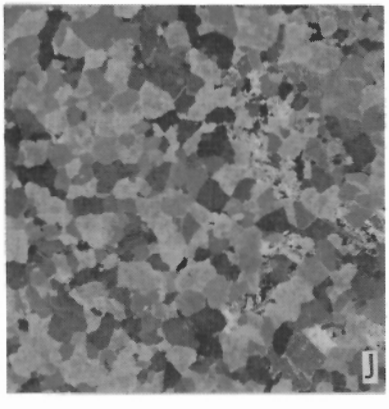
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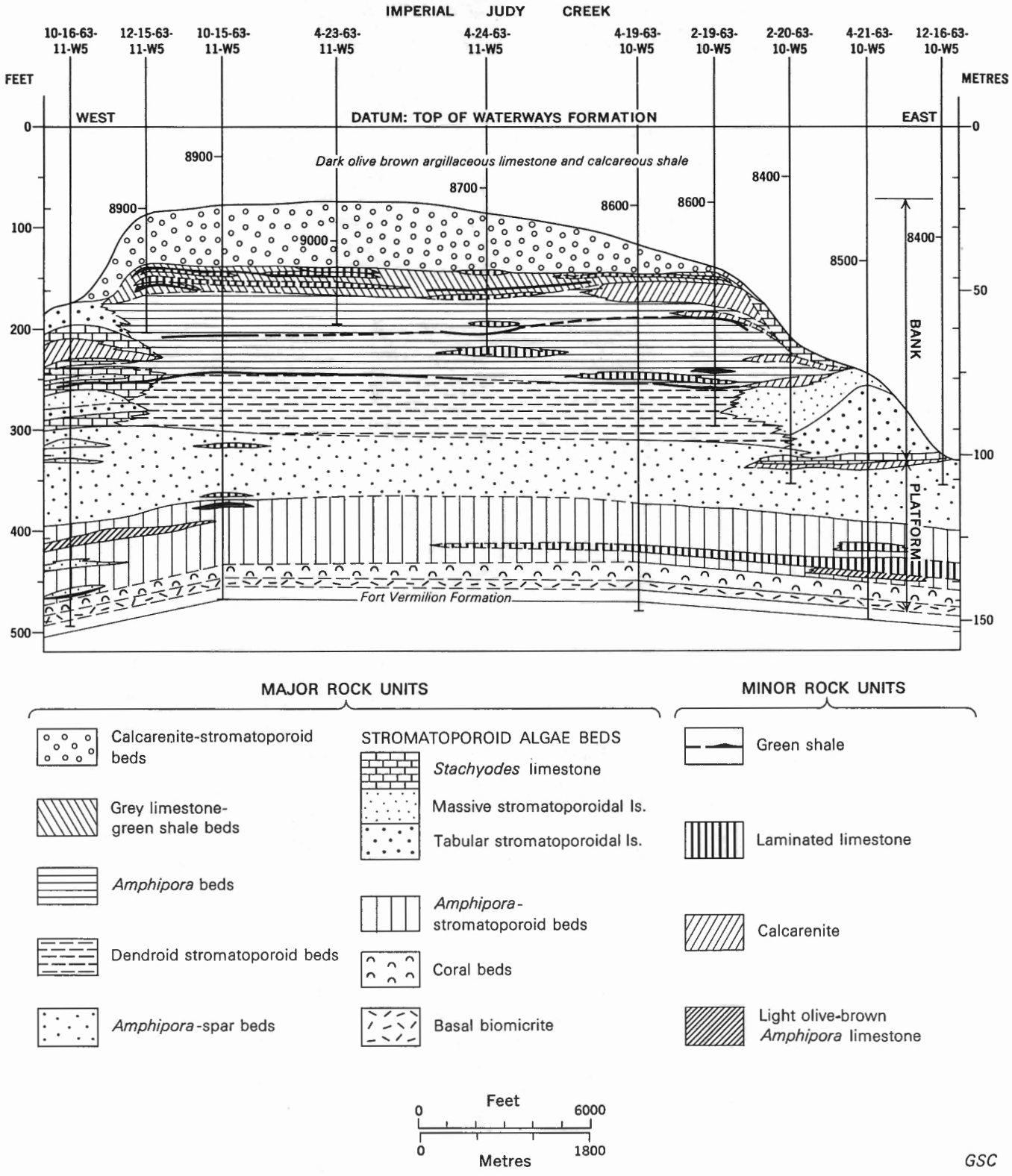


FIGURE IX-17. Cross-section through Upper Devonian, Swan Hills reef, south Judy Creek field, Alberta (after J. W. Murray).

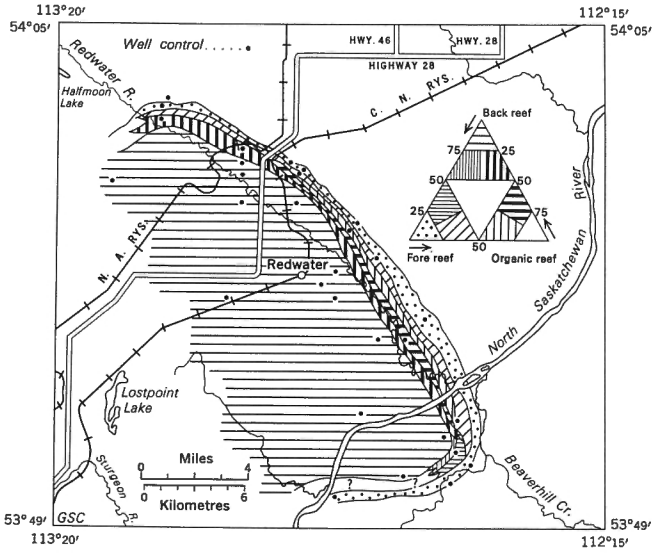


FIGURE IX-18. Distribution of major facies in upper Leduc Formation, Redwater reef complex, Alberta (after J. E. Klovan).

at least 25 to 50 feet below its highest part. In the fore-reef talus there is a progressive change in fauna from fragmented massive stromatoporoids to tabular stromatoporoids, colonial corals, dendroid stromatoporoids, *Megaladon*, and tabulate and rugose corals set in a coarsely fragmented matrix. Oncolites and algal coatings on fossil fragments are common. Farther off-reef the sediments become finer and dips reach 20 degrees. Within the reef complex, the organic facies intertongue with cleanly washed, porous, skeletal sands, in places containing large stromatoporoid fragments, or composed of finely comminuted grain-supported debris and lime mudstone. In the interior of the complex are laminated limestones, lime mudstone with *Amphipora*, and non-skeletal calcarenites containing pellets, lumps, and coarse structureless grains of various shapes and sizes largely cemented by sparry calcite but with some scattered intergranular porosity. The best porosity, both intergranular and vuggy, occurs in the grain-supported skeletal calcarenites and in the organically bound reefs or in the coarse skeletal calcarenites and calcirudites of the talus slopes dipping into the basin. Porosity is also present in the non-skeletal calcarenites and *Amphipora* calcarenites and calcilutites. Oil of 34 to 36 degrees A.P.I. gravity has accumulated on the up-dip northeast flank of the structure. Porosity averages 6.53 per cent and permeability is excellent in the producing zones, maximum oil zone thickness being 220 feet and average 100 feet.

In most oil and gas fields the Leduc Formation is composed of fine to medium crystalline dolomite with intercrystalline and vuggy porosity. The trap is a mound-shaped structure enclosed and covered by greenish grey impervious marine shales, and it is generally assumed that the initial internal constitution of these dolomite bodies was the same as that of the Redwater reef complex described above. The platform beds of the Cooking Lake Formation are similar to those at Redwater and the location of the mounds is similarly controlled by build-up of stromatoporoid-rich shoals where dolomitization has failed to destroy the stromatoporoid structure completely. Some production has been obtained from the lower part of Leduc in some fields along the Rimbey–Meadowbrook reef chain. The upper part, in which most oil and gas accumulations occur, is light grey to buff, medium to coarsely crystalline dolomite, with vuggy porosity and some

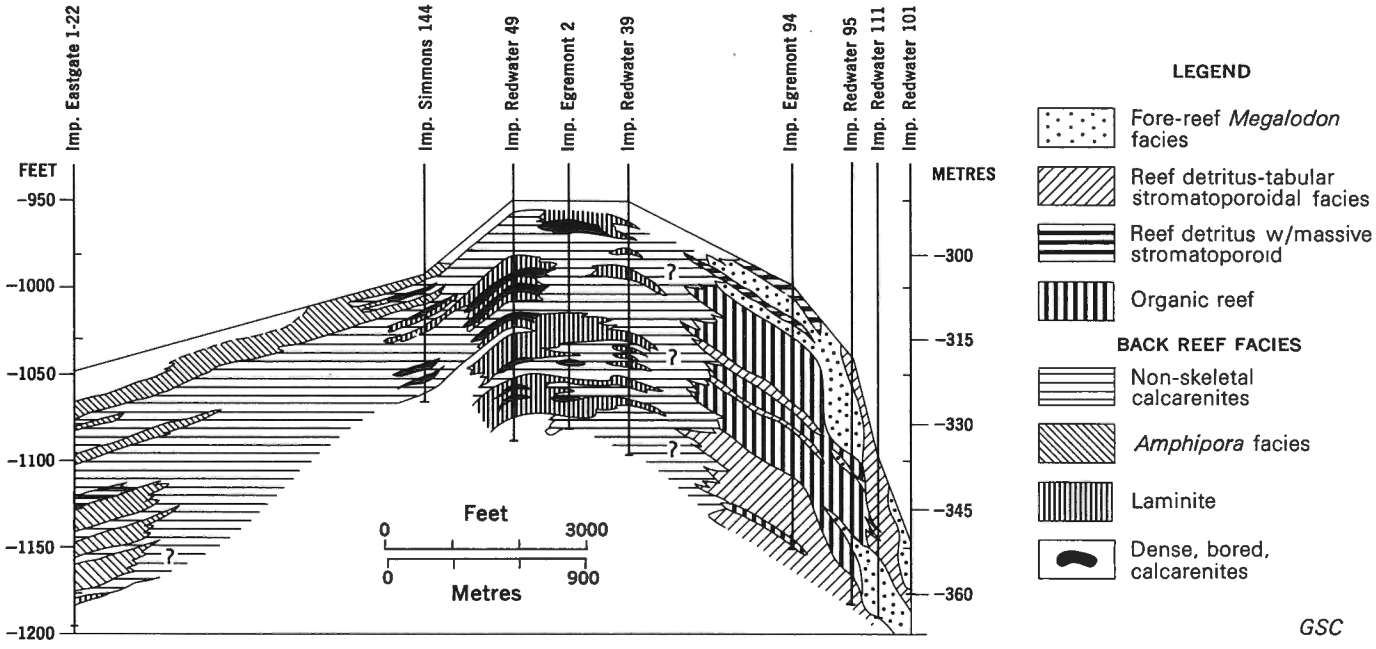


FIGURE IX-19. Composite cross-section through the upper Leduc Formation, Redwater reef complex, Alberta (after J. E. Klovan).

intercrystalline porosity and fractures. Dolomitization has completely destroyed all vestiges of the original structure except for occasional ghosts of stromatoporoids and corals. Fields along the Rimbey–Meadowbrook reef chain (Fig. IX-20) produce both oil and gas, the thickness of the gas cap increasing towards the south. Porosity in these fields ranges from 4 to 10 per cent and permeability is good. Gravity of the oil ranges from 27 degrees at Fairydell (52) to 44 degrees A.P.I. at Westerose (61). In the Bashaw (86) carbonate complex from which the Malmo-Duhamel (85) chain extends north-northeast and in the Stettler and Big Valley fields to the southeast pay thicknesses are generally less than in the Rimbey–Meadowbrook chain but in other respects they are similar. Porosities range from 5 to 8 per cent and gravity of oil from 27.7 to 43.3 degrees A.P.I.

West of the Rimbey–Meadowbrook reef chain production is largely gas. Light grey crystalline dolomite with vuggy and intercrystalline porosity forms reef complexes and banks enclosed by shale and resting on a widespread stromatoporoid-rich carbonate. Depths to production are great and the area has not yet been completely

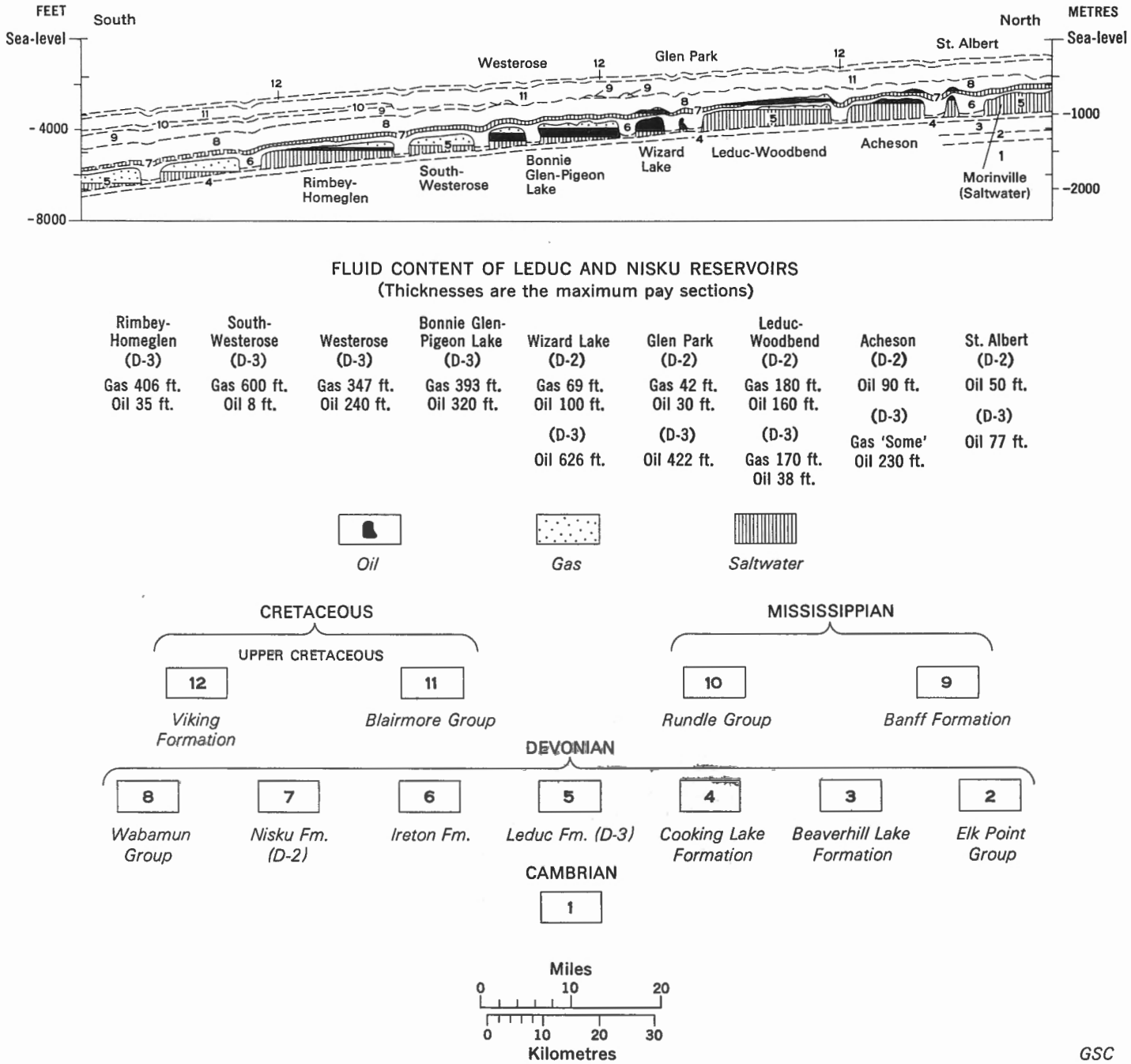


FIGURE IX-20. Fluid content of Upper Devonian Leduc and Nisku reservoirs, Bonnie Glen–Wizard Lake reef trend, Alberta (after W. C. Gussow).

explored: 8,700 feet at Windfall (31), 11,460 feet at Simonette (26), and 12,500 feet at Berland River. Pay zones vary from 142 to 350 feet, porosities from 4.5 to 7.9 per cent, and gravity of the oil from 37 to 48 degrees A.P.I.

Dolomite considered to have originated from reefs fringing the Peace River landmass produced gas at Worsley (4) and oil from 169 feet of pay at Normandville (13a). A gas pool at the up-dip eroded edge of the Grosmont Formation has been partly explored at Craigend. Porosity averaging 10.7 per cent in the form of vugs and fractures occurs in finely crystalline and sucrosic weathered dolomites. They have a tarry oil stain and are overlain by shales of Cretaceous age.

The *Nisku Formation* of the Woodbend Group is productive of oil and gas where overlying Leduc fields of the Rimbey–Meadowbrook reef chain north of Bonnie Glen, and together with the subjacent Camrose Formation where overlying those of the Bashaw complex, and other more easterly fields (Fig. 15c). The Nisku reservoirs are light grey dolomite with good vuggy and fracture porosity. East of the Leduc chain the upper part of the Nisku grades to anhydrite and the lower Nisku and underlying brown coral-stromatoporoid-rich dolomite of the Camrose form the reservoirs. Porosity is partly filled with anhydrite and distribution is irregular. The traps are partly stratigraphic, partly structural due to draping over the underlying Leduc reefs (Fig. IX-20). Pay zones vary to a

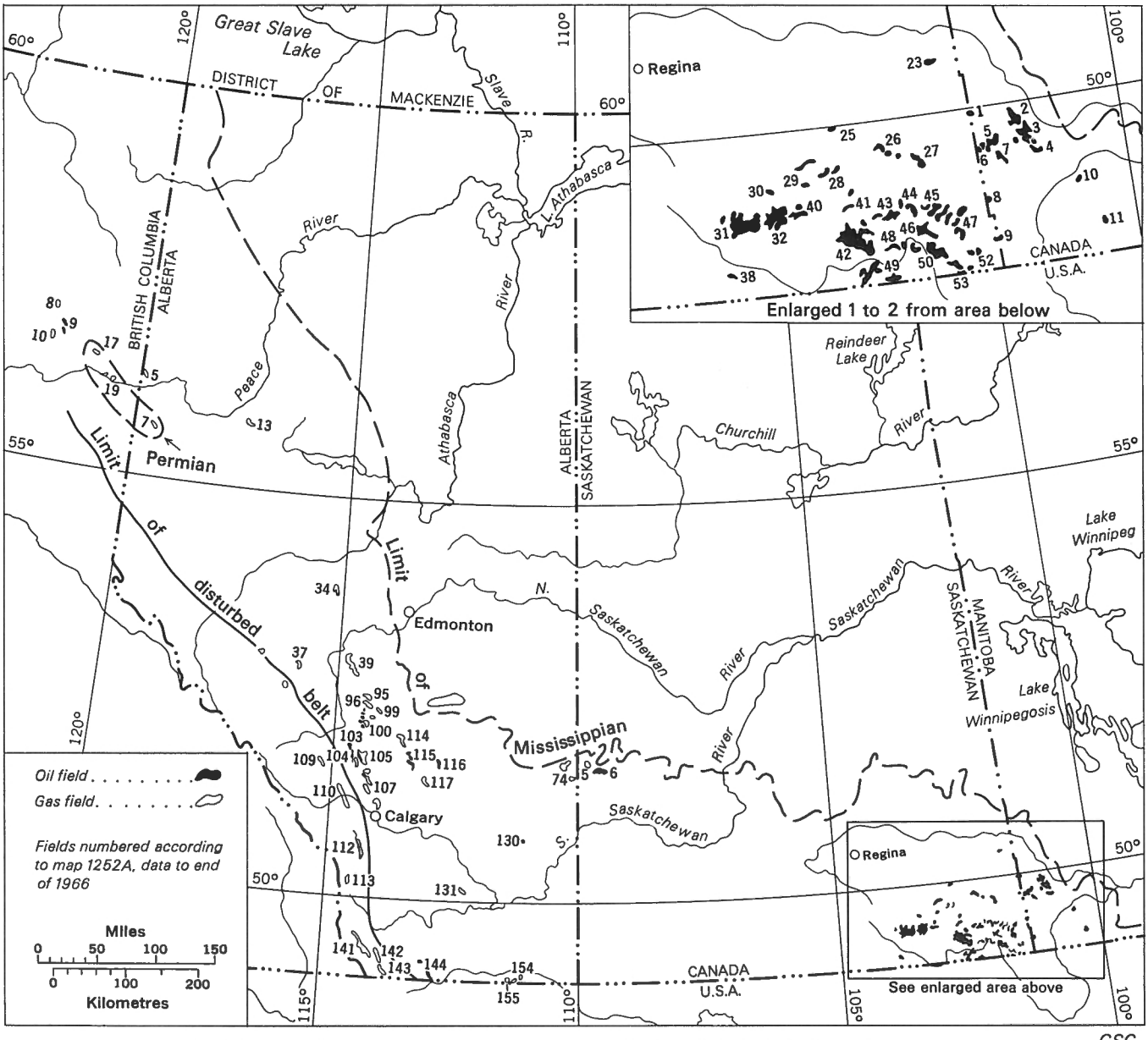


FIGURE IX-21. Mississippian and Permian oil and gas fields of Western Canada (by B. MacLean). GSC

maximum of 119 feet, porosity is generally from 3 to 5 per cent, and the gravity of the oil is 30 to 40 degrees A.P.I. In 1966, some 12,984,866 barrels of oil was produced from the Nisku and 36.5 million Mcf of gas.

Wabamun Group. Sour gas is produced from the Crossfield dolomite of the Wabamun Group at a number of fields that trend northwesterly between Okotoks (111) and Olds (106b) in Alberta (Fig. IX-15d). The gas occurs in the intercrystalline and pin-point porosity of fine to medium crystalline dolomites which grade eastward up-dip into non-permeable dolomite and anhydrite. A number of small fields and discoveries produce gas from the up-dip edge of the Wabamun truncated at the pre-Cretaceous surface. The hydrocarbons occur in crystalline sugary dolomites with good intercrystalline porosity and from vuggy brecciated dolomites and limestones. Porosity has been increased by leaching and weathering of the dolomite at the erosion surface. Gas with no associated oil or water occurs in porous vuggy dolomite in the lower part of the Wabamun at Simonette (26), Pine Creek (32a), and South Sturgeon Lake (21b). The gas is trapped in the dolomite against overlying fine-grained tightly cemented Wabamun limestone. In 1966, the Wabamun produced 68,451 barrels of oil and 76.3 million Mcf of gas.

Mississippian Fields

The Mississippian is the second most productive system in Western Canada, in 1966 yielding 74,186,392 barrels of oil and 484.5 million Mcf of gas, some 22.7 and 29.8 per cent respectively of the oil and gas produced in Western Canada. It is third in gas and oil reserves, accounting for 11 per cent of the oil and 27 per cent of the gas (Tables IX-3, -4). Productive Mississippian regions are, however, more extensive than Devonian (Figs. IX-1, -21).

Williston Basin

Oil of 27 to 38 degrees A.P.I. with very little gas is produced from Mississippian strata on the northeast flank of Williston Basin in southeastern Saskatchewan and southwestern Manitoba (Fig. IX-21). In 1966, 5,230,904 barrels were produced in Manitoba and 60,245,964 barrels in Saskatchewan. The oil is trapped in cyclically alternating porous and non-porous limestones, and some has accumulated where the porous beds are truncated at the pre-Jurassic or pre-Cretaceous erosion surface. As a result, the fields form a series of belts trending west to northwest, producing from successively higher strata down-dip to the southwest.

The Bakken sandstones produce a little 36-degree A.P.I. oil at Coleville–Smiley (6), Saskatchewan. The Souris Valley Formation produces in Manitoba in the Daly (5) and Virden (2, 3) areas where the porous Virden–Whitewater Lake crinoidal and oölitic limestones are truncated. The Nottingham field (47) produces 30- to 37-degree A.P.I. oil from the Alida-Tilston beds and the Parkman field (28) produces 34-degree A.P.I. oil from the Tilston and Souris Valley zones. The traps are topographic or structural highs at the erosion surface. Thirteen fields produce from the Frobisher–Alida beds, the largest being Hastings (46), Willmar (43), and Workman (52b), and ten from the Midale beds the largest being Steelman (42a), Weyburn (36b), and Midale (39). In these fields most traps are the result of porous car-

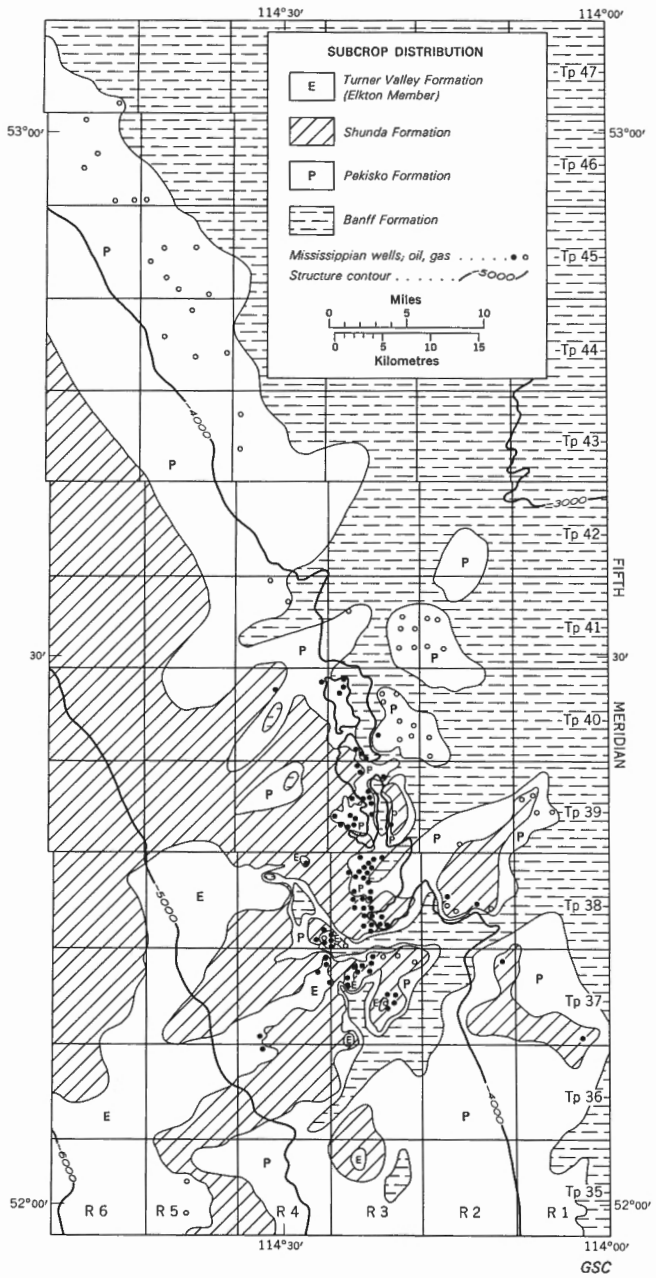


FIGURE IX-22. Mississippian oil and gas fields, Sylvan Lake–Minnehik trend, Alberta, showing paleogeology of the pre-Jurassic and pre-Albian erosion surface (by H. Martin).

bonates grading to non-porous carbonates or anhydrite, but in some the oil is trapped at the erosion surface.

Thomas (1962) and Fuzesy (1966) point out that production is largely controlled by porosity and permeability which in turn are dependent on the original constitution of the rock. The best reservoir rocks are oölite-pisolite, cryptocrystalline to microcrystalline limestones, encrinites, spergenites, algal oölitic-pisolitic limestones, and crinoidal limestone. All have less than 10 per cent matrix and good to excellent intergranular porosity. The presence of fine-grained limestone matrix decreases permeability, but dolomitization of the matrix to form sucrosic dolomite results in good intercrystalline porosity and a productive reservoir. Interconnecting vuggy porosity is developed in algal, oölitic, and pisolitic limestones where their matrix is leached and cavities enlarged by solution. Cavities in fossils account for a considerable proportion of the total porosity, lessened in places by infilling of the pore spaces by calcite, anhydrite, or silica.

Anhydrite and fine-grained, argillaceous, non-porous limestones underlie the reservoirs. The cap rock may be primary anhydrite with hard very finely crystalline argillaceous limestone, or result from secondary infilling of original porosity by anhydrite or chert at the post-Mississippian erosion surface, or be the argillaceous beds of the Jurassic Watrous Formation.

Alberta and British Columbia

Mississippian oil and gas produced in Alberta exclusive of that from the Foothills is from stratigraphic traps at the up-dip eroded edge of the Mississippian carbonates where deeply incised by pre-Mesozoic stream valleys (Fig. 22). Most production is from the Elkton Member of the Turner Valley Formation in west-central Alberta but the Pekisko and Banff Formations also produce (Fig. 23). The Elkton Member consists of buff and grey, fine to medium crystalline dolomite ghosts of corals and crinoids and with good vuggy and intercrystalline porosity. The dolomite grades laterally into and is interbedded with fine- to medium-grained crinoidal limestone with little or no porosity. The traps are formed partly by the up-dip change from porous dolomite to non-porous limestone and partly by truncation of the Elkton with sealing by impervious overlying beds. In places stratigraphic traps are modified by structural nosing along a north-northwest trend. Gravity of the oil ranges from 30 to 40 degrees A.P.I. Porosity ranges from 5 to 10 per cent and pay thickness from 40 to 55 feet in these fields. Fields along the Elkton trend from Harmattan (104) to Sundre (103) produce at depths of 9,000 feet to 10,000 feet. A few small gas fields occur where the Pekisko Formation is truncated at the erosion surface or where it occurs as outliers. Producing zones are dolomite with intercrystalline and leached vuggy porosity and incompletely cemented crinoidal limestones that grade laterally into tightly cemented limestones. Production from the Mississippian in Alberta in 1966, excluding Foothills fields, was 7,810,156 barrels of oil and 456.1 million Mcf of gas.

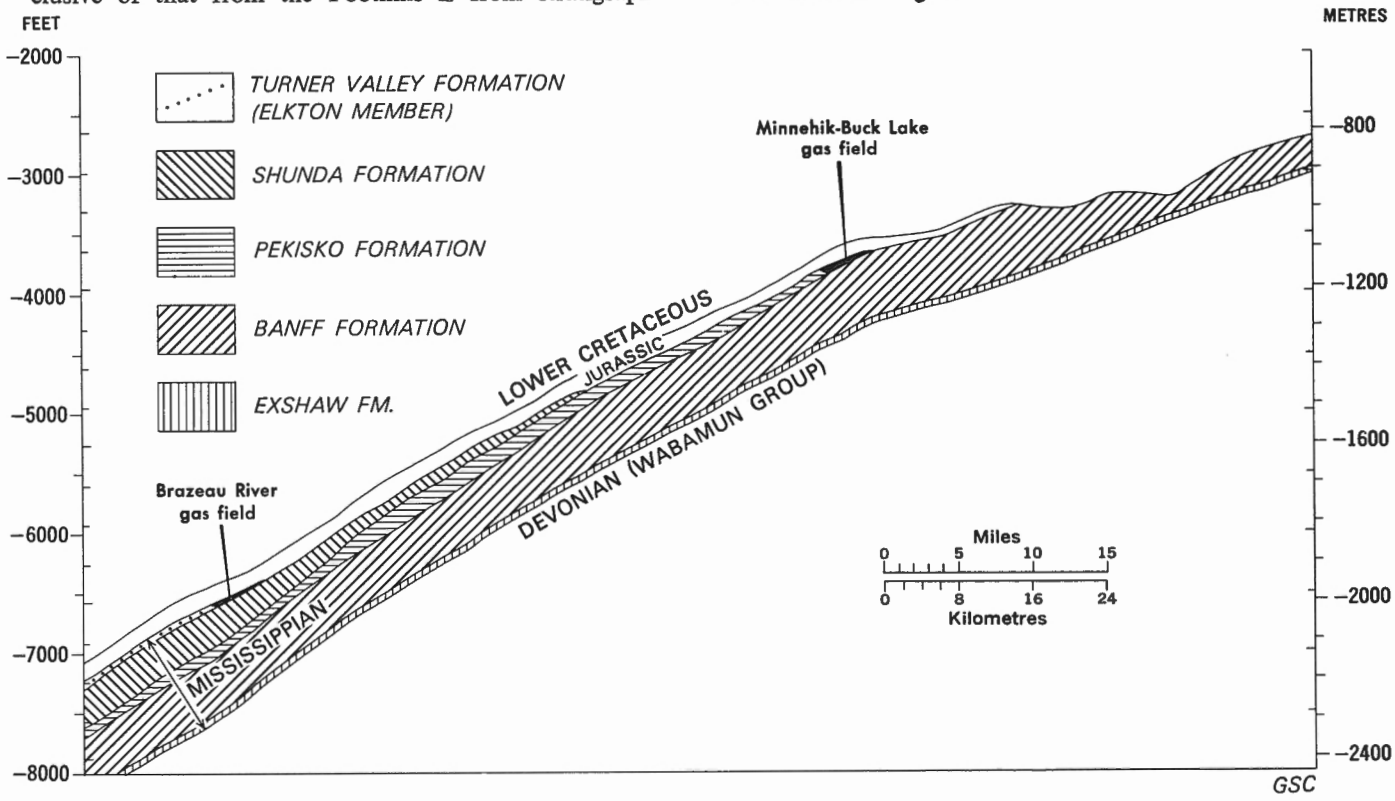
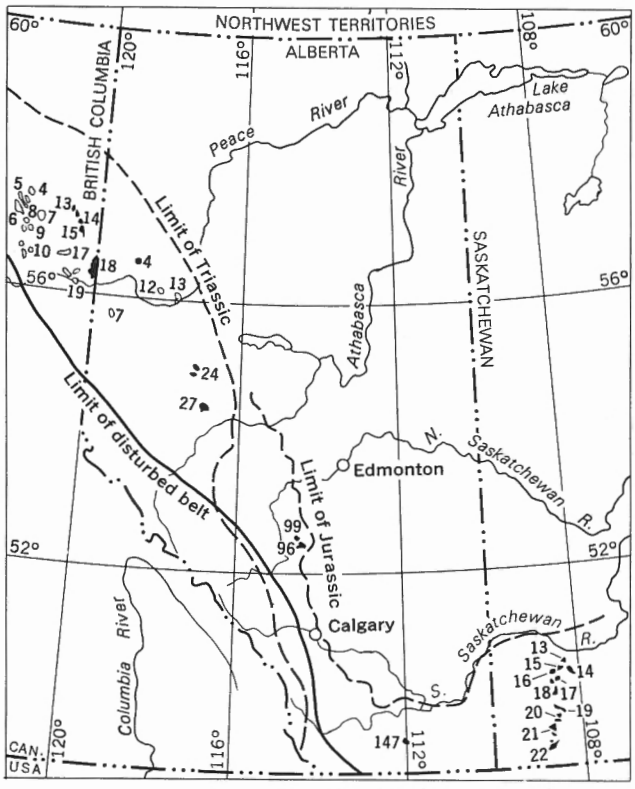
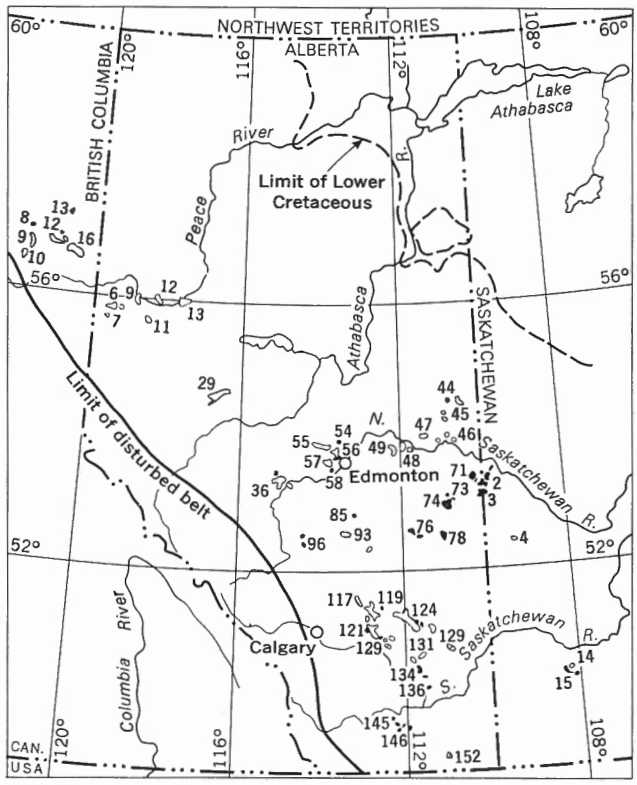


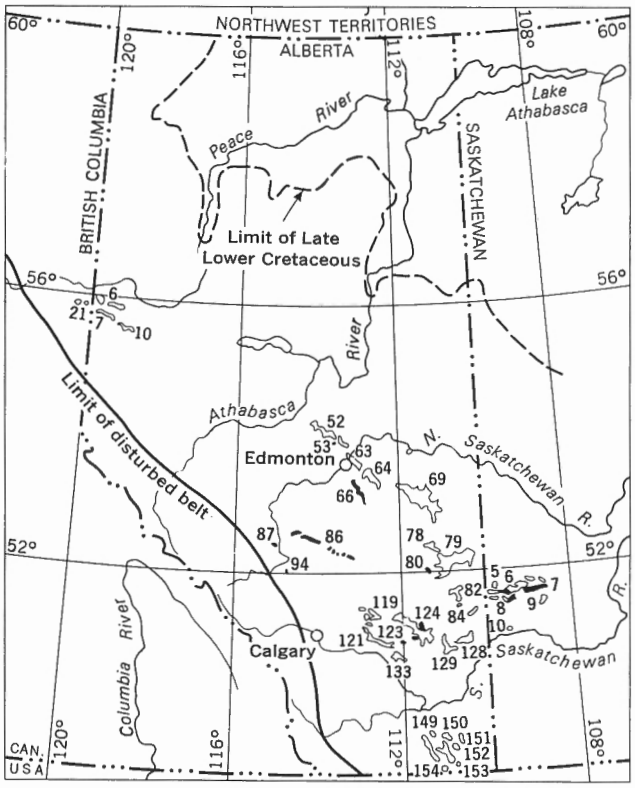
FIGURE IX-23. Cross-section through the Mississippian from Brazeau River field to Minnehik–Buck Lake field, Alberta (by H. Martin).



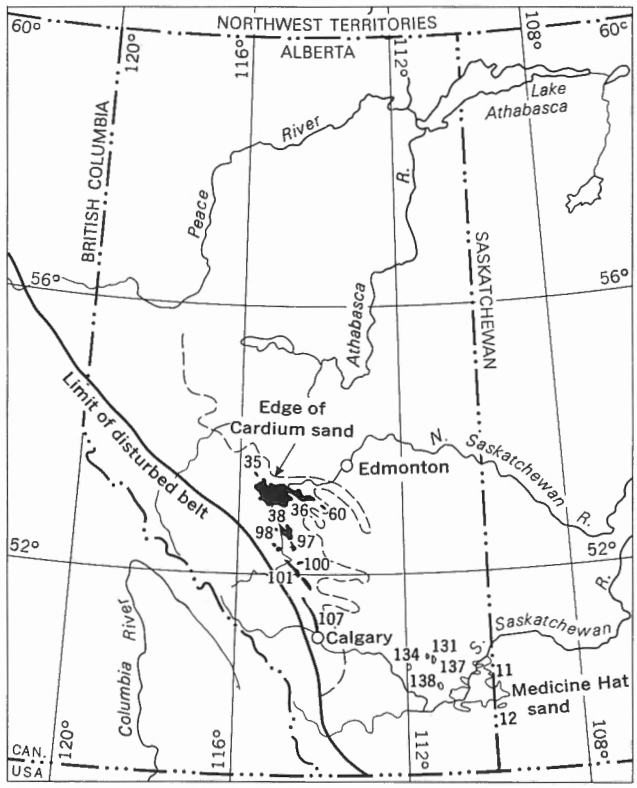
(a) Triassic and Jurassic oil and gas fields.



(b) Early Lower Cretaceous oil and gas fields; Mannville, and Blairmore producing horizons.



(c) Late Lower Cretaceous oil and gas fields; Viking, Cadotte and Bow Island producing horizons.



(d) Upper Cretaceous oil and gas fields; Cardium and Medicine Hat producing horizons.

GSC

Scale: 1 inch to 190 miles or 305 kilometres

FIGURE IX-24. Mesozoic oil and gas fields of Western Canada. Oil fields shown as solid areas, gas fields shown as open areas; fields numbered according to Map 1252A, data to end of 1965 (by B. MacLean).

In northeastern British Columbia the Blueberry oil field (9) and a number of gas fields produce from the Debolt Formation. The reservoir is generally brown, cherty, bioclastic limestone partly replaced by dolomite with intercrystalline and solution cavity porosity. The fields occur along the up-dip eroded margin of the Debolt and accumulation is partly the result of erosion and sealing of the beds at the pre-Mesozoic erosion surface and partly due to porosity pinchout from porous dolomite to limestone. Lenticular sandstones of the Kiskatinaw Formation produce gas at Boundary Lake South. Production from the Mississippian of British Columbia in 1966 was 899,368 barrels of oil and 3.8 million Mcf of gas.

Permian

The Belloy Formation produces gas at several small fields in British Columbia and Alberta (Fig. IX-21). The reservoir is a porous sandstone interbedded with dense limestones and cherty limestones. In 1966 production totalled 13.5 million Mcf.

Triassic Fields

Triassic sediments contain the greater part of the oil and gas reserves of British Columbia but only a small part of those of Alberta (Tables IX-3, -4). The system is productive in northeastern British Columbia and nearby

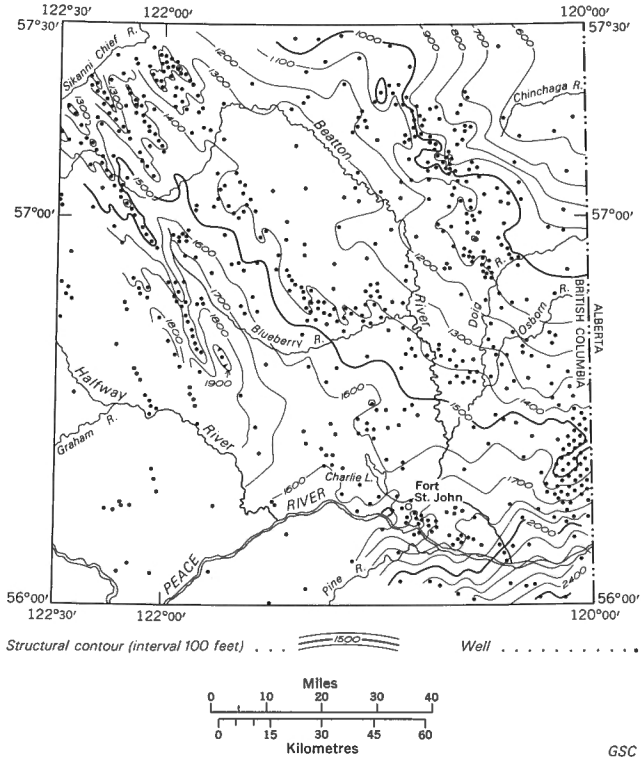


FIGURE IX-25. Structural contours on top of the Triassic system, Fort St. John area, British Columbia (by R. M. Procter).

Alberta (Fig. IX-24a). In 1966, 109 million Mcf of gas and 17,000,347 barrels of oil were produced.

In British Columbia the accumulations of oil and gas are controlled by a close interrelationship of folding, facies, and porosity, and pre-Albian topography. Most structural closures are on northwest-trending Laramide anticlines but some are on northeasterly flexures related to faults along the north side of the Devonian Peace River Arch (Fig. IX-25). Production is obtained at about 5,000 feet. Gas fields within the Baldonnel Formation (Fig. IX-26) include the Fort St. John (19), Nig Creek (7), Blueberry (9), Beg (6), Bubbles (5a), Jedney (5b), and and Laprise Creek (4) fields. The porosity occurs mainly as interstitial voids in an allochthonous limestone composed of fragmented pelecypods, brachiopods, and crinoids. In the Jedney and Laprise fields, the carbonate was deeply eroded in early Cretaceous time. Some gas is present in the sandstone of the Halfway Formation in the Fort St. John and Jedney fields. A major Triassic gas discovery was made in 1965 near Sukunka River in the Foothills west of Dawson Creek. In Alberta, gas is produced from dolomitic coquina and sandstone of the Montney Formation in the Tangent field and from the Baldonnel at Saddle Hills. Production has been obtained also at Kaybob South, Braeburn, Clear Hills, Fox Creek, Whitelaw, Snipe Lake, Mountain Park, and Nordegg River. Most oil produced is obtained from the Boundary Member of the Charlie Lake Formation at Boundary Lake field (17). The reservoir is similar to the Baldonnel, being composed of dolomitized skeletal limestone. Oil is also produced from Halfway sandstone at Beatton River (13a), Milligan Creek (13c), Wildmint (14), and Peejay (15c) fields. The oil in the Halfway is trapped stratigraphically by variations in porosity of the sandstone which formed as near-shore sand bars (Fig. IX-27). The gravity ranges from 26 degrees A.P.I. near the subcrop edge to nearly 50 degrees A.P.I. adjacent to the Foothills (Hitchon, 1964).

In Alberta, oil and gas are obtained from the Sturgeon Lake (24) area where individual sand lenses in the lower part of the succession are truncated at the post-Triassic erosional surface.

Jurassic Fields

Jurassic rocks contain only a small proportion of the oil and gas reserves of Western Canada, the producing fields occurring mostly in southern Alberta and Saskatchewan (Fig. IX-24a). Total production in 1966 was 1.8 million barrels of oil and 11.2 million Mcf of gas from Alberta, and 24.5 million barrels of oil and 7.2 million Mcf of gas from Saskatchewan.

In Alberta, the largest gas reserves are in the Nordegg Member in Medicine River (96), Gilby (95), and Paddle River (34) fields in central and west-central parts of the province. In the southeast, a small amount of gas is produced from the upper Shaunavon sandstone. Oil is

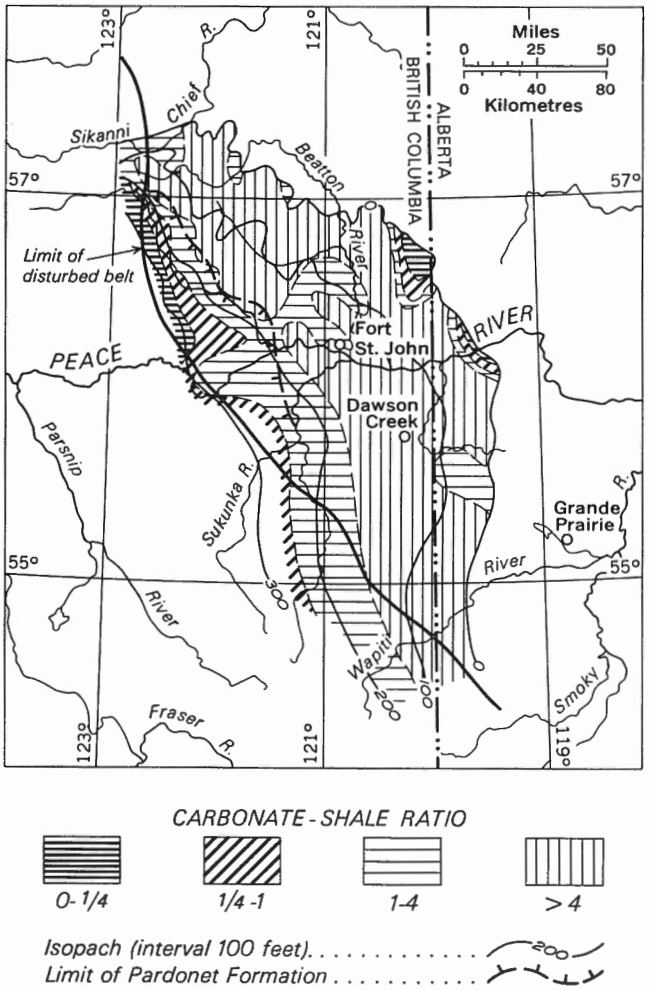


FIGURE IX-26. Isopachs and facies of the Triassic Pardonet and Baldonnel Formations, Fort St. John area, British Columbia (after Barss, et al.).

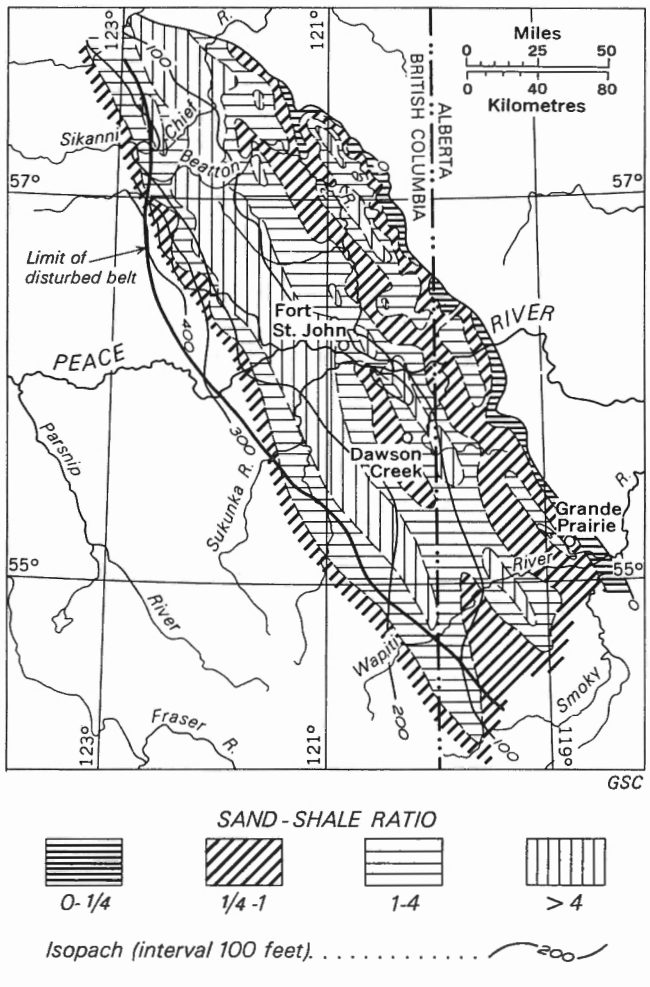


FIGURE IX-27. Isopachs and facies of Triassic Halfway Formation, Fort St. John area, British Columbia (after Barss, et al.).

produced from quartzose sandstone and belemnite conglomerate of the upper Sawtooth Formation at Skiff, Conrad (147), Glenphyll, and Burdett. The gravity of the oil is about 25 degrees A.P.I.

In Saskatchewan, oil of about 22 degrees A.P.I. is produced from dolomitic sandstones and bioclastic limestone of the Shaunavon Formation in a combination of stratigraphic and structural traps. Of these, the Rapdan (22), Dollard (21), Bone Creek–Instow (19) fields represent beach and channel-fill deposits, in part controlled by pre-existing topography (Christopher, 1964). The Gull Lake (18), Eastend (21b), and Leon Lake (20b) fields are developed on anticlines and the Leitchville–Shaunavon (20) field on the flank of a southeasterly plunging syncline. Several small fields find production in the middle Vanguard sandstone over the Swift Current Arch, a post-Jurassic structure in which the sandstone is truncated by pre-Albian erosion, trapping the oil stratigraphically.

The Wapella field (24) in southeastern Saskatchewan produces oil of 26 degrees A.P.I. gravity from fine-grained sandstone of Jurassic or Cretaceous age. In western Saskatchewan in the Fosterton field (15b), oil of 21 degrees A.P.I. is obtained from Roseray sandstones lying unconformably between rocks of known Jurassic and Cretaceous age.

Cretaceous Fields

Cretaceous rocks particularly those of Alberta (Fig. IX-24) contain a major portion of the oil and gas reserves of Canada (Tables IX-3, -4). In Alberta, Cretaceous sediments contain more than 13,000 million Mcf cubic feet of gas, about one third of the known marketable gas, and to end of 1966 have produced about one half of the marketed gas. Oil reserves in 1966, over 1,500 million barrels, account for about one quarter of the recoverable crude oil of Alberta. In 1966, production of natural gas from the Lower Cretaceous totalled 391 million Mcf or 24 per cent of that of Western Canada, and 97.8 million Mcf from the Upper Cretaceous, 6 per cent of the total. Oil production totalled 24.3 million barrels from the Lower

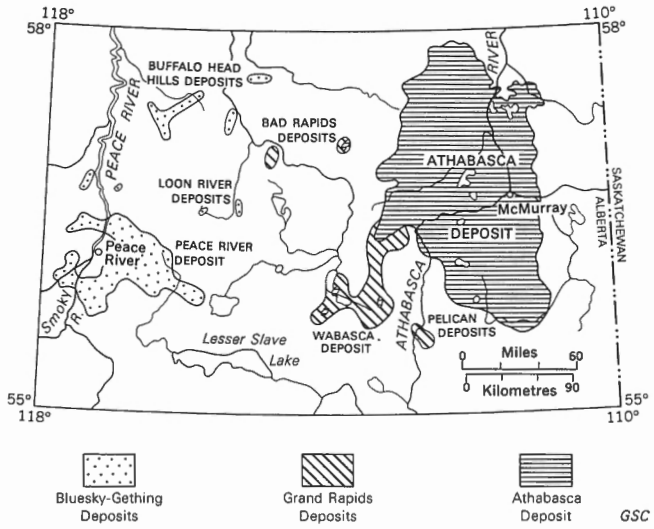


FIGURE IX-28. Oil sands of northern Alberta evaluated by the Oil and Gas Conservation Board, 1963 (after Pow, et al.).

Cretaceous in 1966, some 7.4 per cent of total from Western Canada, and 43.8 million barrels or 13.4 per cent from the Upper Cretaceous.

Athabasca Oil Sands

Of the known oil reservoirs in the world, the Athabasca oil sands of northeastern Alberta (Figs. IX-28, -29) is one of the largest, if not the largest. It is estimated (Pow, *et al.*, 1963) to contain more than 626 billion barrels of heavy oil in place, 369 billion barrels of recoverable raw oil-sands oil, and 266 billion barrels of recoverable up-graded synthetic crude oil. The evaluated sands extend over an area of 5¾ million acres and are well exposed along the banks of Athabasca River. The deposits were first reported by explorers as early as 1778, but only in recent years has their recovery been considered economically feasible by such methods as open-strip mining and injection of steam and aqueous solutions. The oil impregnation occurs in fluviatile, lacustrine, and lagoonal sands of the McMurray, Clearwater, and Grand Rapids Formations. The McMurray sandstones, containing the largest reserves, consist predominantly of quartz and are very lenticular. They lie on a highly irregular erosion surface carved on Devonian rocks in pre-Cretaceous time and are in part draped over some of the higher features.

Lower Cretaceous Fields

In the lower Mannville Formation (Fig. IX-24b) basal well-sorted quartzose sandstones were deposited as shoreline sediments or as channel-fills in part controlled by the topography of the underlying erosional surface. Structural traps are also found in these and higher sands of the lower Mannville where draped over Upper Devonian reefs as a result of differential compaction of the surrounding shale. Oil fields are numerous but small, the largest being Bellshill Lake (76a). Gas fields are also numerous, the largest being Kaybob (29), Princess (131), and Cessford (124a). Gas has been discovered also within the older Nikanassin sandstones south of Dawson Creek in British Columbia. In the upper Mannville, stratigraphic traps formed in part by intertonguing of sandstones and shales. Oil, having a gravity ranging from 10 to 18 degrees A.P.I. is found in numerous small fields near Lloydminster (71) and along the Alberta–Saskatchewan border.

Reservoirs in beds equivalent to the Lower Colorado are found in sandstone of the Peace River, Viking, and Bow Island Formations (Fig. IX-24c). Variation in the porosity of the lenticular Viking sands, related to winnowing and current action, together with up-dip loss of permeability, produce many stratigraphic traps. In general, the reservoirs are elongate sandbars, trending northwesterly in central Alberta. The Joarcam–Joseph Lake (66) trend extends for 30 miles, the oil and gas field being 2 to 4 miles wide. The Joffre–Gilby (94, 95) oil and gas field is 45 miles long and 1 to 1½ miles wide. The Viking gas reserves, including the largest fields of Viking-Kinsella (69), Provost (79), and Bindloss (128), are estimated at 9,200 million cubic feet, about 17 per

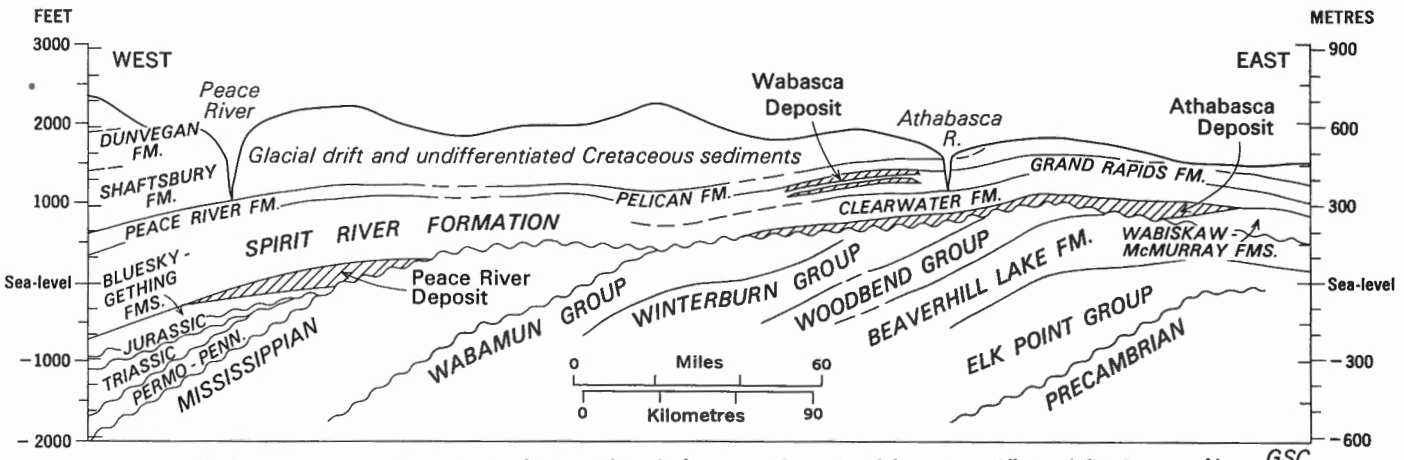


FIGURE IX-29. Schematic cross-section showing oil sands deposits and associated formations, Alberta (after Pow, et al.).

cent of total gas reserves of Western Canada. In Saskatchewan, the Viking produces oil at Coleville-Smiley (6) and gas at Brock (9). Basal Colorado sands are productive at Cessford (124a), Alberta. Gas is obtained from the Cadotte sandstone of the Peace River Formation in the Pouce Coupe field (6b) and other nearby fields of northern Alberta. The Cadotte is of near-shore origin and is associated with deltaic sediments. Gas accumulations are present in southern Alberta in the Bow Island sandstones at Pendant d'Oreille (152) and Foremost (149a) fields. The Bow Island Formation includes a succession of intertonguing sandstone and shale, the sandstone being of similar origin to the Viking. The former Bow Island field is now used for gas storage.

Upper Cretaceous

The Pembina (36b) field contains a considerable proportion of the recoverable oil reserves of Canada, being areally one of the largest known in North America and accounting for nearly 20 per cent of the oil reserves of Western Canada (Fig. IX-24d). The reservoir is found in barrier and epineritic sandstones of the Cardium Formation which has a regional dip to the southwest. The oil, of 37-degree A.P.I. gravity, is trapped by the up-dip change from sandstone to shale. A number of smaller Cardium oil fields with some gas, such as Crossfield (107), Garrington (101), and Willesden Green (38), extend southeasterly from Pembina.

Marine sandstones at the base of the Belly River Formation produce oil and gas in Pembina and Willesden Green fields. The sandstones are well sorted and are similar to the Cardium, having been formed as beaches and epineritic sand bars during a major regression of the Cretaceous seas.

About 5 per cent of Western Canada's recoverable gas reserves are present in the Medicine Hat sandstone in the combined Medicine Hat field (137) of southeastern Alberta and Hatton field (12) southwestern Saskatchewan, the largest single Cretaceous reserve of gas. Gas has been reported also from Dunvegan, Edmonton, Brazeau, and Wapiti strata.

Foothills Fields

In the late Cretaceous and Tertiary Laramide Orogeny, the eastern parts of Cordilleran Geosyncline were deformed. A series of folds and thrust sheets were produced which constitute the main control for entrapment of

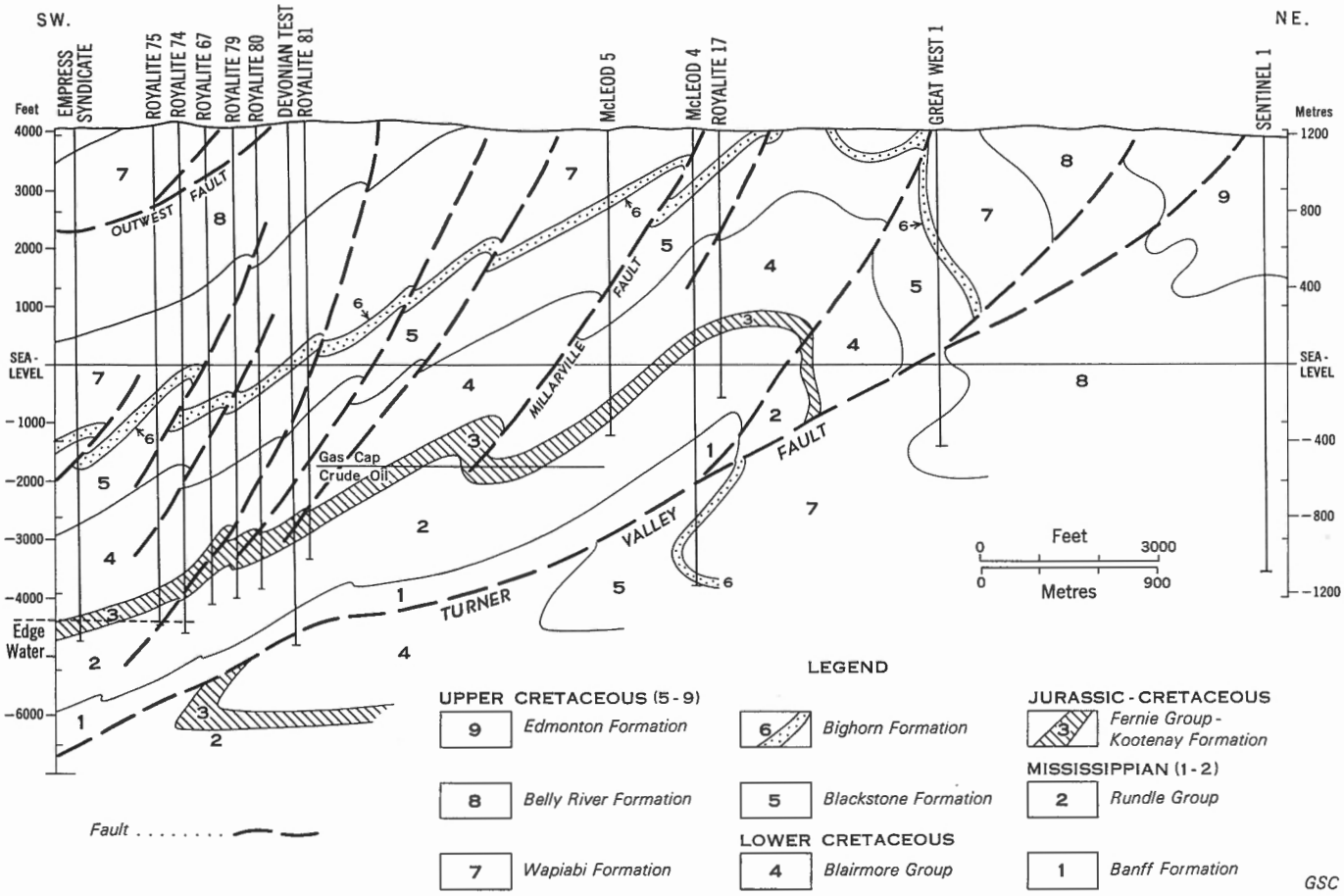


FIGURE IX-30. Structure-section through the central part of the Turner Valley oil and gas field, Alberta (after W. B. Gallup).

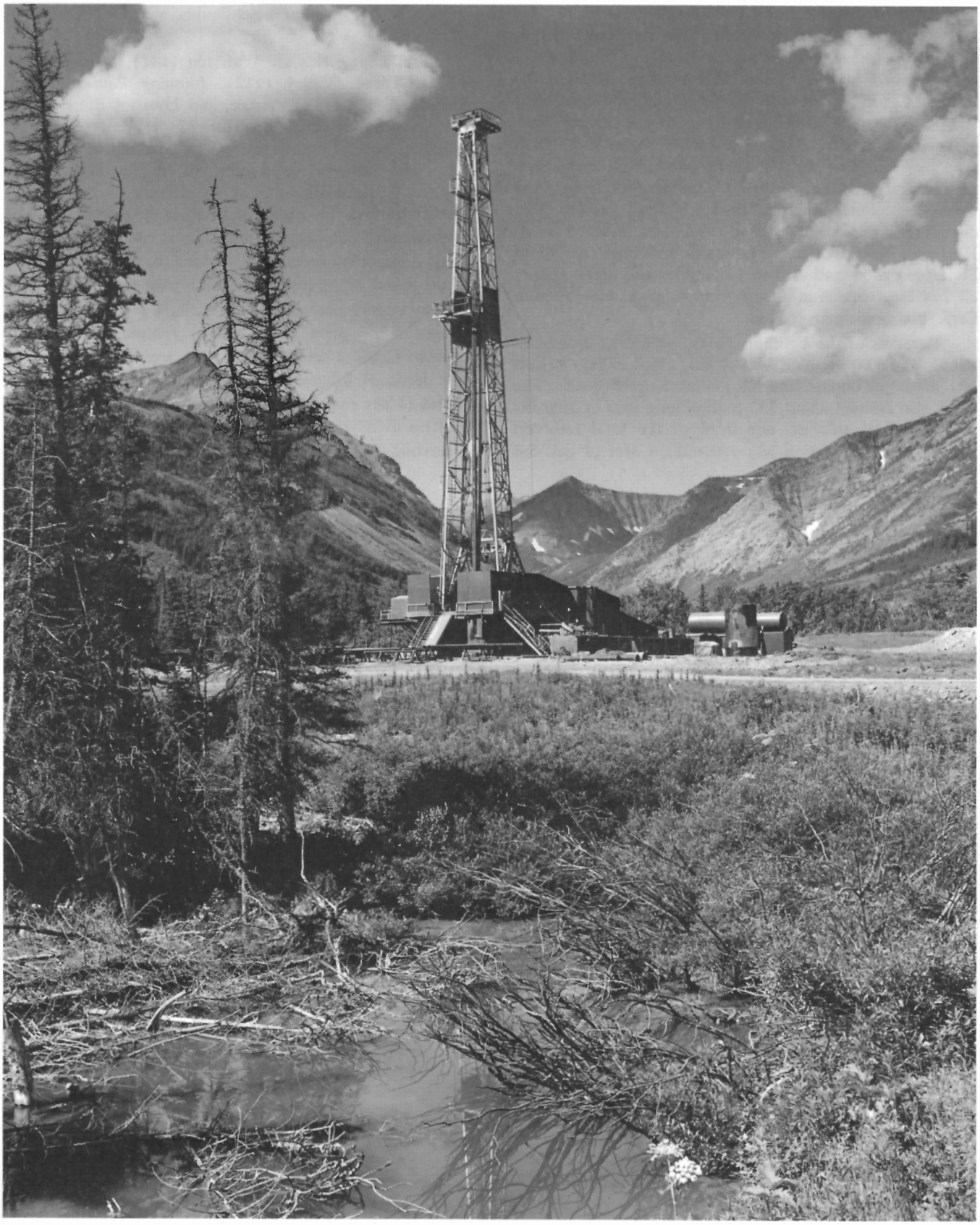


PLATE IX-5. Drilling rig, Shell Waterton #1 well.

hydrocarbons in the Eastern Cordillera (Figs. IX-30, -31, and -32). The natural gas fields of the southern Foothills of Alberta, which produce mainly from the Mississippian are described in this section. Some other fields in the Eastern Cordillera are only partly structurally controlled and have been described according to the age of their producing horizons. These are the Norman Wells field producing oil from the Middle Devonian Kee Scarp reef, and several fields in British Columbia producing from Triassic dolomites and sand bars. Natural gas has been discovered in Middle Devonian carbonates in the cores of Kotanelee and Pointed Mountain anticlines of Liard Plateau, Yukon and British Columbia, and light gravity oil has been found in Pennsylvanian sandstones beneath Eagle Plain, northern Yukon.

Reserves of the Foothills fields, established to end of 1966, were 5,000 million Mcf of gas, 116.2 million barrels of natural gas liquids, and 5.8 million barrels of oil. The fields also contain about 333 million long tons of recoverable elemental sulphur, one third of the total reserves. During 1966 production was 130 million Mcf of gas, 5.8 million barrels of natural gas liquids, and a million barrels of crude oil.

Producing fields are Waterton (141), Pincher Creek (142), Turner Valley (112), Jumping Pound (110a), Savanna Creek (113), Sarcee, and Lookout Butte (143), the seven fields accounting for 75 per cent of the gas reserves, most of the natural gas liquids, and all the crude oil. The principal reservoirs are in the Mississippian Turner Valley Formation of the Rundle Group. They consist of fine- to medium-crystalline dolomite with fine intergranular and vuggy porosity, or of coarsely crystalline, vuggy, crinoidal limestones. Much of the permeability is the result of fracturing in the brittle carbonates. The gas commonly contains dissolved light oil, especially the deeper fields, such as Waterton, Jumping Pound, and Pincher Creek where the reservoirs are 10,000 to 12,000 feet deep. Hydrogen sulphide content is generally between 5 and 15 per cent, but is 25 per cent in the Waterton field and up to 87 per cent in the Panther River wells. The reservoirs for the latter gas are in the upper Devonian

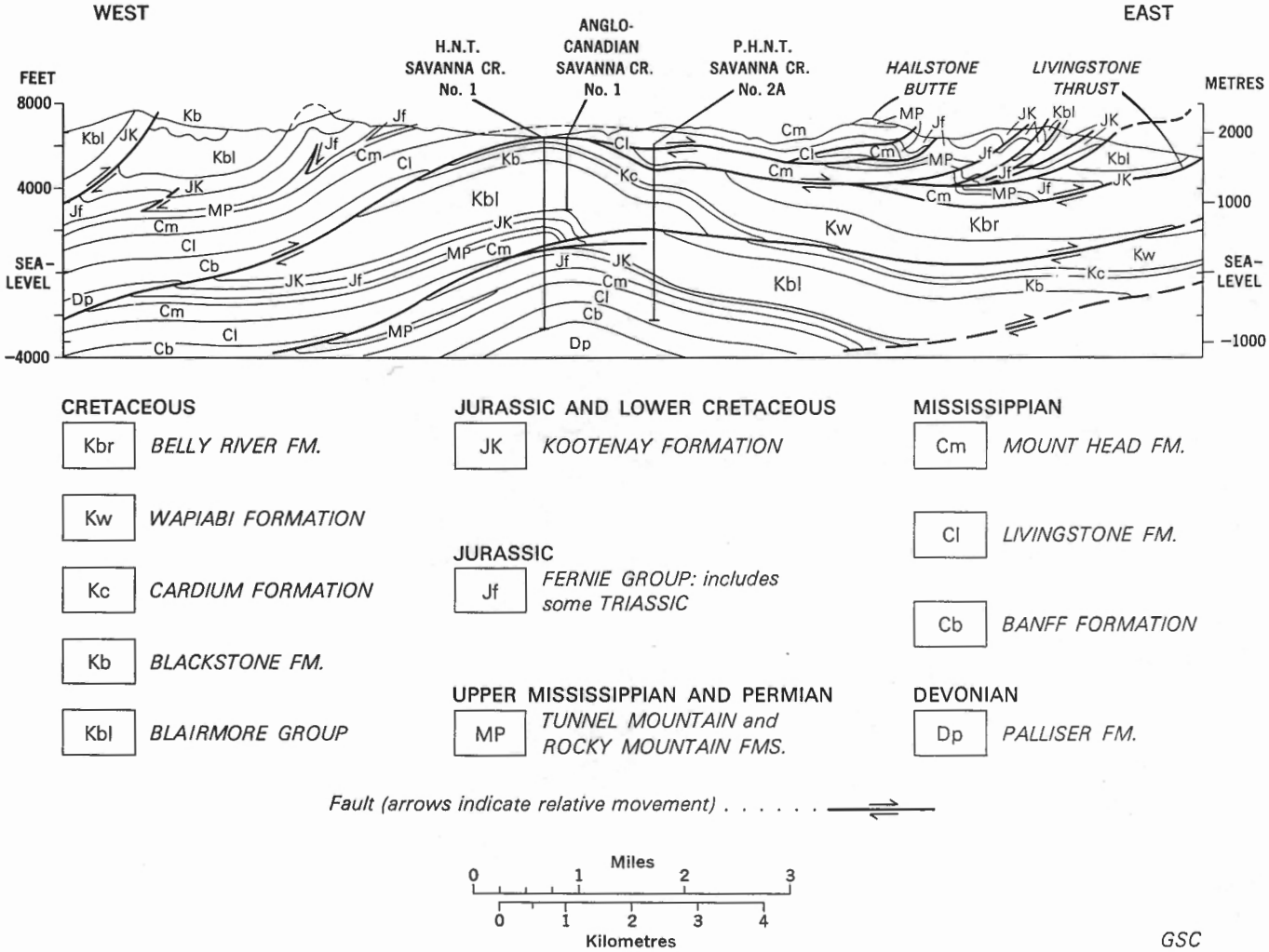


FIGURE IX-31. Structure-section through Savanna Creek gas field, Alberta (after Scott, et al.).

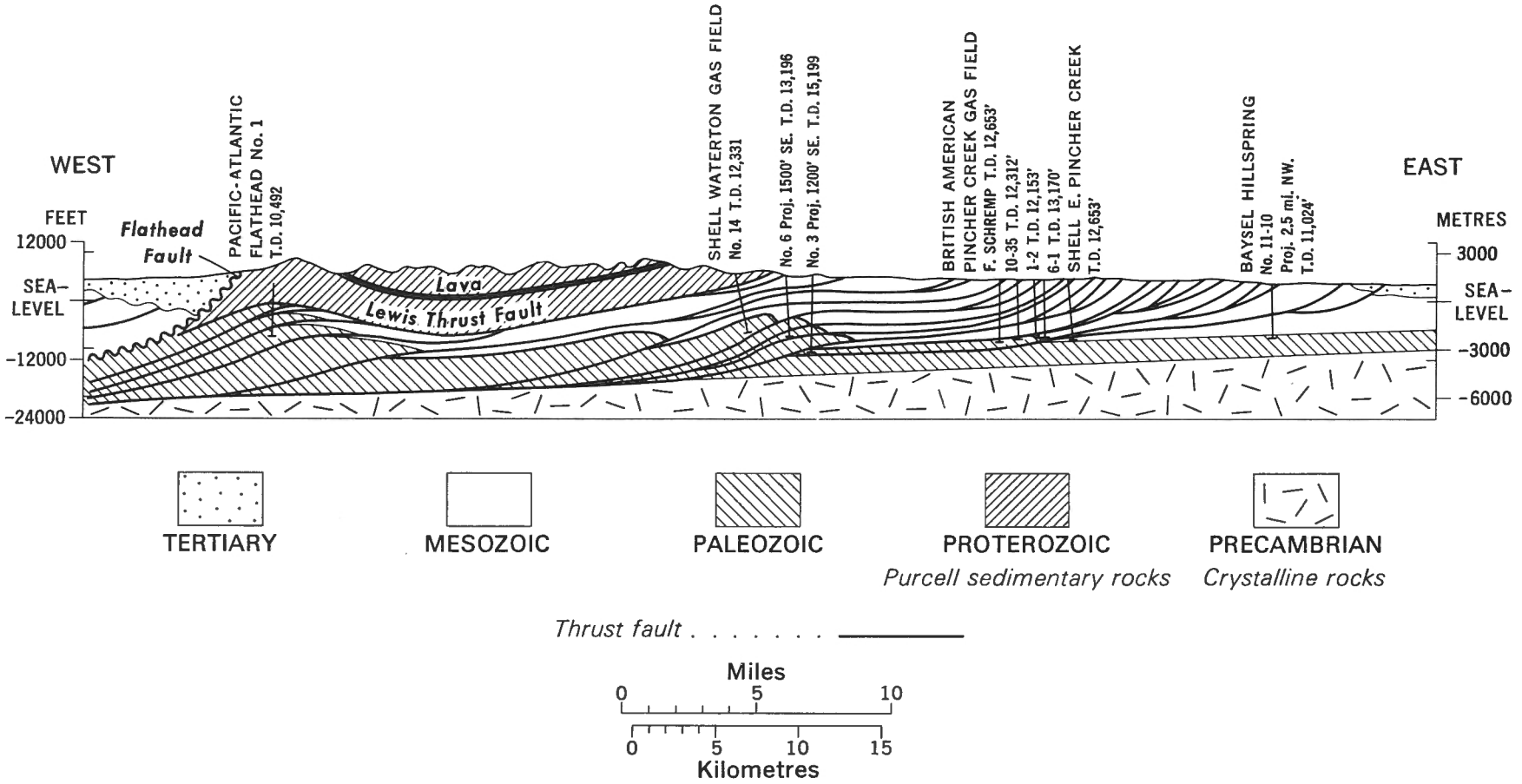


FIGURE IX-32. Structure-section through Waterton and Pincher Creek gas fields, Alberta (after Bally, et al.).

Nisku Formation and Wabamun Group; the remaining 13 per cent consists of inert gas, mainly nitrogen and a little methane. The Wabamun is productive of natural gas only in the Waterton field. In several small non-productive fields gas is encountered in Triassic dolomites as well as in the Mississippian. In the Turner Valley field some light oil is found in Lower Cretaceous sandstones but the bulk of the crude oil, 38 to 42 degrees A.P.I., is in the Mississippian.

The hydrocarbon accumulations occur where the Mississippian reservoirs are cut off by underlying thrusts, either in simple anticlines, west-dipping homoclines, or in multiple slices and more complicated folds. The surface structure may be cored with Mississippian strata as in the anticline at Turner Valley (Fig. IX-30), but in the other fields the reservoir lies beneath one or more thrust sheets (Figs. IX-31, -32). Oil and gas accumulations are considered to have been present in gently west dipping undisturbed Mississippian rocks prior to the Laramide Orogeny and to have migrated again into structurally high and salient parts of the thrust sheets during the orogeny.

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X. Geology of the Arctic Archipelago

Introduction 548

Geological Provinces 548

Precambrian 552

Lower and Middle Paleozoic 552

Upper Paleozoic and Mesozoic 568

Upper Paleozoic 569

Mesozoic 574

Cenozoic 584

Economic Geology 588

Selected References 589



INTRODUCTION

The geology of that part of the Arctic Archipelago that lies north of the Canadian Shield is dealt with in this chapter. The archipelago includes a land area of about 300,000 square miles underlain for the most part by rocks of Phanerozoic age. Precambrian rocks of other parts of the archipelago, particularly of Baffin, Somerset, and Victoria Islands, are treated in Chapter IV. Rocks of all Paleozoic and Mesozoic systems and of Cenozoic age are present. They are mainly sedimentary, but some plutonic and volcanic rocks are also found. Several intervals of Paleozoic orogeny affected parts of the area. The Mesozoic was not a time of orogenic activity, but in the Cenozoic some deformation occurred.

The area is divisible into seven major geological provinces (Fig. X-1): The *Arctic Lowlands,* formed of little disturbed, mainly lower Paleozoic rocks overlying the Precambrian basement; *Boothia Uplift,* a feature characterized by north-trending structures, mainly of Paleozoic age, with a Precambrian core exposed in the south dividing Arctic Lowlands into eastern and western parts; *Franklinian Geosyncline,* folded in the Paleozoic; *Sverdrup Basin,* which in late Paleozoic and Mesozoic time was superimposed on the folded Franklinian Geosyncline and was itself folded in the Cenozoic; *Prince Patrick Uplift,* an area of intermittent tectonic elevation in Mesozoic and Cenozoic time; *Arctic Coastal Plain,* composed of gently seaward dipping late Cenozoic strata, probably concealing the extensions of the Prince Patrick Uplift; and *Arctic Continental Shelf,* the seaward extension of Arctic Coastal Plain.

The structural provinces are described followed by a chronological account of the sedimentary, volcanic, plutonic, and tectonic events that have affected the area. Correlations of stratigraphic units and tectonic events are shown on Chart IV.

[1] R. Thorsteinsson prepared the account of the geology of the Paleozoic and Precambrian; E. T. Tozer prepared the parts treating the Mesozoic and Cenozoic. When acknowledging, reference should be made to individual author and page number.

GEOLOGICAL PROVINCES

Arctic Lowlands

The Arctic Lowlands include the terrains where the Precambrian basement is overlain by flat-lying or little disturbed sedimentary strata for the most part of Cambrian to Late Devonian age. In the western part, Cretaceous and early Cenozoic sediments form the surface. The lowlands represent the northern continuation of the Interior Plains of the continental mainland. They may also be



X

Geology of the Arctic Archipelago

R. Thorsteinsson and E. T. Tozer[1]

Triassic Blaa Mountain shales with gabbro sills on Raanes Peninsula, east side of Eureka Sound, Ellesmere Island.

regarded as the northern part of the North American Craton.

The lowlands are divided into segments by north- and northeasterly-trending uplifted belts and inliers of Precambrian rocks. The segments between the inliers appear to be basins that have experienced substantial subsidence. Aeromagnetic data suggest that the sedimentary thickness is more than 10,000 feet. Boothia Uplift, which bisects the lowlands, is so large and has experienced such a complex history that it is treated as a separate geological province. Foxe Basin represents the eastern part of the lowlands.

In the western lowlands a 250-mile-long inlier of Precambrian rocks flanked by gently inclined Paleozoic strata is known as the Minto Arch. This is essentially a northeasterly trending and plunging anticline with a structural relief of more than 3,000 feet. The strike parallels that of the exposed Precambrian rocks. Northeasterly trending gravity faults dislocate the Paleozoic rocks throughout the greater part; at the west end there is at least one north-trending fault. These faults indicate that the Minto Arch has characters of a horst as well as those of an arch.

Precambrian rocks forming inliers on southern Victoria Island have been termed the Wellington Arch. Adjacent Paleozoic rocks appear to be horizontal; the inliers seem to represent monadnocks on the basement surface rather than exposed parts of an area that has experienced uplift in Paleozoic or later times. On the other hand, the strike of Archean rocks at the head of Hadley Bay near the middle of Victoria Island coincides with those of the Slave Province of the Canadian Shield, which lies immediately south of the island. This suggests that much of southern Victoria Island may be underlain by an extension of the Slave Province, which has been resistant to subsidence since Precambrian time.

Boothia Uplift

This uplift is composed of Precambrian rocks flanked by a Cambrian to Lower Devonian sedimentary sequence and characterized by north-trending structures. Outliers of Middle and Upper Devonian and Cenozoic rocks are also present. Boothia Uplift as used here includes both the "Boothia Arch" and the "Cornwallis Fold Belt" of earlier accounts (Thorsteinsson, 1959; Kerr and Christie, 1965).

Both folding and faulting have contributed to the uplift. Most of the tectonic activity took place in the Devonian but faulting also occurred in the Cenozoic. The uplift strikes north, is 600 miles long, and about 125 miles wide at its widest point. Precambrian crystalline and sedimentary rocks in the southern two thirds form a conspicuous salient, penetrating Arctic Lowlands. Folded and faulted Paleozoic rocks flank the Precambrian core on Prince of Wales and Somerset Islands. The maximum structural relief expressed in terms of the level of the basement surface is about 25,000 feet. In Peel Sound the Precambrian rocks disappear beneath the Paleozoic formations. To the north, across Parry Channel, Boothia Uplift is expressed as a belt of northerly trending folds,

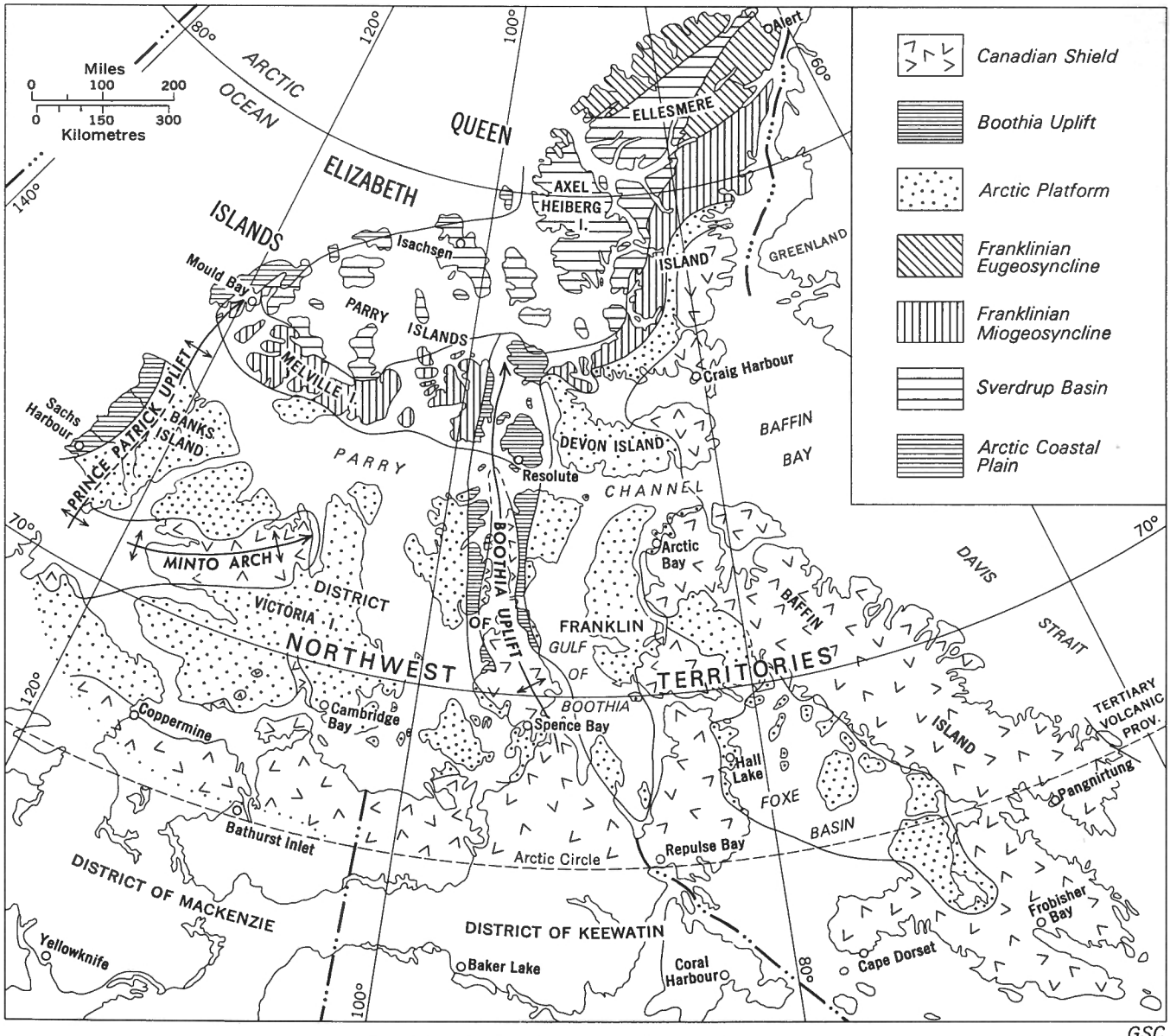

FIGURE X-1. Geological provinces of the Arctic Archipelago.

with linear grabens bounded by normal and transcurrent faults, forming a block interrupting the east–west folds of Franklinian Geosyncline. This northern part of the uplift is known as the Cornwallis Fold Belt, where the folds are characterized by generally steep flanked anticlines and broad open synclines. North of Grinnell Peninsula at Belcher Channel the Paleozoic rocks of the Boothia Uplift disappear beneath the younger formations of Sverdrup Basin, and a major anticlinal feature, the Cornwall anticline, lies on strike with the uplift.

Five unconformities within the environs of the uplift attest to episodes of deformation that range from Devonian to Cenozoic time. 1. The oldest is a Lower Devonian angular unconformity of regional extent. It is most precisely dated in eastern Bathurst Island where it is bracketed between the Bathurst Island and Stuart Bay Formations, both of Early Devonian age. Comparable unconformities occur within the Peel Sound Formation on Prince of Wales Island and within the Snowblind Bay Formation on Cornwallis Island. 2. An angular unconformity in eastern Bathurst Island separates the Lower Devonian Stuart Bay Formation from the early Middle Devonian, Disappointment Bay Formation. On Cornwallis Island, the latter formation unconformably overlies the Cornwallis Group and Cape Phillips Formation. 3. The Middle Devonian Bird Fiord Formation and the Upper Devonian Griper Bay Formation are separated by an angular unconformity on eastern Bathurst Island. 4. An episode of broad upwarping of the Boothia Uplift is indicated by a disconformity that separates Carboniferous and Lower Permian rocks along the northern extremity of the uplift. 5. Late Cretaceous or Cenozoic

volcanism and faulting have also affected the uplift on Bathurst and Cornwallis Islands.

Franklinian Geosyncline

This geosyncline was the site of more or less continuous, heavy sedimentation between late Precambrian and Late Devonian times. It is now the site of a complex fold belt that can be traced from Melville Island on the west to the central part of the archipelago, and thence northeasterly through Ellesmere Island to the northern part of Greenland—a distance of some 1,300 miles. Extensions of the fold belt probably underlie Sverdrup Basin and the Arctic Continental Shelf, extending southwest to join with the British Mountains of northern Yukon.

Rocks of Franklinian Geosyncline that border the Arctic Lowlands are miogeosynclinal in character; carbonates, quartzose sandstones, and shales are dominant facies. The total thickness of sedimentary rocks is about 40,000 feet. The northwesternmost exposures are of eugeosynclinal facies, mainly volcanic and volcanic-derived rocks, together with slate, chert, and impure limestone. The boundary between the miogeosyncline and eugeosyncline, as drawn on Figure X-1, is placed where the dominantly carbonate–shale–sandstone sequence is replaced by sections dominated by clastic rocks. Volcanic rocks are known only in the northern part of the eugeosyncline, adjacent to the Arctic Ocean.

The most extensive and profound tectonic event to affect the Franklinian Miogeosyncline was the Ellesmerian Orogeny of Late Devonian or Early Carboniferous age which produced the Parry Islands, Central Ellesmere, and Northern Ellesmere Fold Belts. The age of the clastic wedge deposits in the miogeosyncline suggests that the eugeosyncline was affected by orogeny and uplift prior to the Middle Devonian, and no doubt the eugeosyncline was folded during the Ellesmerian Orogeny as well.

Sverdrup Basin

Sverdrup Basin is a regional depression, superimposed on the Franklinian Geosyncline. It contains an essentially concordant succession of formations ranging in age from Lower Carboniferous (Viséan) to Upper Cretaceous. In places early Cenozoic beds form the youngest part of the concordant sequence; elsewhere they unconformably overlie the older beds. The relations of these beds show that in Cenozoic times Sverdrup Basin had ceased to be a simple, continuously subsiding area. Partly during but mainly after the deposition of the Cenozoic beds the basin experienced Eurekan Orogeny, which involved folding, thrust faulting, and the emplacement of piercement bodies composed of Carboniferous evaporites.

Originally a simple crustal depression, as a result of Eurekan Orogeny the basin today is a great synclinorium, the axis of which extends northeast for 600 miles from northern Melville Island to northern Ellesmere Island. Apparently this structural axis was also the depositional axis, for it more or less marks the site of heaviest sedimentation for much of the history of the basin.

The maximum thickness of sediment in any one part of the basin is about 35,000 feet; the aggregate thickness of all the formations is more than 50,000 feet, but it is unlikely that this amount accumulated in any one area. The rocks are mainly sedimentary, but upper Paleozoic and Cretaceous basalts are present and thick gabbroic sills intrude much of the sequence. In the axial area, as time progressed the depocentres migrated westward: from Ellesmere Island in the upper Paleozoic, to eastern Axel Heiberg Island in the Triassic, to western Axel Heiberg Island in the Cretaceous. On the margins of the basin the sections are much thinner, partly through thinning of individual formations, and partly through the disappearance of beds.

From the thickness and lithology of the rock sequences there is no doubt that present south and east margins of the basin (Fig. X-1) lie close to the original limits. The northwestern edge is concealed by late Cenozoic beds of the Arctic Coastal Plain. Adjacent to the plain the basin exposures show that the regional dip is to the southeast, and the beds thicken down-dip. Rocks of shallow-water facies are found at some levels, suggesting that an intermittently elevated sill separated Sverdrup Basin from the Arctic Ocean during the upper Paleozoic and Mesozoic. Prince Patrick Uplift is probably an exposed remnant of this sill.

Prince Patrick Uplift

On the west side of the archipelago, adjacent to the coast, a series of inliers indicate uplifted areas. They include the Nelson Head Proterozoic inlier and Cape Crozier anticline, respectively forming the north and south capes of Banks Island, and the area of Devonian rocks that forms much of southern Prince Patrick Island. These inliers are characterized by north-trending faults, mostly of Cenozoic age. They are on strike with one another and are presumably related structural features. Collectively termed the Prince Patrick Uplift they probably represent culminations on a largely buried structural elevation. A smaller, complementary feature is the Eglinton Graben.

In Paleozoic time the east boundary of the uplift may have been the locus of dextral strike-slip faults. This implies that during the Paleozoic the uplift lay south of its present position and would account for the absence of the Ellesmerian structures of Parry Islands Fold Belt. These are well developed on western Melville Island, 40 miles to the east. The Devonian rocks of Prince Patrick Island lie directly on strike but are devoid of similar structures. The fold belt may end abruptly in the vicinity of the Eglinton Graben; alternatively it is truncated by faults. An abrupt ending for the Parry Islands Fold Belt is difficult to reconcile with the widely held theory that the Franklinian Orogen is related to the Paleozoic orogenic belt of northern Yukon. This justifies favouring the

alternative that the fold belt is truncated by a strike-slip fault, or faults. The locus of the Devonian inlier of Cape M'Clure, with structural features related to those of Prince Patrick Island rather than to those of the Parry Islands Fold Belt, renders it unlikely that the folds are displaced to the south by transcurrent movement. This leaves displacement to the north, or dextral movement, as the only alternative. According to this interpretation, the Parry Islands Fold Belt is truncated by a dextral fault and displaced to the north.

In late Paleozoic and Mesozoic time Prince Patrick Uplift appears to have been resistant to subsidence and as such defined the western boundary of heavy sedimentation in Sverdrup Basin. There is no evidence that wrench faulting continued into the Mesozoic, but vertical movements probably occurred. The local unconformity between the Wilkie Point and Mould Bay Formations suggests that movement occurred in the Upper Jurassic. Movement also took place in the early Cenozoic and again in the late Cenozoic or Pleistocene. Parts of Prince Patrick Island are seismically active suggesting that movement on the faults is continuing.

Arctic Coastal Plain

Arctic Coastal Plain comprises a narrow strip of little disturbed sediments that extends the coast of the Arctic Ocean from Banks Island to Meighen Island. The sediments are Pliocene or Pleistocene, unconformably blanketing the rocks of Prince Patrick Uplift and Sverdrup Basin. In general these late Cenozoic beds appear to dip gently northwest, towards the sea, but on Ellef Ringnes Island the dip is apparently to the northeast. On Prince Patrick Island the beds have been dislocated by faults of small displacement. The southern faults, near the exposed part of the Prince Patrick Uplift, strike north as do those of the uplift. Farther north the faults strike northeast.

Northwest of the Arctic Coastal Plain the *Arctic Continental Shelf* is about 100 miles wide. The continental slope is also about the same width, and lies between 600 and 3,000 metres depth and the Canada Basin. Analysis of the magnetic properties of Canada Basin (King, *et al.*, 1966) suggests that it is underlain by rocks similar to those of the Canadian Shield and not by typical oceanic crust.

PRECAMBRIAN

Late Proterozoic sedimentary rocks are brought to the surface in parts of Franklinian Miogeosyncline, and metamorphosed rocks, probably of Proterozoic age, occur in the eugeosyncline. In northeastern Ellesmere Island, Hadrynian dolomite, shale, and sandstone disconformably underlie Lower Cambrian strata (Fig. X-2). At Ella Bay it is more than 5,600 feet thick, but towards the south it is much thinner, and near Bache Peninsula is absent, the Cambrian resting directly upon basement gneisses (Christie, 1967).

In northern Ellesmere and Axel Heiberg Islands, metamorphic rocks of uncertain age are exposed. The Rens Fiord Complex of northern Axel Heiberg Island (Fricker and Trettin, 1962; Trettin, 1964) includes slate, phyllite, sandstone, chert, dolomite, greenschist, spilitic volcanic flows, quartzite, hornfels, and marble. The Cape Columbia Group of northern Ellesmere Island (Blackadar, 1954; Christie, 1964) is composed of a variety of gneissic and schistose rocks, equivalents of which are unconformably overlain by Ordovician Wilderness strata. Apart from stromatolites collected in the Rens Fiord Complex, neither that complex nor the Cape Columbia Group has yielded fossils. A K–Ar radiometric age of 550 ± 35 m.y. has been obtained from gneiss of the Cape Columbia Group and 535 ± 35 m.y. from a schistose sandstone of the Rens Fiord Complex. These ages suggest a period of metamorphism in Cambrian or Hadrynian time. The rocks themselves are presumably Cambrian or older.

LOWER AND MIDDLE PALEOZOIC

Cambrian

In the Arctic Lowlands, Lower and Middle Cambrian rocks are discontinuously exposed along the contact with the Canadian Shield (Pl. X-1); in the Franklinian Geosyncline they are represented only in the miogeosynclinal belt of Ellesmere Island. In both provinces the Cambrian rocks are separated from the Precambrian by a regional unconformity. Rocks of Upper Cambrian age are unknown in the archipelago, and a regional disconformity separates Cambrian and Lower Ordovician rocks.

Arctic Lowlands and Franklinian Miogeosyncline

Lower and Middle Cambrian rocks of eastern Devon Island aggregate 900 feet thick (Kurtz, *et al.*, 1952). The Lower Cambrian Rabbit Point Formation consists of 20 to 30 feet of sandstone, variably argillaceous, arkosic and quartzitic, and containing *Olenellus*. It is separated from the overlying Middle Cambrian Bear Point Formation by a sharp yet conformable contact that may be a disconformity. The Bear Point contains *Glossopleura*, and consists mainly of dolomite, intraformational con-

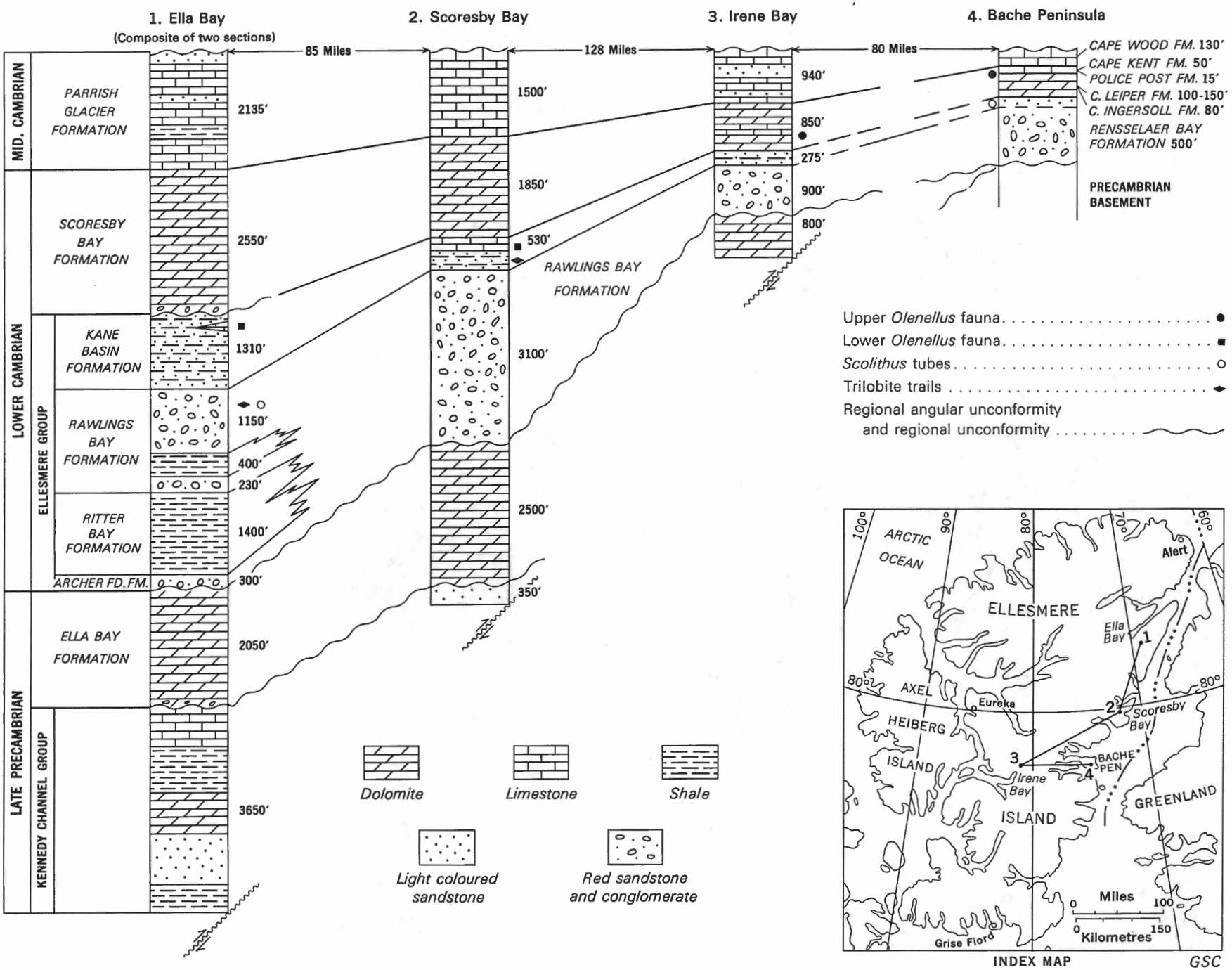

FIGURE X-2. Columnar sections, Proterozoic and Cambrian rocks, central-eastern Ellesmere Island (by J. W. Kerr).



PLATE X-1. Lower Paleozoic strata overlying Precambrian crystalline rocks, south coast of Devon Island.

glomerate, and minor dolomitic limestone up to 520 feet thick. The Ooyahgah Formation with Middle Cambrian brachiopods is composed of relatively soft weathering dolomite and minor intraformational conglomerate, 210 to 400 feet thick.

The Middle Cambrian of Boothia Peninsula and Somerset Island consists of about 300 feet of weakly cemented, brownish weathering sandstone, interbedded with dolomite that is variably shaly and quartzose, and intraformational breccia and conglomerate (Blackadar and Christie, 1963). It rests nonconformably on Precambrian crystalline rocks and is unconformably overlain by Ordovician carbonates.

The basal Paleozoic beds on Victoria Island are red and green sandstone, with minor shale, siltstone, and dolomite, up to 400 feet thick, unconformably overlying various Proterozoic formations (Thorsteinsson and Tozer, 1962). Lower Cambrian olenellids have recently been discovered in these rocks.

Two formations tentatively assigned to the Cambrian mark the base of the Paleozoic succession of northwestern Baffin Island (Trettin, 1965). The Gallery Formation was deposited in a southeasterly trending basin and rests with regional unconformity on various formations of Proterozoic age. It includes reddish sandstone with minor siltstone, conglomerate, breccia, and shale.

The greatest known thickness (1,125 feet) is near Arctic Bay; to the northeast and south it thins to zero. The attitude of crossbedding indicates that the sediment came from these directions. A sharp contact, probably representing a minor disconformity, defines the base of the overlying Turner Cliffs Formation, composed of dolomite, quartzose and dolomitic sandstone, intraformational conglomerate, and minor siltstone and shale. The maximum thickness of 1,000 feet is attained in Navy Board Inlet and, like the Gallery, the Turner Cliffs Formation thins to the north and disappears to the south. The depocentres of the Gallery lie to the west of that of the Turner Cliffs Formation. The Gallery lacks diagnostic fossils; *Lingulella* has been obtained from the Turner Cliffs. The exact age of these beds is uncertain but they are older than mid-Ordovician.

At Bache Peninsula, Ellesmere Island, Precambrian crystalline rocks are overlain by the Rensselaer Bay Formation comprising about 500 feet of dominantly clastic rocks divisible into three members (Christie, 1967). The lower is sandstone and stromatolitic dolomite, intruded by diabase sills; the middle is unfossiliferous arkose and conglomerate; and the upper is a sandstone with *Scolithus*-like tubes. Both the middle and upper members overlap westward and at the east entrance to Sverdrup Pass the upper member rests directly on the crystalline rocks. The Rensselaer Bay beds are succeeded by five formations, cumulatively 450 feet thick, dominantly carbonate. The Cape Leiper and Cape Ingersoll are unfossiliferous dolomites overlain by 15 feet of Police Post limestone with *Olenellus* and other Lower Cambrian trilobites. The succeeding Cape Kent limestone has not yielded fossils on Bache Peninsula but is correlated with the type Cape Kent in Greenland, which contains *Olenellus*. The highest Cambrian beds, the Cape Wood Formation, composed of dolomite, sandstone, and limestone, have yielded Middle Cambrian *Glossopleura* (Troelsen, 1950). Opinions differ regarding the position of the base of the Cambrian in the Bache Peninsula succession. Troelsen regarded all the beds below the Police Post limestone as "Eo-cambrian," and correlated them with the Thule Group of Inglefield Land, on the Greenland side of Kane Basin. Kerr (1967) suggests that the transgressive *Scolithus*-bearing beds at the top of the Rensselaer Bay are Lower Cambrian (Fig. X-2). Christie (1967) accepts the Cambrian dating for the *Scolithus* beds, and suggests that the two underlying members of the Rensselaer Bay Formation, which thicken to the east and disappear to the west, are possibly Precambrian.

Thick Lower and Middle Cambrian miogeosynclinal deposits occur northwest and northeast of Bache Peninsula in northeastern Ellesmere Island (Kerr, 1967). Figure X-2 shows the salient characters of three Proterozoic and Cambrian sections studied by Kerr, and his correlations with the thin Lower and Middle Cambrian platform deposits of Bache Peninsula.

Ordovician, Silurian, and Lower Devonian

Attention has already been drawn to the widespread disconformity between Middle Cambrian and early Ordovician strata. The next younger widespread Paleozoic unconformity in Arctic Archipelago occurs between Lower and Middle Devonian rocks. Ordovician to Early Devonian time is marked by general tectonic quiescence and more or less continuous sedimentation in the Arctic Lowlands, and the Franklinian Miogeosyncline (Fig. X-3). In these regions neither the Ordovician–Silurian boundary nor the Silurian–Devonian boundary is marked by unconformities.

On the Arctic Lowlands the rocks are mainly carbonates, with some evaporites. In Franklinian Miogeosyncline two principal facies were deposited: carbonates and evaporites, similar to those of the lowlands, but thicker; and graptolitic shales and argillaceous limestone. Commonly the graptolitic beds overlie the carbonates but on northwest Melville Island the section is almost entirely of graptolitic facies. On Melville and Ellesmere Islands there are thick carbonates which probably formed as banks within the troughs that accommodated the graptolitic sediment.

In Franklinian Eugeosyncline a great variety of rocks is present, mainly fine clastic rocks, some graptolitic, volcanic rocks, and locally carbonates. A very thick succession of clastic rocks, the Cape Rawson Formation, perhaps as old as Late Ordovician and certainly Silurian and Early Devonian, is transitional between the typical miogeosynclinal and eugeosynclinal sequences (Fig. X-1).

In the eugeosyncline orogenic activity probably started in the Silurian. On the Boothia Uplift tectonic elevation started in the Early Devonian. Adjacent to these areas the Silurian and Lower Devonian rocks include a variety of clastic sediments, both marine sandstone and beds of "Old Red Sandstone" facies.

Lower Ordovician, Canadian

Carbonate rocks of Canadian age are not widely exposed. On the Arctic Lowlands they are known from Devon and Ellesmere Islands; on Boothia Uplift from Cornwallis Island; and within the miogeosyncline, from Ellesmere and Melville Islands. In the southern lowlands their presence has been proved only by an isolated occurrence of limestone with shaly interbeds, containing Arenigian graptolites, on Jens Munk Island in Foxe Basin (Blackadar, 1963).

On Devon Island rocks of Canadian age disconformably overlie Middle Cambrian strata (Kurtz, *et al.*, 1952). The Mingo River Formation is 260 feet thick, predominantly limestone in the east but grades to dolomite in the west, with minor sandstone, intraformational conglomerate, and chert. The overlying Nadlo Point Formation comprises 590 feet of limestone and dolomitic limestone. Both formations are dated by the presence of *Hystricurus*.

The Canadian on Bache Peninsula is represented by three formations: the Cass Fiord, about 1,500 feet of

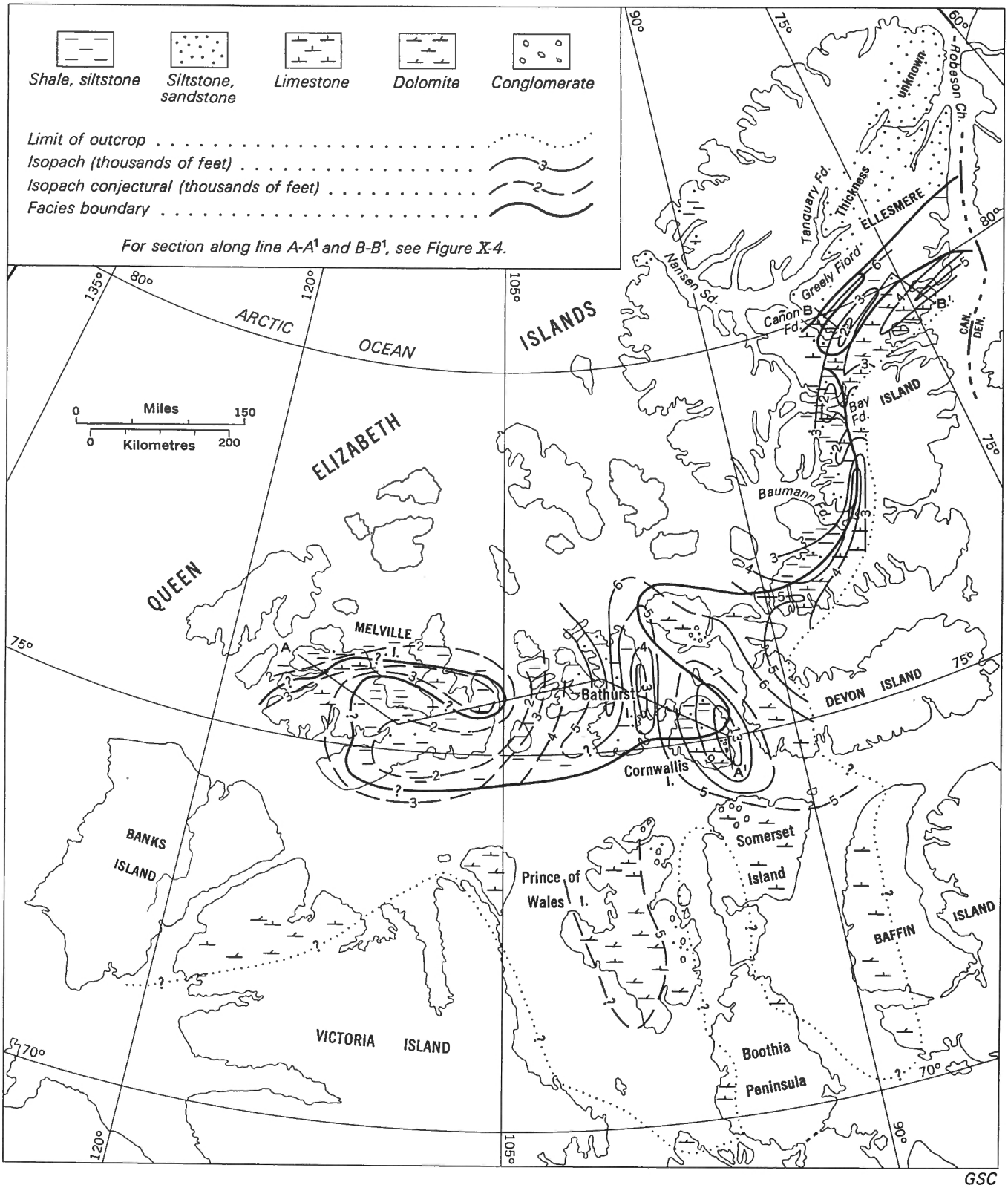

FIGURE X-3. Distribution of Upper Ordovician to Lower Devonian rocks in Arctic Archipelago (by R. Thorsteinsson).

limestone, dolomitic limestone, and intraformational conglomerate, with minor anhydrite and containing undescribed trilobites; the Cape Clay, about 300 feet of porous dolomite with gastropods, cephalopods, and trilobites; and an unnamed formation comprising about 300 feet of dolomitic limestone, limestone, and intraformational conglomerate.

The unfossiliferous Copes Bay Formation outcrops extensively on Ellesmere Island and includes 1,800 to 4,800 feet of thin-bedded impure limestone with anhydrite and limestone autobreccia as a minor but conspicuous component. Baumann Fiord Formation overlies both the Copes Bay and also Canadian rocks on Bache Peninsula. It consists of recessive weathering anhydrite, anhydritic limestone, and limestone from zero to 2,500 feet thick. Fossils are rare but *Hystricurus* is known from Flagler Fiord. This formation is exposed in the core of the Centre anticline on Cornwallis Island.

Conformably above the Baumann Fiord is the Eleanor River Formation composed mainly of resistant, thick-bedded limestone, from 800 to 2,700 feet thick. On Ellesmere Island the lower part contains *Ceratopea,* and on Cornwallis Island ostracods of Canadian or possibly younger age. The beds assigned to the Eleanor River Formation at Burnett Inlet, Devon Island, contain corals and cephalopods that have been dated as post-Canadian.

The Canrobert Formation of northwestern Melville Island consists of at least 1,000 feet of limestone and dolomite with intraformational conglomerate, calcarenite, and calcareous sandstone. Arenigian graptolites occur in the uppermost part; the base is nowhere exposed.

Mid-Ordovician and Silurian Carbonates and Evaporites

The Cornwallis Group is widely distributed in the Franklinian Miogeosyncline, in the northern part of Boothia Uplift, and in the adjacent lowland areas north of Parry Channel. It rests conformably on the Eleanor River Formation. South of Parry Channel correlative rocks are wholly carbonates and are assigned to several formations. Three formations are now distinguished within the Cornwallis Group: (1) The lower Bay Fiord Formation (780 to 1,800 feet of recessive weathering) is mainly thin-bedded limestone in part silty, shaly, and quartzose, and contains *Gonioceras* of Wilderness age. Locally the lower third consists of anhydrite. (2) The Thumb Mountain Formation comprises 1,000 to 2,800 feet of thick-bedded, resistant limestone, with some dolomite containing corals, brachiopods, and *Receptaculites,* probably of Barneveld age. (3) The uppermost formation, the Irene Bay, is a recessive unit of thin-bedded, shaly limestone with greenish grey shale partings, 120 to 675 feet thick. Cephalopods and other fossils of the Arctic Ordovician fauna of about Maysville age are abundant.

In northwest Baffin Island the sequence comprises, in ascending order, the Ship Point, Baillarge, and Cape Crauford Formations (Trettin, 1965). The Ship Point beds consist mainly of dolomite, in part shaly, silty, and sandy, with lesser dolomitic shale, siltstone, and sandstone. The lower beds may be as old as Arenigian, for graptolites occur in possible Ship Point correlatives in Foxe Basin. Fossils of about Wilderness age occur in higher beds. Disconformably above the Ship Point is the Baillarge Formation, divided into two members: Member A, up to 450 feet of limestone and dolomite, in part laminated and shaly, from which no fossils have been obtained; Member B, 1,140 feet of dolomitized cliff-forming limestone, with recessive units of shaly limestone. The lower 200 feet of this member contains corals and gastropods that indicate a correlation with the Thumb Mountain Formation of Ellesmere Island. This is overlain by 80 feet of green, shaly limestone with a fauna similar to that of the Irene Bay Formation; still higher beds in Member B contain Niagaran corals and brachiopods. The Cape Crauford Formation conformably overlies Baillarge and consists of up to 1,340 feet of dolomitic limestone, calcareous dolomite, and evaporite solution breccias. This formation, the youngest Paleozoic unit in northwest Baffin Island, also contains Niagaran fossils.

Ordovician shale and limestone with faunas of Wilderness, Barneveld, and probably also Richmond age, occur on the east and north margins of Foxe Basin. At Putnam Highland, southern Baffin Island, these rocks are at least 600 feet thick. Silurian (Niagaran) limestone and dolomitic limestone is preserved in the middle of the basin, on Prince Charles and adjacent islands.

In the western part of Arctic Lowlands from Somerset Island to Victoria Island all Ordovician rocks and part of the Silurian succession form a remarkably uniform and drab sequence consisting almost wholly of dolomite, dense to porous and vuggy, and fine to coarse grained. Chert, dolomitic shale and sandstone, and intraformational breccia are minor constituents, especially near the base. This sequence overlies Lower Cambrian strata paraconformably or rests nonconformably on Precambrian rocks. Rough estimates of the thickness vary from about 1,000 feet on Somerset Island to about 3,000 feet on Stefansson and Victoria Islands. The lower beds are unfossiliferous and of uncertain age. Higher beds contain a *Gonioceras* fauna of Wilderness age, the somewhat younger Arctic Ordovician fauna, and in the upper half Niagaran fossils are known. Ashgillian graptolites are known from one locality, a thin shale on Falsen Island off the east coast of Victoria Island.

Upper Ordovician, Silurian, and Lower Devonian Carbonates

Complex facies relationships, mainly within Franklinian Miogeosyncline and near Boothia Uplift, characterize the Upper Ordovician to Lower Devonian rocks (Figs. X-3, -4). The Allen Bay Formation (Upper Ordovician to Middle Silurian) and overlying Read Bay Formation (Middle Silurian to Lower Devonian) are carbonates that grade laterally into graptolitic shale and siltstone of the Cape Phillips Formation (Upper Ordovician to Lower

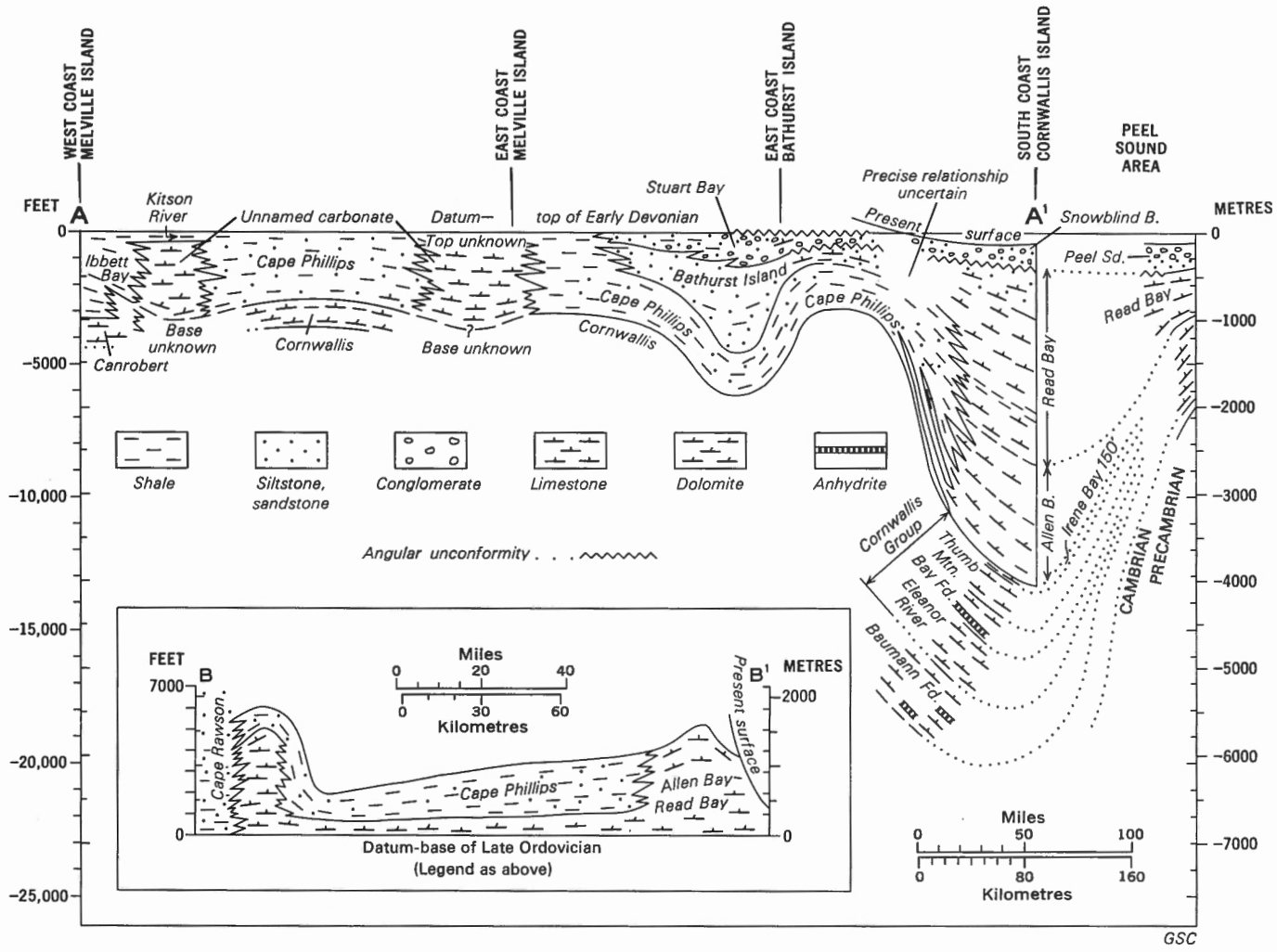

FIGURE X-4. Diagrammatic restored sections of Upper Ordovician to Lower Devonian rocks of Arctic Archipelago; (a) along line A–A', Figure X-3; (b) along line B–B', Figure X-3 (by R. Thorsteinsson).

Devonian). In northern regions of the Franklinian Geosyncline the Cape Phillips Formation passes laterally into sandstones and siltstones of the Cape Rawson Formation.

The Allen Bay Formation rests with sharp yet conformable contact on the Cornwallis Formation. In the type section on Cornwallis Island it is a uniform sequence, predominantly yellowish dolomite, about 5,500 feet thick (Thorsteinsson, 1959). Highly porous and massive beds, presumably representing biostromal reefs, are commonly impregnated with solid bitumen. Subordinate limestone and dolomitic limestone comprise 10 per cent of the formation. Elsewhere limestone and dolomitic limestone increase in proportion. South of Parry Channel rocks correlative with the Allen Bay Formation are present in the Baillarge and Cape Crauford Formations and in the unnamed dolomite of the western part of the lowlands. Ordovician, probably Richmondian, fossils occur in the lower beds of the Allen Bay Formation and Wenlockian ostracoderms in the upper part.

The Read Bay Formation is a varied assemblage of limestone, sandstone, dolomite, siltstone, and shale. In some parts of the archipelago it has the rank of a group. The basal contact is conformable, variably sharp or gradational. The formation outcrops in northern Victoria, Stefansson, Prince of Wales, Somerset, Devon, and southwestern Ellesmere Islands, and on Boothia Peninsula. It also outcrops in the Franklinian Miogeosyncline from Cornwallis Island, through northwestern Devon Island and Ellesmere Island. North of Baumann Fiord, the Allen Bay and Read Bay equivalents are indivisible although both limestone and dolomite occur in the lower part and limestone predominates above.

The type section of Read Bay Formation on Cornwallis Island (Thorsteinsson, 1959) is divisible into four members, in ascending order as follows: Member A, 1,875 feet of mainly thin bedded argillaceous limestone alternating with generally thick bedded biostromal and fossil fragmental limestone; Member B, 60 to 100 feet of shale and minor sandstone; Member C, 3,775 feet of beds similar to Member A, but with biostromal and biohermal develop-

ments that are thicker and commonly dolomitized; Member D, about 1,800 feet of alternating red calcareous sandstone and argillaceous limestone. The greater part of Member A contains the late Wenlockian to Ludlovian fauna characterized by *Atrypella scheii*. Higher beds of Member A contain *Monograptus bohemicus* of Lower Ludlovian age; Member C contains fossils that probably indicate a Lower Devonian age; Member D has not yielded diagnostic fossils.

On northwestern Devon Island the Read Bay Group (Thorsteinsson, 1963) embraces the Douro Formation, about 1,300 feet of interbedded thin-bedded argillaceous limestone and skeletal limestone with *Atrypella scheii*; the Devon Islands Formation, about 480 feet of shale containing the *Monograptus ultimus* Zone, which is provisionally regarded as Lower Devonian; the Sutherland River Formation, about 500 feet of medium-bedded, fine-grained dolomite with minor limestone and sandstone, containing a fauna formerly dated as Upper Silurian (Boucot, *et al.*, 1960); and the Prince Alfred Bay Formation, about 700 feet of unfossiliferous, thin- to medium-bedded, fine-grained sandstone with lesser limestone and dolomite. A somewhat similar sequence occurs in southwestern Ellesmere Island (Greiner, 1963, p. 292). On the north coast of Stefansson Island three unnamed formations are included in the Read Bay Group (Thorsteinsson and Tozer, 1962): the lower, about 400 feet of limestone and lesser dolomite with *Atrypella scheii*; the middle, about 125 feet of shale with lower Ludlovian graptolites; and the upper, about 150 feet of thin-bedded unfossiliferous dolomite. The Prince Alfred Bay Formation and the upper dolomite of Stefansson Island have not yielded diagnostic fossils. Their age is uncertain and they may be younger than the highest beds in the type section of the Read Bay Formation.

Carbonate banks occur north and west of the main belt of graptolitic shales on Ellesmere Island and Melville Island (Kerr, 1962; Tozer and Thorsteinsson, 1964). That of Ellesmere Island is about 5,000 feet thick and occurs as an apophysis of the main mass of Allen Bay and Read Bay rocks (Fig. X-3). It is bordered on the northwest by the Cape Rawson Formation and on the southeast by the graptolitic Cape Phillips Formation. The spatial relationship of the carbonate bank on Melville Island is uncertain owing to the nature of the exposures. This bank is situated north of a belt of Cape Phillips Formation and is possibly isolated. The thickness is estimated at 6,000 feet, but this figure probably includes about 3,000 feet of beds correlative with the upper part of the Cornwallis Group.

Graptolitic Rocks of Franklinian Miogeosyncline

Rocks of graptolitic facies are widely distributed in Franklinian Miogeosyncline, but the span of time covered by this type of sedimentation varies considerably from place to place. The longest span, recorded by the Ibbett Bay Formation on northwestern Melville Island, extends from the Arenigian (Lower Ordovician) to the Lower Devonian. The most widespread formation is the Cape Phillips, the basal beds of which are Upper Ordovician, Ashgillian, and locally the upper beds are Lower Devonian. On the west side of Boothia Uplift the Lower Devonian Bathurst Island and Stuart Bay Formations, characterized by fine-grained sandstone and graptolitic shale, overlie the Cape Phillips strata. Where graptolitic rocks overlie the carbonate banks they record only a short span of time, as shown by the 300-foot Lower Devonian Kitson Formation of Melville Island and by analogous deposits of uppermost Silurian age on Ellesmere Island.

The Ibbett Bay Formation comprises about 3,000 feet of black shale, with minor dolomite, chert, siltstone, and limestone. It is exposed only in the Canrobert Hills of northwest Melville Island, an area that represents the inner part of Franklinian Miogeosyncline. Graptolites indicate beds of Arenigian, Caradocian, Ashgillian, Llandoverian, Ludlovian, and Early Devonian age. It conformably overlies the Canrobert Formation, and is disconformably overlain by the Blackley Member of the Weatherall Formation.

The Cape Phillips Formation, predominantly graptolitic shale and siltstone, was defined originally on Cornwallis Island (Thorsteinsson, 1959). It outcrops in a linear belt that follows much of the Franklinian Miogeosyncline from Melville Island to Ellesmere Island. The transition zone from Allen Bay and Read Bay carbonate to the Cape Phillips Formation shale and siltstone is narrow and sharply defined on Cornwallis Island (Fig. X-4a). There the Cape Phillips Formation embraces the combined time span of the Allen Bay and Read Bay Formations, that is, Ashgillian to Emsian. On Ellesmere Island carbonate sequences of variable time spans and thicknesses, but which are invariably equivalent to the lower part of the Allen Bay Formation, extend northwesterly into regions predominantly of Cape Phillips facies. Beds of limestone, variably argillaceous and cherty, and chert, are common constituents in the Ashgillian and Llandoverian part, especially on Cornwallis and Bathurst Islands. On these islands dolomite is also common in the Ashgillian part. Shale and siltstone comprise the younger parts of the Cape Phillips Formation.

Where Cape Phillips beds represent a time span from Ashgillian to Lower Devonian, the thickness ranges from about 1,400 to more than 8,500 feet. The formation thickens to the south towards the line of facies change, with minimal thickness in medial regions of the miogeosynclinal belt. It also thickens gradually northward changing into the Cape Rawson facies. In general the Upper Silurian and Lower Devonian parts of the Cape Phillips Formation account for more than half of the thickness.

On Bathurst Island the graptolitic Lower Devonian is represented by the Bathurst Island and Stuart Bay Formations (McLaren, 1963b; Kerr and Temple, 1965). The Bathurst Island Formation comprises up to 3,410 feet of very fine grained sandstone, variably argillaceous and calcareous, calcareous and silty mudstone, and minor dolomitic siltstone. The typical Stuart Bay rocks comprise 1,220

feet of calcareous, argillaceous, fine-grained sandstone with some mudstone, siltstone, and minor bioclastic limestone, and thin limestone and chert conglomerate beds at the base. In eastern Bathurst Island, Stuart Bay beds contain a greater proportion of limestone, and include conspicuous reefoid masses and conglomerates. The contact of the Bathurst Island Formation with the underlying Cape Phillips beds is transitional, but nevertheless appears to be approximately isochronous wherever observed. On the other hand, the contact between the Bathurst Island and Stuart Bay beds is markedly diachronous. West and north of the type locality, the Bathurst Island Formation thickens gradually at the expense of Stuart Bay beds, entirely replacing the latter. South and east of the type locality the relationships are reversed and further complicated by an unconformity at the base of the Stuart Bay Formation, beneath which the Bathurst Island Formation is deeply eroded and in places cuts out entirely. This angular unconformity within the Lower Devonian marks the first major movement of Boothia Uplift within the Paleozoic. In central Bathurst Island the Stuart Bay Formation is overlain, presumably disconformably, by the Middle Devonian Eids Formation. In eastern Bathurst Island, the Disappointment Bay Formation, a dolomite equivalent to the lower part of the Eids Formation rests with angular unconformity on Stuart Bay and older beds. The second movement of Boothia Uplift is thus narrowly bracketed in time by late Lower and Middle Devonian rocks. Although the main source for the Bathurst Island and Stuart Bay sediments appears to have been from the north, a secondary source, particularly for the coarse clastic rocks in the Stuart Bay Formation, was Boothia Uplift.

Clastic and Volcanic Rocks of the Franklinian Eugeosyncline

On Ellesmere Island a thick sequence of Ordovician (?) to Devonian clastic rocks extends in a belt from the head of Trold Fiord, through central Canyon Fiord, northeast to Archer Fiord. These rocks, known as the Cape Rawson Formation, form a transition between sequences characteristic of the miogeosyncline to the southeast and the eugeosyncline to the northwest. Cape Rawson rocks are also exposed northwest of Archer Fiord, where they are characterized by more intense folding and low rank metamorphism. Northwest of Archer Fiord the base is not exposed, but at the head of the fiord the Cape Rawson overlies the Ordovician Cornwallis Group, and to the south in Canyon and Trold Fiords it rests gradationally on the Silurian part of the Cape Phillips Formation.

The Cape Rawson consists mainly of sandstone and siltstone in the regions of facies change to the Cape Phillips and Eids Formations. To the north it is composed of greywacke, subgreywacke, impure sandstone, chert, micaceous gritty shale, phyllite, and impure sandstone (Christie, 1964). The top has not been recognized, and it is possible that upper beds are equivalent to the Upper Devonian Okse Bay Formation.

Although great thicknesses of sediments are probably represented, the complexity of structures has precluded accurate measurements. About 15,000 feet of Cape Rawson beds is estimated to occur in Canyon Fiord. Cape Rawson rocks are generally barren of fossils; nevertheless, three collections obtained indicate Lower Silurian, Upper Silurian, and Lower Devonian rocks. It is possible that rocks described by Blackadar (1954) on the north coast of Ellesmere Island as the Mount Disraeli, Sail Harbour, and View Creek Groups are equivalent to the Cape Rawson Formation.

In M'Clintock Inlet, on the north coast of Ellesmere Island, seven formations have recently been distinguished by Trettin (1966) within the rocks classed as the M'Clintock and Challenger Groups by Christie (1964). The total thickness probably exceeds 12,000 feet. The oldest rocks are Ordovician carbonates with *Gonioceras* of Wilderness age, resting unconformably on slates and phyllites that are probably equivalents, at a lower metamorphic rank, of the Cape Columbia gneiss. Above the Wilderness carbonates are red sandstones, succeeded by several thousand feet of volcanic flows, in turn overlain by carbonates with Richmondian Arctic Ordovician fauna and graptolitic beds and carbonates with *Climacograptus* cf. *latus* Elles and Wood of Ashgillian age. The Ashgillian rocks are succeeded by more than 750 feet of sandstone and conglomerate believed on lithological grounds to be correlative with the Imina Group, which contains Llandovery graptolites farther to the west. Finally, above the Imina equivalent there is at least 1,300 feet of limestone and calcareous sandstone, with *Conchidium,* of Late Silurian age.

Silurian rocks are also known farther to the west, in northwestern Ellesmere Island (Trettin, 1964). The Imina Group comprises several thousand feet of beds, predominantly lithic arenite and siltstone with lesser shale and conglomerate, and contains Llandoverian graptolites in the upper part. Between 3,000 and 6,000 feet of slaty shale and slaty siltstone apparently overlies the Imina, locally with volcanic flows of keratophyric or quartz keratophyric composition (especially in the upper part) and containing graptolites of Llandoverian or Wenlockian, and Ludlovian age. The succeeding 5,000 feet of siltstone, shale, sandstone, and conglomerate carries Ludlovian graptolites; these rocks unlike the underlying volcanic-derived formation are composed mainly of chert and quartz, laminated and thin bedded with low-angle crossbedding and evidence of soft sediment deformation.

On northern Axel Heiberg Island there is at least 3,000 feet of slate and siltstone with some lithic arenite, tuffaceous arenite, and tuff containing Wenlockian graptolites. As on Ellesmere Island, the volcanic material is chiefly keratopyhric and quartz keratophyric. Groove casts indicate southeasterly to easterly current directions. These rocks lie in fault contact with the Rens Fiord Complex and are overlain unconformably by the Lower Devonian Stallworthy Formation. This is the only locality

known within Franklinian Eugeosyncline where there is evidence for an unconformity within the Silurian–Devonian sequence.

Lower Devonian Red Beds of the Boothia Uplift

In the Boothia Uplift widespread Lower Devonian red beds reflect the onset of tectonic elevation. The Snowblind Bay Formation on Cornwallis Island comprises up to 1,500 feet of red limestone-conglomerate, dolomite, siltstone, limestone, dolomitic limestone, and dolomitic siltstone and sandstone. On the east coast of the island these beds appear to overlie the Read Bay Formation conformably. In the interior there is an unconformity within the Snowblind Bay sequence. There the upper Snowblind Bay beds rest with angular unconformity on the lower beds and also on older formations. Ostracoderms indicate a Gedinnian age. Invertebrate fossils from clasts in the limestone-conglomerate attest to a provenance of eroded Read Bay and Allen Bay Formations. The Snowblind Bay Formation evidently represents a syntectonic deposit, essentially correlative with the Stuart Bay Formation of Bathurst Island. The angular unconformity within the Snowblind Bay Formation is correlated with the unconformity at the base of the Stuart Bay beds.

A sedimentary environment and tectonic history comparable to that of the Snowblind Bay beds is attributed to the approximately correlative Peel Sound Formation of Somerset and Prince of Wales Islands. Adjacent to the Precambrian exposures of the uplift the Peel Sound beds contain large boulders of Precambrian crystalline rocks. In western Prince of Wales Island, Peel Sound beds are relatively fine grained and intergrade and intertongue with the uppermost part of the Read Bay Formation; they are eventually replaced by a purely marine sequence devoid of red beds and conglomerate.

Lower Devonian Red Beds and Volcanic Rocks of the Eugeosyncline

In the eugeosynclinal terrane of northern Axel Heiberg Island a sequence of Lower or Middle Devonian red beds, the Stallworthy Formation, is abruptly but concordantly succeeded by the Svartevaeg volcanic rocks (Fricker and Trettin, 1962; Trettin, 1964). The Stallworthy Formation is divisible into three members. Member A at the base comprises up to 3,000 feet of beds. A basal pebble-conglomerate and sandstone is overlain by red siltstone, sandstone, and shale. Member B includes 1,000 to 2,500 feet of conglomerate, breccia, sandstone, and minor siltstone. Member C comprises 8,000 feet of red and green siltstone, sandstone, conglomerate, breccia, and calcareous and tuffaceous arenite with Lower or early Middle Devonian ostracoderms. Chert and quartz are the dominant constituents of all the sedimentary rocks. The Svartevaeg Formation comprises turbidites and volcanic rocks, about 10,000 feet of volcanic arenite, tuff, keratophyric and spilitic flows, siltstone, shale, conglomerate, and breccia. Silurian brachiopods occur in limestone boulders in a basalt, but autochthonous fossils are unknown. The Svartevaeg Formation is overlain unconformably by the Viséan Emma Fiord Formation.

Middle and Upper Devonian

Middle and Upper Devonian rocks are widely distributed in the Franklinian Miogeosyncline and occur as outliers within the Boothia Uplift (Fig. X-5). They are also exposed in the western part of the Arctic Lowlands, where they show close similarities with the succession on the mainland (Chap. VIII). Throughout Franklinian Miogeosyncline Middle Devonian rocks succeed the Lower Devonian conformably, but in the region of Boothia Uplift an angular unconformity separates the Middle Devonian from older rocks. The Middle and Upper Devonian rocks are divisible into two gross lithologic units, the lower of which is dominantly carbonate and shale and the upper mainly terrigenous sediments representing a clastic wedge derived from the north. In the inner part of the miogeosyncline, adjacent to the eugeosyncline, both the Middle and Upper Devonian are dominated by clastic rocks.

The lower part of the sequence in the outer part of the miogeosyncline includes dominantly carbonate formations (Disappointment Bay and Blue Fiord); drab coloured shales, siltstones, and fine-grained sandstones (Goose Fiord and Eids Formations); and red and green clastics with evaporites. Most or all of these rocks are of early Middle Devonian (Eifelian) age. The upper part of the sequence shows the initiation of clastic sedimentation throughout the whole of the miogeosyncline and the western part of the lowlands. Except for some isolated Frasnian reefs on Banks Island, carbonate formations are completely absent. Initially the sedimentation was mainly marine; later it was almost entirely non-marine. The principal marine formations are the Bird Fiord and the greater part of the Weatherall, both of Givetian age. The younger, mainly non-marine, formations are the Okse Bay and its approximate equivalents the Hecla Bay and Griper Bay Formations. Intercalated marine bands of Frasnian age occur on Prince Patrick Island and of Famennian age on Melville and Cameron Islands.

In the inner parts of the miogeosyncline the two divisions are not discernible, but are represented by the Cape Rawson Formation, a clastic wedge deposit which may range from Upper Ordovician to Middle or even Upper Devonian. Parts of the Devonian succession of northwestern Melville Island resemble the Cape Rawson Formation, but there the oldest thick clastics (Blackley Member of Weatherall Formation) are probably early Middle Devonian.

Deposition of clastic rocks of northerly derivation in the miogeosyncline commenced with the Upper Ordovician or Silurian Cape Rawson Formation, and continued until possibly the Upper Devonian. By the early Middle Devonian they had spread to areas such as northwest Melville Island, and Trold and Canyon Fiords on Ellesmere

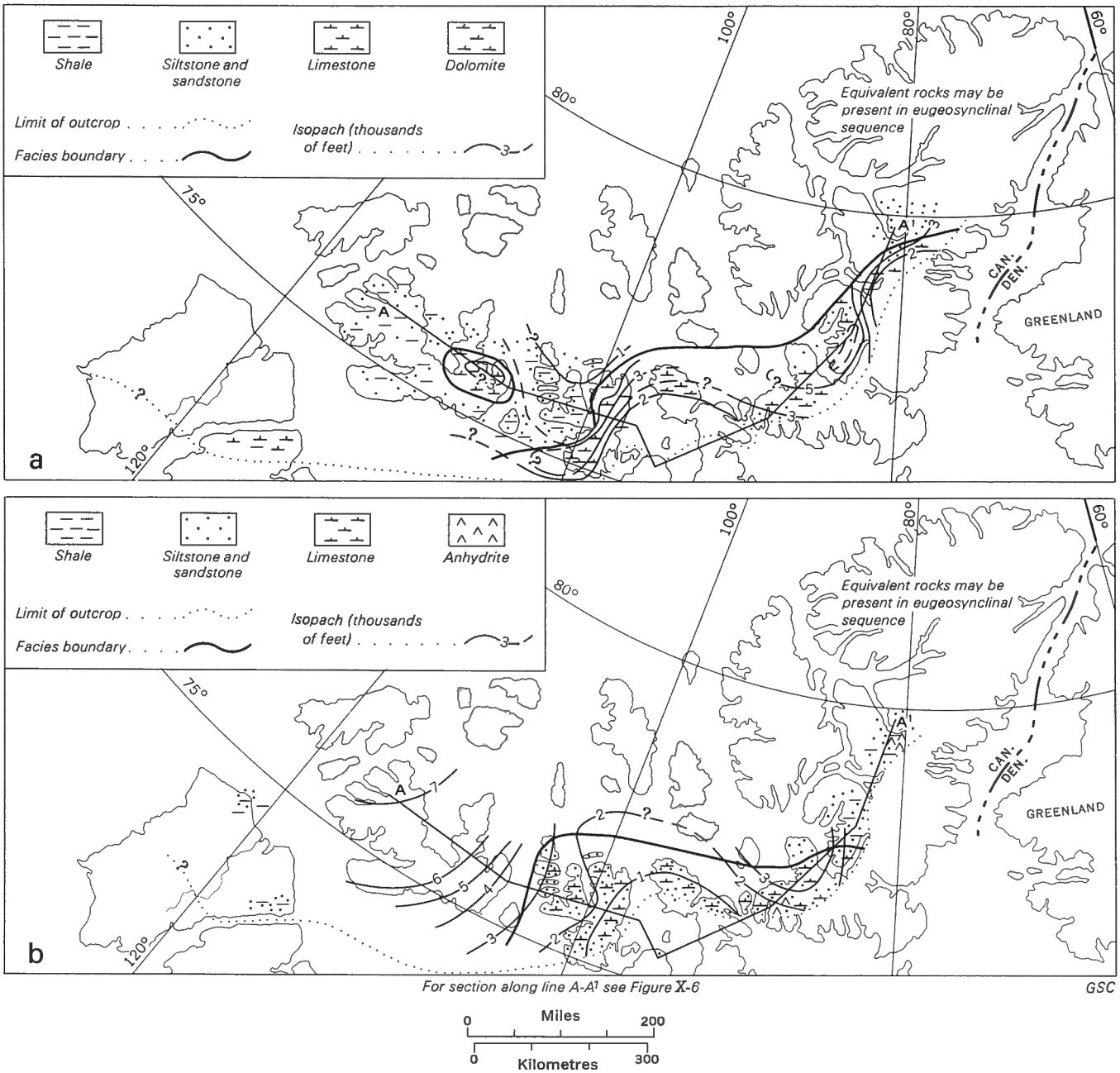

FIGURE X-5. Distribution of Middle Devonian rocks in Arctic Archipelago; (a) Eifelian, (b) Givetian (by R. Thorsteinsson).

Island. In late Middle Devonian (Givetian) time the clastic beds achieved their greatest extent by covering the miogeosyncline and spreading beyond, south of Parry Channel, to the lowlands. These conditions persisted until latest Devonian (Famennian) time.

Early Middle Devonian (Eifelian)

The early Middle Devonian rocks of the miogeosyncline include a great variety of carbonate, clastic, and evaporitic rocks. The Disappointment Bay Formation where originally defined on Cornwallis Island comprises about 280 feet of porous dolomite with basal sandstone and conglomerate overlying Silurian rocks with angular unconformity (Thorsteinsson, 1959). The upper contact of the formation is not preserved on Cornwallis Island. Rocks previously assigned to the Sherard Osborn and Driftwood Bay Formations on eastern Bathurst Island by Thorsteinsson and Glenister (1963) have been consigned by J. W. Kerr to the Disappointment Bay Formation. On Bathurst Island, the Disappointment Bay includes 700 to 2,100 feet of beds primarily resistant, generally porous dolomite, including a local basal sandstone and pebble-

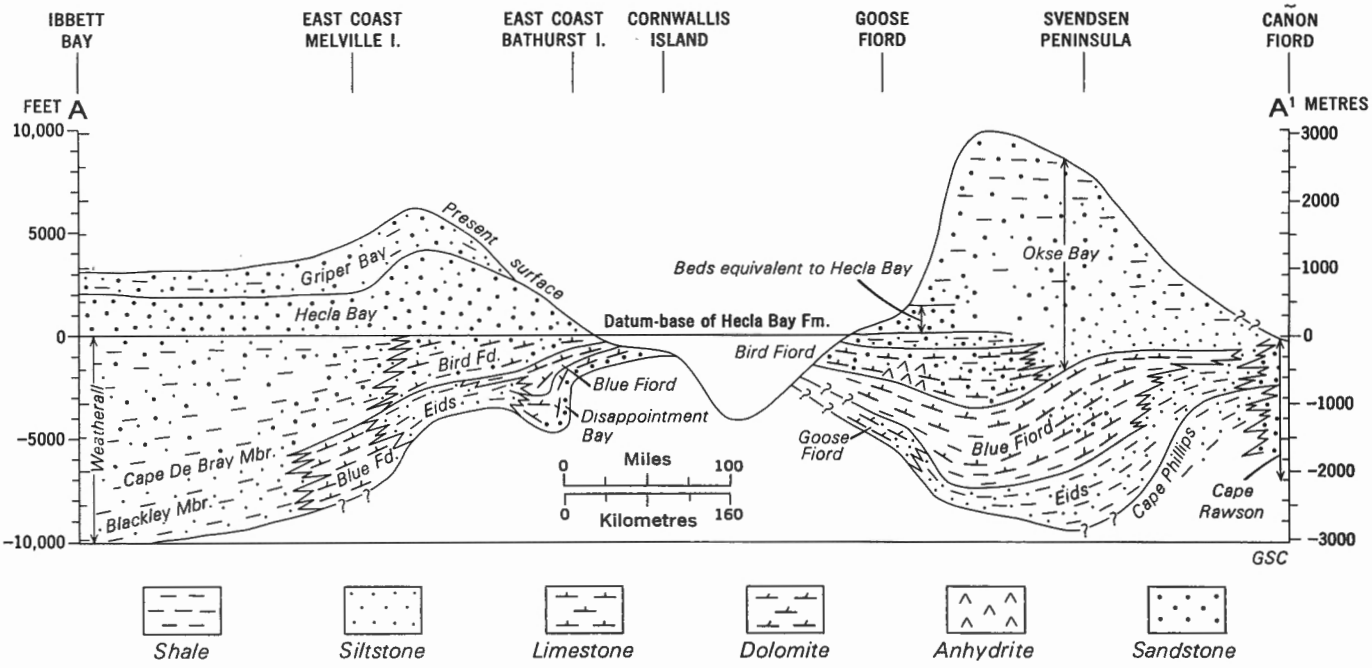

FIGURE X-6. Diagrammatic restored section of Middle and Upper Devonian rocks of Arctic Archipelago along line A–A′ of Figure X-5 (by R. Thorsteinsson).

conglomerate, formerly considered to represent the Driftwood Bay Formation. The Disappointment Bay Formation is overlain with sharp yet conformable contact by limestones of the Blue Fiord Formation. The thickest sections of Disappointment Bay occur in southern Bathurst Island where the formation includes numerous interbeds of dolomitic sandstone. In eastern Bathurst Island it rests with angular unconformity upon various Ordovician to Lower Devonian formations. To the west the angular unconformity passes into a disconformity with Disappointment Bay rocks overlying the Stuart Bay Formation. Still farther west the Disappointment Bay Formation grades laterally into the lower part of the Eids Formation through a transition zone characterized by a gradual decrease in dolomite and corresponding increase in siltstone.

The Eids Formation is composed of thin-bedded, recessive rocks, 600 to more than 3,000 feet thick, mainly of medium grey siltstone, variably shaly and calcareous, and minor calcareous, locally gypsiferous shale. The base of the Eids is a disconformity approximating a time-stratigraphic surface. The upper contact however is diachronous, overlain on Ellesmere Island with gradational contact by carbonate rocks of the Blue Fiord Formation. The Eids is developed in western Bathurst Island but eastward it grades laterally into the Disappointment Bay and Blue Fiord Formations. Correlation of the Eids Formation with the Blue Fiord and Weatherall Formations of Melville Island is speculative, as faunal evidence is lacking and exposures are discontinuous.

The Goose Fiord Formation comprises 1,010 feet of dolomitic siltstone and sandstone, silty dolomite and dolomite preserved in southwestern Ellesmere Island where it lies with an abrupt yet conformable contact on the Read Bay Group (Greiner, 1963). It is overlain with similar contact relationships by the Blue Fiord Formation. Goose Fiord rocks have not yielded diagnostic fossils but they probably represent a relatively sandy facies equivalent to the more argillaceous Eids Formation that occurs to the north, in the axial part of the miogeosyncline.

The Vendom Fiord Formation is confined to parts of the Arctic Lowlands and Franklinian Miogeosyncline on Ellesmere Island (Fig. X-7). It comprises up to 600 feet of red and green, thin-bedded, variably calcareous and silty sandstone, with quartz-pebble conglomerate, siltstone, and anhydrite as minor constituents. These rocks outcrop as a narrow belt that extends from Vendom Fiord where they grade into the Eids Formation, to the head of Canyon Fiord where they merge with the Cape Rawson Formation. They rest disconformably on either the Allen Bay–Read Bay Formations or the Cape Phillips Formation, and are conformably overlain by the Blue Fiord Formation. Their distribution and facies relationships indicate that these sediments, unlike other Devonian clastic formations, were probably derived mainly from an eastern source.

From interfingering relationships and stratigraphic position it appears that the Disappointment Bay, Eids, and Goose Fiord Formations, and also the red and green beds between Vendom and Canyon Fiords, are essentially correlative. Most of these rocks are devoid of diagnostic fossils but the Eids contains trilobites of Eifelian age (Ormiston, *pers. com.*), and it is therefore presumed that the other formations are also Eifelian.

The Blue Fiord Formation is characteristically developed as a resistant, microcrystalline and fossil frag-

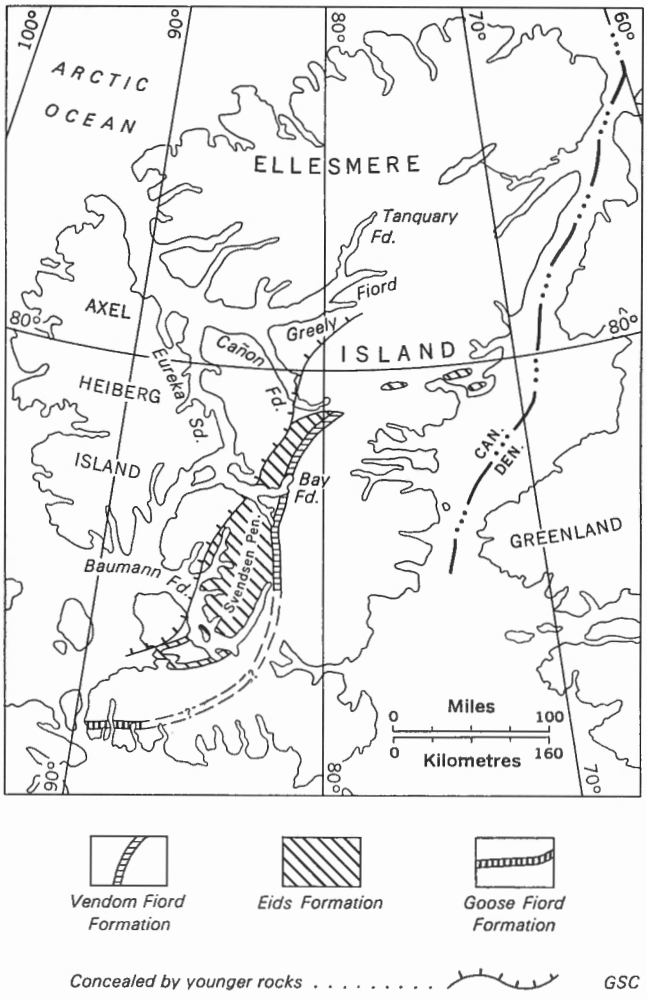

FIGURE X-7. Outcrop of Middle Devonian, Vendom Fiord, Eids, and Goose Fiord Formations, Ellesmere Island (by R. Thorsteinsson).

mental limestone, with lesser dolomite, shale, and siltstone (Figs. X-5a and 6). Dolomite forms a conspicuous constituent and locally is dominant. Outcrops of Blue Fiord rocks are distributed as an irregular belt along the southeastern and southern margins of the Franklinian Miogeosyncline from Ellesmere Island, northwestern Devon Island, Bathurst Island, and Melville Island. In northeastern Melville Island the formation includes about 2,500 feet of limestone and dolomite. There the formation is interpreted as representing an isolated carbonate bank and may include rocks correlative with the Disappointment Bay Formation. The Blue Fiord Formation outcrops in the Arctic Lowlands of northwestern Victoria Island, where neither the base nor the top of the formation is exposed. The formation is characterized by highly variable thicknesses; from zero where it has graded into correlative formations, to a maximum of more than 4,500 feet between Vendom and Bay Fiords on Ellesmere Island. It contains a rich fauna of Eifelian corals, brachiopods, and trilobites.

Late Middle Devonian (Givetian) and Non-Marine Clastic Rocks

The Givetian is represented by the Bird Fiord Formation on Ellesmere, Devon, and Bathurst Islands, and by the Weatherall Formation of eastern Melville Island (Figs. X-5b, -6). The Weatherall of western Melville Island includes a basal member (Blackley) that is probably of early Middle Devonian age.

The Bird Fiord Formation is composed of light greenish grey calcareous sandstone and quartzose limestone with interbeds of dark green shale and siltstone. Corals, bivalves, brachiopods, and trilobites are common. The contact with the underlying Blue Fiord carbonates is sharp but conformable.

On northeastern Melville Island the Weatherall Formation abruptly and conformably overlies the Blue Fiord Formation. It is divisible into (1) a lower sandstone, siltstone, and shale sequence, some 700 feet thick, and variably calcareous and drab weathering with marine Givetian fossils; and (2) an upper sandstone, siltstone, and shale sequence, about 3,900 feet thick, devoid of marine fossils and probably of non-marine deposition. Carbonaceous fragments occur throughout, but especially characterize the upper member. On northwestern Melville Island the lower part of the Weatherall Formation includes the Blackley Member, which rests with abrupt and probably disconformable contact on the Lower Devonian graptolitic beds of the Ibbett Bay Formation. It comprises about 2,300 feet of apparently unfossiliferous grey micaceous shale with interbeds of siltstone. The succeeding Cape de Bray Member comprises about 2,000 feet of grey to black micaceous shale with siltstone interbeds, in which Givetian brachiopods occur. The Cape de Bray Member is followed by about 6,000 feet of beds that closely resembles the typical Weatherall rocks of eastern Melville Island. Givetian corals, brachiopods, trilobites, and bivalves are common in the beds above the Cape de Bray Member, but towards the top of the Weatherall marine fossils become increasingly rare. In the absence of fossil evidence, the correlation of the Blackley Member with the Disappointment Bay and Blue Fiord Formations is tentative. However, if this interpretation is correct the Weatherall Formation of western Melville Island is both Eifelian and Givetian whereas in the eastern part of the Island it is only Givetian.

Late Middle (?) and Upper Devonian

The Devonian rocks that succeed the Weatherall and Bird Fiord Formations are mainly non-marine. East of Grinnell Peninsula they are known as the Okse Bay Formation, and on Bathurst Island and to the west they are represented by the Hecla Bay and Griper Bay Formations of the Melville Island Group (Pl. X-2). These rocks, in which the youngest dated beds are Famennian, are the youngest deposited in Franklinian Geosyncline. Their upper contact is everywhere a profound angular unconformity.



PLATE X-2. Upper Devonian rocks and late Cenozoic fault, Blue Hills, southwestern Melville Island, view eastward. H–Hecla Bay Formation; G–Griper Bay Formation.

The Hecla Bay Formation is a uniform succession 1,200 to about 4,000 feet thick, of light grey to white sandstone that is generally thick bedded and cross-stratified. Carbonaceous beds characterize the formation which is also relatively resistant to weathering and of distinctive colour. The contact with the Weatherall Formation is abrupt on eastern Melville Island, but becomes transitional to the west. On Bathurst Island and Grinnell Peninsula it gradationally overlies the Bird Fiord Formation. Although the Hecla Bay Formation has not yielded diagnostic fossils, its age is limited by the Givetian and Famennian ages of underlying and overlying formations.

The Griper Bay Formation is mainly sandstone, siltstone, and shale with thin coal seams and conglomerate as minor constituents. The rocks are characterized by green and greyish colours with units of white sandstone like those of the Hecla Bay Formation. Up to 4,000 feet of Griper Bay rocks outcrop in the Franklinian Miogeosyncline on Melville and Bathurst Islands, where they rest concordantly on the Hecla Bay sandstones. They also occur on Boothia Uplift where they overlie with angular unconformity various older formations, the youngest being Givetian. This unconformity attests to the third period of movement of Boothia Uplift. The greater part of the



PLATE X-3. Upper Devonian Frasnian Gyrfalcon Bluffs, northern Banks Island.

Griper Bay was evidently deposited under non-marine conditions but thin beds with Famennian brachiopods occur about 2,000 feet above the base on Melville and Bathurst Islands (Kerr, *et al.*, 1965). Estherians, fragments of armoured fish and spores, have also been described from the Griper Bay of Melville and Bathurst Islands. The rocks assigned to this formation on Prince Patrick Island comprise about 4,500 feet of beds with no base exposed. Thin-bedded sandstone and siltstone with Frasnian brachiopods and corals form the lower part of the section; thick-bedded, light coloured non-marine sandstone, with thin coals, characterizes the upper part. The marine Frasnian beds have no known counterparts on the island to the east. Strata more or less equivalent to the Griper Bay are also present on Banks Island where they include considerable sandstone and siltstone but also some isolated Frasnian limestone reefs (Pl. X-3), the only known Upper Devonian reefs within the archipelago.

The Okse Bay Formation on Ellesmere Island (McLaren, 1963a) includes at least 10,000 feet of sandstone, siltstone, and shale with thin coal seams. They have yielded plant and fish fossils not dated more precisely than Upper Devonian. The formation has four members: a lower sandstone, a lower sandstone and shale, an upper sandstone, and an upper sandstone and shale. Two subdivisions are recognizable in the lower sandstone member, a lower unit of red and green sandy shale, and an upper unit comprising 1,350 feet of thick-bedded to massive sandstone, which appears to be the equivalent of the Hecla Bay Formation. Presumably the upper three members of the Okse Bay Formation are correlative with the Griper Bay Formation. North of the latitude of Vendom Fiord on Ellesmere Island the lower part of the Okse Bay Formation is believed to be older than to the south, and to include equivalents of the Bird Fiord and Blue Fiord Formations.

Ellesmerian Orogeny

The main interval of folding during the Ellesmerian Orogeny was between the Famennian and the Viséan. This is shown by the occurrence of Famennian rocks at the top of the essentially concordant sequence in the Franklinian Miogeosyncline, and by the local occurrence of Viséan strata at the base of the sequence in the Sverdrup Basin, which unconformably overlies the folded rocks of the Franklinian Geosyncline.

In the Franklinian Miogeosyncline the principal structures produced by the Ellesmerian Orogeny are the Parry Islands and Central Ellesmere Fold Belts. The eugeosyncline evidently experienced a long sequence of tectonic events. Some movement certainly preceded the deposition of Lower Devonian rocks. However, from available evidence it would appear that the principal period of orogenic activity within the eugeosyncline, as in the miogeo-



PLATE X-4. Parry Islands Fold Belt, northwest Melville Island, view eastward. Dark rocks in cores of anticlines are Ordovician, Silurian, and Devonian conformable sequences folded during the Ellesmerian and Melvillian Orogenies unconformably overlain by the Upper Carboniferous Canyon Fiord Formation (C) in the synclines, folded in the Melvillian Disturbance. In the distance are gently dipping Permian and Mesozoic beds (M) of southern Sverdrup Basin.

syncline, was between the Devonian and the Viséan.

The Parry Islands Fold Belt of Melville and Bathurst Islands is characterized by long, narrow, more or less symmetrical anticlines separated by broad, shallow synclines (Pl. X-4). A few steeply dipping thrust faults of small horizontal displacement parallel the regional strike of the folds. The major structural features of Parry Islands Fold Belt suggest a *décollement*, probably within the evaporitic sequence of the Ordovician Cornwallis Group.

The east–west strike of Parry Islands structures is nearly at a right angle to the north–south regional trend of structures of Boothia Uplift. On eastern Bathurst Island the two trends meet in an area of structural complexity. It is characterized chiefly by close, northerly trending folds as well as by northerly striking faults, several of which have strike-slip components. Certain folds within the Boothia Uplift adjacent to Parry Islands structures have undergone a second episode of folding as a result of the Ellesmerian Orogeny; however, the greater part of the uplift appears to have acted as a buttress that was little affected by the orogeny.

The Central Ellesmere and the Parry Islands Fold Belts bear similar structural relationships to the Boothia Uplift. Structures of Central Ellesmere Fold Belt strike southeasterly from Grinnell Peninsula across northwestern Devon Island proper, whence they curve northeasterly into southwestern Ellesmere Island and continue diagonally across the island. From Grinnell Peninsula to about the latitude of Baumann Fiord characters of the Central Ellesmere Fold Belt are analogous to those of the greater part of the Parry Islands Fold Belt. The generally simple fold patterns displayed there suggest that the southwestern part of the Central Ellesmere Fold Belt was affected by the Ellesmerian period of orogeny only.

Between Baumann Fiord and Kane Basin the Central Ellesmere Fold Belt experienced a long, complex, structural history. Folds, thrust faults, and normal faults, all striking northeast, are present. The thrust faults, commonly *en echelon*, generally dip northwest at moderate to steep angles and extend for more than a hundred miles. Two episodes of deformation, the Ellesmerian and Eurekan in the Cenozoic, are clearly in evidence in this region. This is shown by the presence of numerous outliers of the Cenozoic Eureka Sound Formation, resting with angular unconformity on rocks of early Paleozoic age but nevertheless folded and overridden by thrust faults. The relative effects of the two orogenies is difficult to assess. Many folds and normal faults, but no thrusts, are demonstrably older than the deposition of the Eureka Sound Formation, suggesting that the structures of the Ellesmerian Orogeny were predominantly folds. Several thrust faults demonstrably postdate the Eureka Sound Formation and many normal faults postdate both the Eureka Sound Formation and thrust faults. Probably the major episode of thrust faulting resulted from the Eurekan Orogeny whereas the normal faults represent a still later structural event.

The orogenic history of the Franklinian Eugeosyncline is more complex than that of its complementary miogeosyncline. The Franklinian Eugeosyncline had undergone several orogenic episodes, possibly in the Lower Cambrian in northern Ellesmere and northern Axel Heiberg Islands from the evidence provided by radiometric ages, and between the Middle Silurian and Lower Devonian in northern Axel Heiberg Island. The principal orogeny however was the Ellesmerian as shown by the widespread angular unconformity below rocks of Carboniferous age. The youngest unequivocally dated rocks below this unconformity are of Lower or Middle Devonian age. Consequently, the question of whether the Ellesmerian Orogeny or a somewhat older orogeny of about Middle or early Late Devonian affected the Franklinian Eugeosyncline cannot be answered on the basis of available evidence.

On northwestern Ellesmere Island small granitic plutons range in composition from quartz monzonite to granodiorite, quartz diorite, and diorite (Trettin, 1964). They are variously massive to foliate, and concordant to discordant. Radiometric age determinations from a pluton on Axel Heiberg Island and another from Ellesmere Island yielded ages of 360±25 m.y. (late Devonian) and 335±25 m.y. (early Carboniferous) respectively. The youngest rocks intruded are probably of Early Silurian age. In all probability the emplacement of these plutons was related to the Ellesmerian Orogeny. Accordingly it is unlikely that the unconformity between rocks of Middle Silurian and Lower or Middle Devonian in northern Axel Heiberg Island reflects the principal orogeny in the eugeosyncline. In the sedimentary record of the miogeosyncline, however, there is evidence that the orogenic events and uplift in the eugeosyncline preceded the Ellesmerian Orogeny of the miogeosyncline. This is shown by the clastic deposits that first appeared in the Upper Ordovician or Lower Silurian, and gradually became more extensive, reaching their maximum extent in the late Middle and Upper Devonian.

UPPER PALEOZOIC AND MESOZOIC

The end of the Ellesmerian Orogeny in late Devonian or early Carboniferous time marks a complete change in the régime of sedimentation and tectonism in the Arctic Archipelago. Following this orogeny a new cycle of sedimentation was initiated that lasted from the Carboniferous to the Upper Cretaceous. This cycle was terminated by the Eurekan Orogeny in early Cenozoic time.

The principal site of sedimentation from the Carboniferous to the Upper Cretaceous was Sverdrup Basin. In the Upper Carboniferous it is possible that small detached basins may also have existed south of the main area of subsidence, for example on northwest Melville Island. Cretaceous rocks also extend beyond the basin, most extensively in the western part of the lowlands. The Carboniferous and Permian rocks include clastics of varied grain size, carbonates, evaporites, and volcanic rocks, resting with angular unconformity on the bevelled structures of the Franklinian Orogen. The Lower Carboniferous deposits are non-marine; Upper Carboniferous and Permian beds are mainly marine.

In the Mesozoic a more monotonous suite is encountered. Nearly all rocks are medium- to fine-grained quartzose sandstone, siltstone, and shale with minor amounts of conglomerate and some basalts in the Cretaceous. Marine and non-marine deposits alternate in about equal volume. Biogenic carbonate rocks and evaporites are totally absent. Possibly cooler climatic conditions prevailed than in the Paleozoic.

There was relatively little tectonic activity between the Ellesmerian and Eurekan Orogenies, except on the margins of the Sverdrup Basin. The Melvillian Disturbance, of about Early Permian age, produced folds and faults near the south and east margins. At the western margin, defined by the Prince Patrick Uplift, movement on gravity faults was probably initiated in the Late Jurassic. All the margins were relatively resistant to subsidence and show incomplete sequences with many disconformities and gentle discordances for the Carboniferous to Upper Cretaceous interval. In the axial part of the basin, on the other hand, the sequence from the Upper Carboniferous to the Upper Cretaceous is essentially concordant. In places the concordant sequence even includes beds of early Cenozoic age. From a regional standpoint, however, there is no doubt that early Cenozoic time marks the end of uniform subsidence within the Sverdrup Basin.

Gabbroic dykes and sills intrude most of the Carboniferous to Cretaceous formations. They were probably emplaced in Cretaceous time.

UPPER PALEOZOIC

Carboniferous and Permian

For the Carboniferous and Permian the record of sedimentation includes beds of the following ages: Lower Carboniferous (Viséan); Upper Carboniferous (Bashkirian, Moscovian, Gzhelian or Zhigulevskian, and Orenburgian); and Permian (Asselian, Sakmarian, Artinskian, and Guadalupian). Earliest Carboniferous (Tournaisian), mid-Carboniferous (Namurian), and latest Permian (Dzhulfian) rocks are unknown and probably absent.

The Viséan rocks are all non-marine and are separated from the younger beds by a regional unconformity.

The Bashkirian to Guadalupian sequence is one of great complexity with intertonguing formations of conglomerate, sandstone, shale, limestone, and evaporites (Fig. X-8). Volcanic rocks also occur locally. In general the carbonate rocks, in part reefoid, intertonguing with dark shales and associated with evaporites characterize the sequence within the axial part of Sverdrup Basin. On the south and east margins sandstone and conglomerate dominate. However a substantial unit of sandstone and conglomerate also forms the base of the sequence in the axial part, and tongues of limestone extend into the marginal areas at several levels. For purposes of description the Bashkirian to Guadalupian sequence is divided into two major units; an older unit extending from the Bashkirian to the early Artinskian and a younger unit of late Artinskian and Guadalupian rocks. In the axial part of Sverdrup Basin the whole sequence is essentially concordant. There appears to be, however, a widespread disconformity within the Artinskian and this permits the discrimination of the two major units. A contemporary break is also encountered on the south and east margins of the basin. On the margins, however, there is clearly more than one break in the succession because there are overlapping formations underlain by disconformities in the Sakmarian and Guadalupian as well as in the late Artinskian.

The Upper Carboniferous and Permian formations have been dated by their fusulinacean and ammonoid faunas. Thorsteinsson has determined the fusulinaceans, W. W. Nassichuk the ammonoids. Study of these faunas is incomplete but the following provides an indication of the scope of the faunas and of the way in which the stage names are interpreted.

The oldest rocks are dated as Bashkirian (Morrowan). Rocks of this age have been positively identified on the basis of ammonoids only in reef limestone masses brought to the surface in piercement domes on Axel Heiberg and Melville Islands. Faunas of Moscovian and younger age are widely distributed. Lower Moscovian (Atokan) is proved by the presence of *Profusulinella* and *Eofusulina* and also a rich ammonoid fauna that includes *Winslowoceras*. Younger Moscovian beds are recognized by a zone with *Wedekindellina* and a second zone with *Fusulina*. The Gzhelian (Zhigulevskian) Stage is characterized by *Quasifusulina, Triticites,* and *Fusulinella;* Orenburgian by schwagerinids that indicate the Russian Zone of *Pseudofusulina*; Asselian by characteristic species of *Schwagerina*; Sakmarian by a zone with *Monodiexodina* and *Pseudoschwagerina*. Two fusulinacean zones are recognized in the lower part of the Artinskian; the lower with *Parafusulina,* the upper with *Schwagerina hyperborea* (Salter). The later Artinskian and Guadalupian rocks lack fusulinaceans and are dated by ammonoids that represent at least three stratigraphic levels. The older two, characterized successively by *Metalegoceras* and *Spirolegoceras* are dated as Artinskian; the youngest ammonoid, *Neogeoceras,* provides evidence for the recognition of Guadalupian strata.

Lower Carboniferous

The Emma Fiord Formation rests unconformably on folded rocks of the Franklinian Geosyncline and is the oldest in the Sverdrup Basin (Kerr and Trettin, 1962). Outcrops are sporadic and occur only in a belt extending from Svartevaeg on northern Axel Heiberg Island to M'Clintock Inlet in northern Ellesmere Island. The formation includes at least 1,000 feet of dark grey carbonaceous siltstone and fine-grained sandstone, variable shaly, with minor light grey sandstone, and thin coal seams with fossil plants and a microflora of Viséan age. The upper contact with the younger Carboniferous strata is invariably a disconformity.

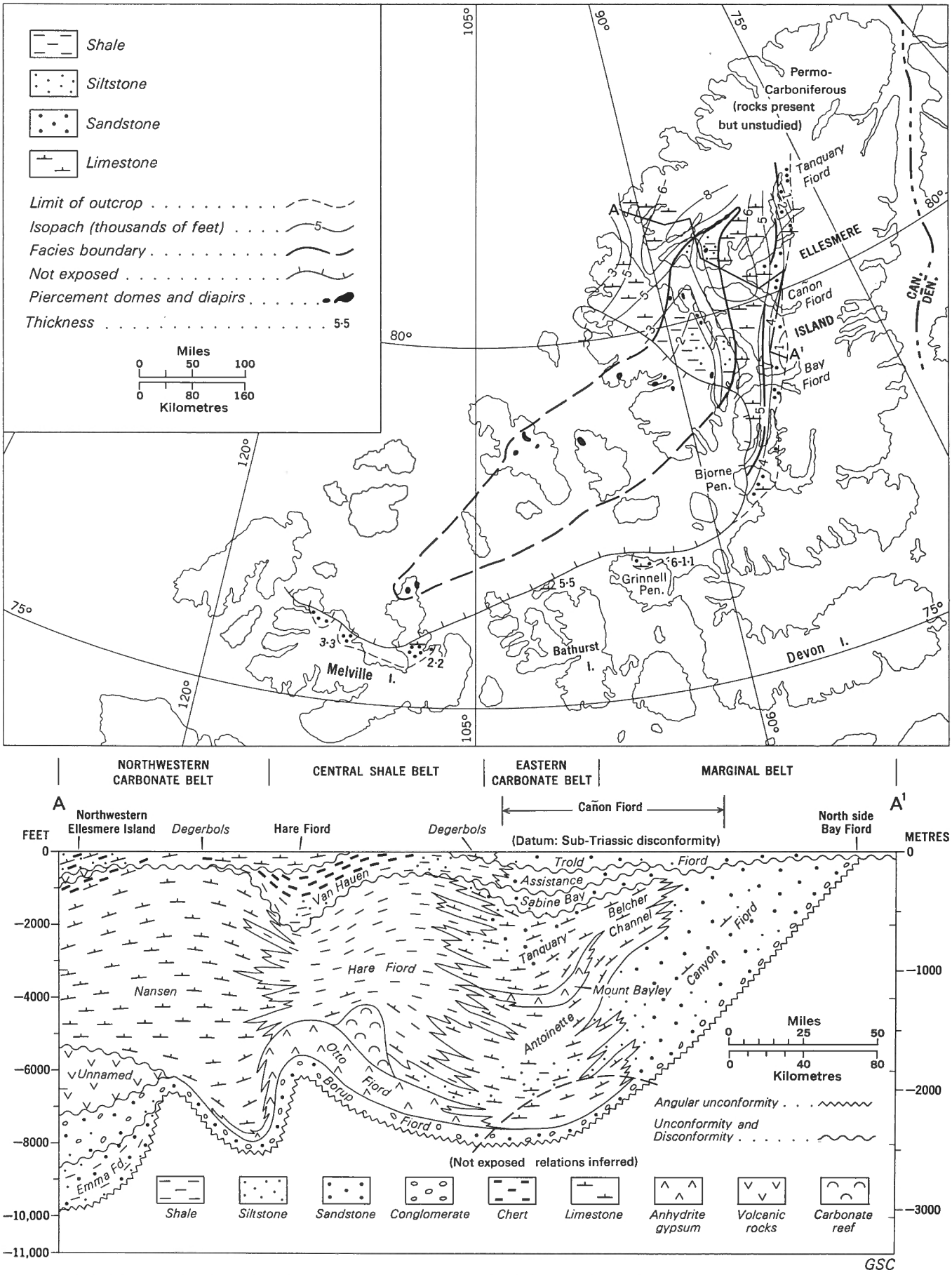

FIGURE X-8. Distribution and restored section of Pennsylvanian and Permian rocks in Arctic Archipelago (by R. Thorsteinsson).

Upper Carboniferous and Early Permian

The Bashkirian, Moscovian, Orenburgian, Gzhelian, Asselian, and Sakmarian Stages, and the early part of the Artinskian, are well represented in the Sverdrup Basin by an essentially conformable succession. Higher in the Permian, within the Artinskian, there is an unconformity throughout the basin. The late Artinskian and Guadalupian rocks that occur above this unconformity are described later.

The basal beds of the Bashkirian to early Artinskian sequence, virtually everywhere, are red conglomerate and sandstone. These rocks are represented by the Borup Fiord Formation and by the lower conglomeratic part of the Canyon Fiord Formation. The higher and greater part of the Bashkirian–early Artinskian succession is developed in four facies belts. These belts are arranged parallel to the northeasterly trending axis of the Sverdrup Basin (Fig. X-8a). From southeast to northwest the belts are as follows: (1) a marginal facies belt developed on the south and east sides of the Sverdrup Basin characterized by the clastic and carbonate rocks, respectively the Canyon Fiord and Belcher Channel Formations; (2) an eastern carbonate belt, in which the principal rock units are the Antoinette, Mount Bayley, Tanquary, and Belcher Channel Formations; (3) a central shale belt where the rocks known as the Hare Fiord Formation are mainly shale and siltstone; (4) a northwestern carbonate belt, in which the Nansen Formation is the principal rock-unit. Evaporitic rocks occur in the eastern carbonate, central shale, and northwestern carbonate belts. The eastern and northwestern carbonate belts merge east of the head of Hare Fiord in northern Ellesmere Island. The marginal facies belt with both clastic and carbonate formations marks the southern and eastern margins of the Sverdrup Basin, and is exposed as a narrow band of outcrops traceable from Melville Island to the north coast of Ellesmere Island, a distance of about 1,000 miles. Extensions of the eastern carbonate belt, central shale belt, and the northwestern carbonate belt probably underlie the central and western parts of the Sverdrup Basin.

The Borup Fiord Formation is a heterogeneous, dusky red weathering assemblage of conglomerate, conglomeratic sandstone, and sandstone, with minor siltstone and shale, that form the basal beds of the sequence in the central shale belt and adjacent carbonate belts. It varies in thickness from zero to about 1,300 feet. The formation is widely exposed in the northern parts of Axel Heiberg and Ellesmere Islands, and grades laterally into the basal conglomerate of the Canyon Fiord Formation of the marginal facies belt. The Borup Fiord lies either with disconformity upon the sporadically distributed Emma Fiord Formation or with angular unconformity upon older rocks. Throughout much of their extent the Borup Fiord and Canyon Fiord Formations constitute the oldest rocks of Sverdrup Basin. Borup Fiord lacks diagnostic fossils but is dated tentatively as Bashkirian.

The Otto Fiord Formation occurs in the central shale belt and the adjacent carbonate belts. It is composed mainly of medium- to thick-bedded anhydrite and minor interbeds of limestone, and ranges in thickness from zero to nearly 1,200 feet. The contact with the underlying Borup Fiord Formation is gradational. In the axial region of Sverdrup Basin it is approximately coextensive with, though somewhat more extensive than, the immediately younger Hare Fiord Formation. It grades laterally into the basal limestone beds of the Nansen Formation northwest and north, and into the basal limestones of the Antoinette Formation to the southeast.

The Nansen Formation typically developed in the northwestern carbonate belt is a uniform succession of limestone, 4,000 to more than 7,000 feet thick. It is widely exposed in northern Axel Heiberg Island and northwestern Ellesmere Island. It is also exposed at the edge of the marginal facies belt on Raanes Peninsula in central Ellesmere Island where the upper beds are reefoid. The basal contact of the Nansen Formation is variable. In the vicinity of facies change with correlative rocks of the Hare Fiord Formation, the Nansen gradationally overlies the Otto Fiord evaporites, but where the evaporites are missing it gradationally overlies the Borup Fiord Formation (Fig. X-8b). At the northwestern extremity of Ellesmere Island about 1,800 feet of volcanic flows and pyroclastic rocks is intercalated between the Nansen and Borup Fiord Formations. The Nansen Formation evidently attains its greatest thickness where it grades laterally into the shale of the Hare Fiord Formation. In such areas, for example, reefoid development is common in the lower part of the Nansen Formation. On Raanes Peninsula the Nansen rocks contain much interbedded quartzose sandstone and chert is important in the upper part on Axel Heiberg Island. The Nansen contains abundant fusulinaceans that range in age from Lower Moscovian to early Artinskian.

The eastern carbonate belt is best represented between Bay and Tanquary Fiords where the characteristic succession is four, or locally, three formations. From oldest to youngest these are: the Canyon Fiord Formation which consists mainly of reddish terrigenous clastic rocks, and includes conglomerate like that in the Borup Fiord Formation. It ranges from Bashkirian or Moscovian to early Orenburgian. The Antoinette Formation consists mainly of limestone and sandstone, 1,400 to over 2,600 feet thick, intertonguing with and overlying the Canyon Fiord Formation. Where overlying the Canyon Fiord it is late Orenburgian to Asselian but at East Cape in Canyon Fiord it includes equivalents of the Canyon Fiord Formation ranging in age from Moscovian to Asselian. The Mount Bayley Formation includes up to 820 feet of anhydrite with minor limestone. It lies gradationally on the Antoinette Formation and is probably Asselian. The overlying Tanquary Formation comprises 800 to 2,200 feet of Sakmarian to early Artinskian limestone and sandstone. The formation is everywhere underlain by the Mount Bayley Formation, the contact being gradational.

The Hare Fiord Formation, the characteristic unit of the central shale belt, includes 1,000 to more than 4,000 feet of beds, mainly grey siltstone and shale with lesser amounts of impure limestone and chert, resting gradationally on the Otto Fiord evaporites. Limestone bioherms, a few tens of feet to more than 1,500 feet thick, containing a brachiopod and ammonoid fauna of lower Moscovian age (Nassichuk, 1965) occur in the lower part. The non-reef parts of the Hare Fiord Formation contain fusulinacean and ammonoid faunas of Moscovian to early Artinskian age. The Hare Fiord Formation outcrops between Whitsunday Bay on Axel Heiberg Island and the head of Hare Fiord on Ellesmere Island, grading laterally into carbonate rocks. The conjectured distribution of the Hare Fiord Formation in southwestern regions of the Sverdrup Basin where Permian–Carboniferous rocks are hidden beneath younger rocks (Fig. X-8a) is based upon extrapolation of the geological relationships evident on Axel Heiberg and Ellesmere Islands and the occurrence of anhydrite diapirs which are derived from the Otto Fiord Formation. Inclusions of reef-like limestone in the anhydrite contain ammonoids of Bashkirian age. The large intrusions occur only where the Otto Fiord evaporites are overlain by incompetent siltstone and shale beds of the Hare Fiord Formation, for example on Svartefjeld Peninsula. In northwest Axel Heiberg Island where Otto Fiord evaporites are overlain by competent carbonate beds of the Nansen Formation only small bodies of intrusive anhydrite are found. Furthermore, the Mount Bayley Formation which outcrops in western Ellesmere Island is lithologically similar to the Otto Fiord Formation and the two formations attain comparable maximum thickness, but the Mount Bayley rocks are invariably overlain by competent limestone and sandstone beds and do not form diapirs.

The sequence of formations in the marginal facies belt is particularly variable owing to the presence of several disconformities and an angular unconformity. The lower beds are clastics of the Canyon Fiord Formation and the upper parts are carbonates of the Belcher Channel Formation. The latter limestones are, in places, entirely represented by clastics inseparable from the Canyon Fiord Formation and accordingly are included in it. On Ellesmere Island the Canyon Fiord Formation consists mainly of reddish sandstone and conglomerate with thin beds of limestone. It is thickest, at least 5,200 feet, in central Canyon Fiord and contains Moscovian to Sakmarian fusulinaceans, the upper part here being equivalent to the Belcher Channel carbonates. Between Canyon and Bay Fiords the thickness is much less mainly the result of bevelling prior to the deposition of the Lower Cretaceous Isachsen Formation. Farther south near the head of Trold Fiord the formation is 1,750 feet thick and ranges from Moscovian to Orenburgian. At least 1,000 feet of Canyon Fiord beds are present north of Eids Fiord.

On Melville Island the Canyon Fiord rocks occur in two tectonic situations. On eastern Melville Island about 2,000 feet of sandstone and conglomerate, with Bashkirian or Moscovian brachiopods, form the lower part of the Sverdrup Basin sequence, and are overlain concordantly but disconformably by Permian rocks. On northwestern Melville Island, south of Sverdrup Basin, Canyon Fiord rocks are preserved in isolated synclines. South of Raglan Range about 4,000 feet of sandstone, conglomerate, and minor limestone with Moscovian fusulinaceans occurs, but a few miles to the north, on the north side of the range, these rocks are absent and the Guadalupian Trold Fiord Formation rests directly on the Devonian. The present isolation of the Canyon Fiord rocks is partly related to the Melvillian disturbance described later, but it is also probable that some Canyon Fiord rocks of northwestern Melville Island were deposited in depressions separate from Sverdrup Basin.

The Belcher Channel Formation which outcrops throughout much of the marginal facies belt consists mainly of limestone, in part quartzose and bioclastic, and minor sandstone. The formation is most complete on Ellesmere Island between Bjorne Peninsula and Canyon Fiord, where 1,000 to 1,900 feet is present, resting with gradational contact on the Canyon Fiord Formation, and ranging in age from Orenburgian to early Artinskian. On Grinnell Peninsula it includes about 600 feet of Sakmarian and early Artinskian limestone which lies either with disconformity on the Canyon Fiord Formation, or with angular unconformity on rocks of lower Paleozoic age (Nassichuk, 1965). On Helena Island the formation includes some 600 feet of beds of uncertain age resting with angular unconformity on lower Paleozoic rocks. Belcher Channel rocks in northeastern Melville Island disconformably overlie the Canyon Fiord Formation and are entirely early Artinskian.

The sub-Belcher Channel disconformity is apparently responsible for the absence of Canyon Fiord rocks from northern Bathurst Island and parts of Grinnell Peninsula. It is probable that this disconformity reflects late Paleozoic movement of Boothia Uplift.

Melvillian Disturbance

In at least two areas on the margin of Sverdrup Basin folding and faulting termed herein the Melvillian Disturbance took place during the interval between deposition of Upper Carboniferous and Upper Permian rocks. The effects are best displayed in northwest Melville Island between Raglan Range and Canrobert Hills. There, the Upper Carboniferous Canyon Fiord Formation was folded along the same east–west axis first established during the Ellesmerian Orogeny (Pl. X-4). Nearby the Guadalupian Trold Fiord Formation rests unconformably upon the Canyon Fiord rocks and shows no sign of Melvillian folds. At the head of Trold Fiord, Ellesmere Island, an angular unconformity separates the Canyon Fiord and Trold Fiord Formations; the Upper Carboniferous rocks were faulted against Ordovician limestones prior to deposition of the Permian beds.

Late Permian

The Permian rocks representing the late Artinskian and Guadalupian are approximately co-extensive with those of the Bashkirian to early Artinskian interval, but parts of the late Permian succession are transgressive. Two regional facies belts may be distinguished. The basinal facies is composed of fine-grained clastics and chert; the van Hauen Formation is overlain by carbonates and chert of the Degerbols Formation. The marginal facies, indicating near-shore conditions on the south and east edges of Sverdrup Basin, consists of glauconitic sandstone divided into the Sabine Bay, Assistance, and Trold Fiord Formations. Regional disconformities mark the base and top of most of these formations, and at various localities one or more may have been removed by erosion, so that various formational sequences occur. All the formations except the Sabine Bay are entirely marine. The marine beds contain abundant brachiopods, bryozoans, and rare ammonoids. Fusulinaceans so abundant in the underlying Moscovian–early Artinskian rocks are completely unrepresented in these younger Permian faunas.

The Sabine Bay Formation consists of mainly nonmarine, light coloured sandstone, conglomeratic sandstone, and conglomerate of late Artinskian age (Tozer and Thorsteinsson, 1964; Nassichuk, 1965). On northeastern Melville Island where the thickness varies between 20 and 400 feet it lies between the Belcher Channel and Assistance Formations. On western Ellesmere Island the Sabine Bay is exposed in a narrow belt from south of Canyon Fiord to Tanquary Fiord varying in thickness from a feather edge to 640 feet and resting on the Belcher Channel Formation or the Tanquary Formation. The formation is overlain by either the Assistance or the Trold Fiord Formation. Sabine Bay rocks are believed to pinch-out towards the northwest, and to have no equivalents within the basinal facies of Sverdrup Basin.

The Assistance Formation is composed principally of greyish sandstone and siltstone, variably glauconitic, commonly poorly consolidated, replete with brachiopods. It ranges in thickness from zero to 1,200 feet and contains late Artinskian ammonoids. It is distributed discontinuously along south and east margins of the Sverdrup Basin, from northeastern Melville Island to Greely Fiord on Ellesmere Island and is apparently equivalent to the van Hauen Formation of the basinal facies.

The Trold Fiord Formation is mainly glauconitic sandstone with minor chert and limestone, variably consolidated and as much as 1,000 feet thick. Fossils are abundant, mainly brachiopods and at one locality the Guadalupian ammonoid *Neogeoceras* (Nassichuk, *et al.*,



PLATE X-5. Lower Devonian Eids Formation (E), folded during the Ellesmerian Orogeny unconformably overlain by 100 feet of Permian glauconitic sandstone of the Trold Fiord Formation (T) and the Lower Triassic Bjorne Formation (B). View north of Trold Fiord, Ellesmere Islands.

1965). It transgresses south and southeast, and overlies several formations: the Sabine Bay, Assistance, and Canyon Fiord, and the lower Paleozoic (Pl. X-5).

The van Hauen Formation of the basinal facies belt is most commonly black siltstone and shale, variably calcareous and non-calcareous, and up to 2,460 feet thick. Bedded black chert, apparently an alteration product of siltstone, forms a conspicuous constituent in the upper part of some sections. Van Hauen rocks have not yielded diagnostic fossils but are dated as late Artinskian because they grade laterally into the Assistance Formation on Bjorne Peninsula of Ellesmere Island.

The Degerbols Formation, the youngest Permian formation of the basinal facies, comprises up to 1,100 feet of predominantly cherty and fossiliferous limestone. The formation outcrops extensively in northern Axel Heiberg and Ellesmere Islands where it commonly rests on the van Hauen Formation. It appears to be at least approximately the same age as the Trold Fiord Formation and is therefore tentatively dated as Guadalupian. In some places, however, the Degerbols rests directly on the Nansen Formation, and it would appear that the van Hauen beds are truncated. The Degerbols of northern Axel Heiberg Island is mainly light coloured, fossiliferous chert, with minor siltstone. This chert is apparently altered limestone. About 120 feet of basaltic flows and pyroclastic rocks are locally present between the Nansen and Degerbols Formations on northeast Axel Heiberg Island. On southern Ellesmere Island and Melville Island the marginal and basinal sequences evidently intertongue as bioclastic limestone of the Degerbols type is associated with the marginal sandstone formations. On Bjorne Peninsula at least 465 feet of white limestone overlies the Assistance Formation, this limestone forms the classical outcrops of Great Bear Cape, discovered by Per Schei. A similar limestone on Sabine Peninsula, Melville Island, lies between beds of Assistance and Trold Fiord facies (Nassichuk, 1965).

MESOZOIC

In the Arctic Archipelago the principal site of Mesozoic sedimentation was Sverdrup Basin (Pl. X-6). Cretaceous rocks occur south and east of the basin as outliers resting on the folded rocks of the Franklinian Geosyncline, and they also form an extensive cover over the little disturbed Paleozoic rocks on Banks Island of the western Arctic Lowlands.

In the axial part of Sverdrup Basin Mesozoic strata from the Triassic to the Upper Cretaceous are structurally conformable as are the Permian and Triassic rocks. Dis-



PLATE X-6. Mesozoic rocks north of Blaa Mountain, Ellesmere Island, folded during the Eurekan Orogeny. Formations are: Triassic, H–Heiberg; Jurassic, B–Borden Island; S–Savik; A–Awingak; Lower Cretacous, I–Isachsen; dark cuesta capped by Cretaceous basalt. Recessive Jura-Cretaceous Deer Bay Formation, not visible, occurs between Awingak and Isachsen Formations.

conformities occur however beneath the Triassic and at several levels in the Jurassic. On the margins many formations overstep. Such known formations are: Triassic: Karnian (Schei Point), Norian (Heiberg); Jurassic: Sinemurian (Borden Island), Toarcian (Wilkie Point), and Volgian (Mould Bay); mid-Lower Cretaceous (Isachsen).

During the Mesozoic the archipelago was unaffected by orogeny. In many areas the rocks are folded, but the folding took place during Eurekan Orogeny in Cenozoic time.

The Mesozoic rocks consist mainly of fine- to medium-grained quartzose sandstone, siltstone, and shale. Conglomerate is volumetrically unimportant. Basalt flows are interbedded with Cretaceous sedimentary rocks in northern Sverdrup Basin. Marine and non-marine sediments alternate. Commonly the marine Mesozoic formations consist of shale in central Sverdrup Basin, with contemporaneous sands on the south and east margins. The non-marine beds are frequently crossbedded, ripple-marked, and contain coal at several horizons. Shallow-water features are also common in the marine formations. It follows that most of the sediment was deposited in relatively shallow water, upon a surface characterized by low gradients. The non-marine beds were presumably deposited on deltas. During the three principal periods of deltaic development (Triassic Heiberg Formation, Jurassic Awingak Formation, and Lower Cretaceous Isachsen Formation) thin, widely distributed marine beds are interbedded with the non-marine sediments, indicating periodical marine transgressions over the deltas. The wide distribution of these marine beds indicates that the deltas were of very low relief.

The volume of Mesozoic sediment originally deposited in the Sverdrup Basin was large, probably more than 150,000 cubic miles. The boundary between the shale and sand facies belts suggests that the source of most of this sediment lay east and south of the basin. Minor amounts of sediment were derived from the north in Lower Triassic time and from the north or northwest in the Upper Triassic (Karnian) and possibly also the Middle Jurassic (Callovian). The Lower Triassic rocks of northern origin were probably derived by uplift of the metamorphic complex of northern Axel Heiberg Island. These sources were probably volumetrically unimportant, the main source of the great volume of clastic sediments, mainly fine-grained and commonly quartzose, lay to the south and west; the precise provenance is uncertain. Some sediments may have been derived from the denudation of the folded Paleozoic rocks. However, the area embraced by the Franklinian Geosyncline was largely denuded by Upper Carboniferous time. The large volume and fine-grained nature of the Mesozoic sediment suggest that much came from distant sources, such as the west slope of the East Greenland Caledonides or the west slope of the Appalachian Mountains. Sediment from the latter would have been transported by a major drainage system traversing the Canadian Shield. Such a drainage system would be analogous to the Missouri–Mississippi System of today, feeding sediment from distant uplifted areas, across extensive lowlands, into the Gulf of Mexico.

The distribution of sand and conglomeratic beds in the Triassic, Jurassic, and early Lower Cretaceous marine formations (up to the Valanginian) suggests that the shorelines during these periods were not far from the east and south limits of their outcrop. This is also suggested by the complete absence of marine Triassic and Jurassic rocks south and east of Sverdrup Basin. The Triassic, Jurassic, and early Lower Cretaceous faunas are closely related to contemporary assemblages in Spitsbergen, East Greenland, and northern Siberia. These faunas presumably entered Sverdrup Basin from the northwest, and their close relationship with those of other arctic areas suggests that a marine area, probably an oceanic basin, existed on the site of the Arctic Ocean during Triassic, Jurassic, and Cretaceous time, as has been suggested by Diener, Haug, Arkell, Frebold, and Sachs.

The Albian transgression in the arctic islands was much more widespread, but, as in earlier Mesozoic time, the Albian seas probably did not extend far beyond the southeast limit of Sverdrup Basin, although they did extend beyond the basin on Melville and Banks Islands, and were undoubtedly continuous with the Albian sea that covered western Canada.

In the Upper Cretaceous at the time of deposition of the Kanguk Formation in Sverdrup Basin marine strata were deposited not only in western Canada but also on the east shore of Baffin Bay, Greenland. The Upper Cretaceous rocks of west Greenland nonconformably overlie Precambrian rocks and they are the oldest and only marine Mesozoic sediments in the Baffin Bay area. For this period it is possible, as suggested by Teichert and Birkelund, that there was a direct marine connection between the region of Sverdrup Basin and Baffin Bay. The connection could probably have been in the vicinity of southern Ellesmere Island where Cenozoic non-marine beds apparently rest unconformably upon Paleozoic formations. It is possible that Upper Cretaceous rocks were originally present, but removed prior to the deposition of the Cenozoic strata.

Triassic

Triassic rocks of the Sverdrup Basin are exposed on Ellesmere, Axel Heiberg, Cornwall, Table, Exmouth, Cameron, Melville, Prince Patrick, Brock, and Borden Islands. Their distribution and facies are shown on Figures X-9 and X-10.

Permian–Triassic Boundary

In the axial part of Sverdrup Basin and in parts of the marginal area Lower Triassic rocks rest disconformably upon Permian strata. In places the younger Permian rocks are Guadalupian; elsewhere they are late Artinskian. Latest Permian (Dzhulfian or Ochoan) strata are unknown

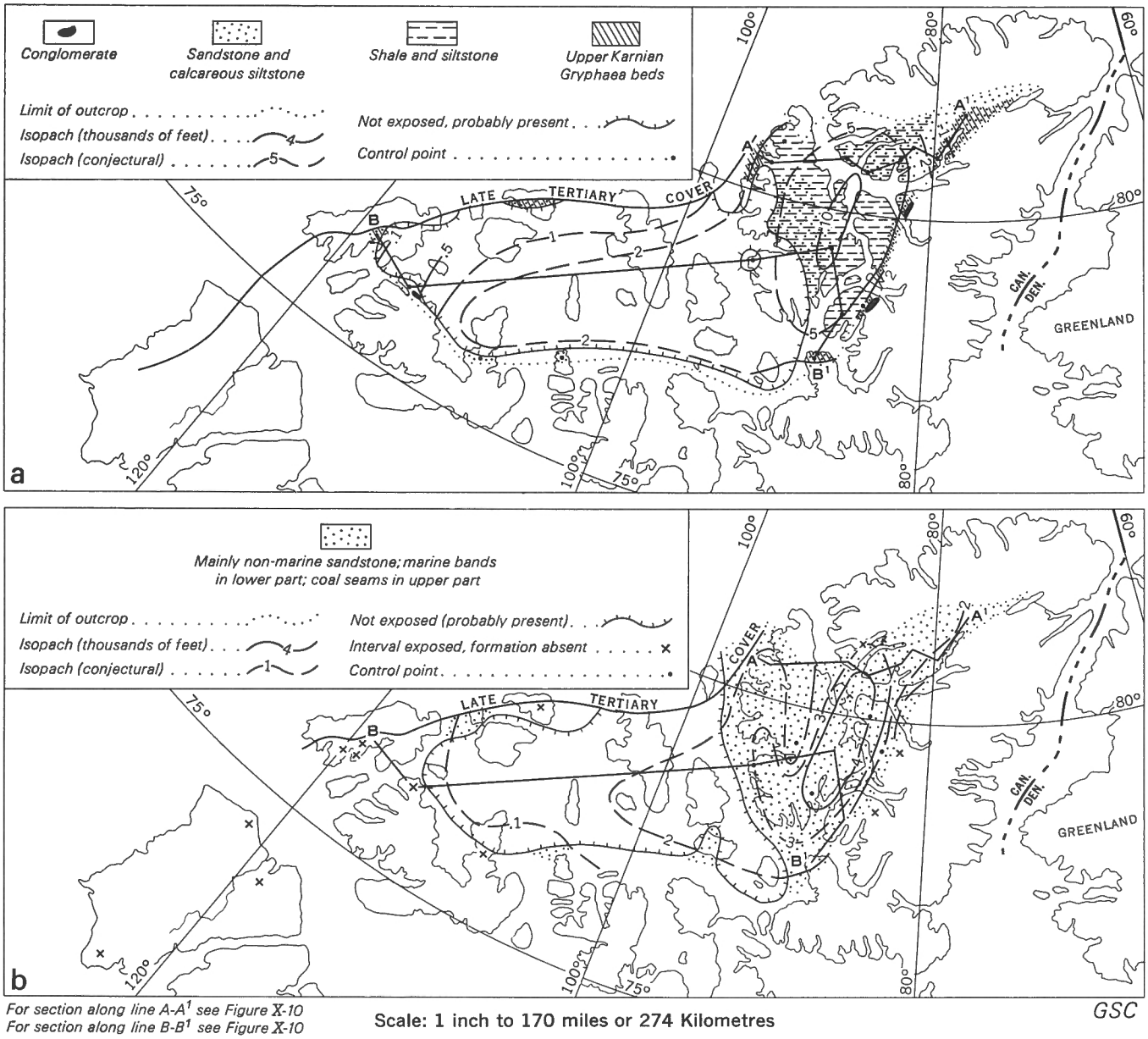

FIGURE X-9. Distribution of Triassic rocks in Arctic Archipelago; (a) Griesbachian to Karnian, (b) Norian, Heiberg Formation (by E. T. Tozer).

in the archipelago. The oldest Triassic beds are Lower Griesbachian, as old as any Triassic rocks known. The contact between the Permian and Triassic rocks is invariably well marked by lithological change but not marked by physical evidence of erosion. From the available evidence it appears that the Permian–Triassic boundary is paraconformable, marks a hiatus, and that the gap in the sedimentary record is within the latest Permian and not in the Triassic.

Marine Rocks

Marine conditions prevailed throughout the axial part of Sverdrup Basin from the earliest Triassic (Griesbachian) to the early Upper Triassic (Karnian). Some of the Lower Triassic beds on the margins were probably deposited in a non-marine environment. For this interval the sections on the margins of Sverdrup Basin differ from those of the axis both in thickness and lithology. Most of the rock on the south and east margins consists of sandstone and calcareous siltstone (Bjorne and Schei Point Formations); in the axial part it is shale and siltstone (Blind Fiord and Blaa Mountain Formations). On the northwest margin of the basin the sections are of mixed character. For Griesbachian, Smithian, Spathian, Anisian, and Ladinian the rocks are mainly shale and siltstone but in the Dienerian and Karnian the sections include sandstone and conglomerate.

The Bjorne Formation disconformably overlies Permian strata and consists mainly of quartzose, commonly crossbedded sandstone, with conglomeratic interbeds on

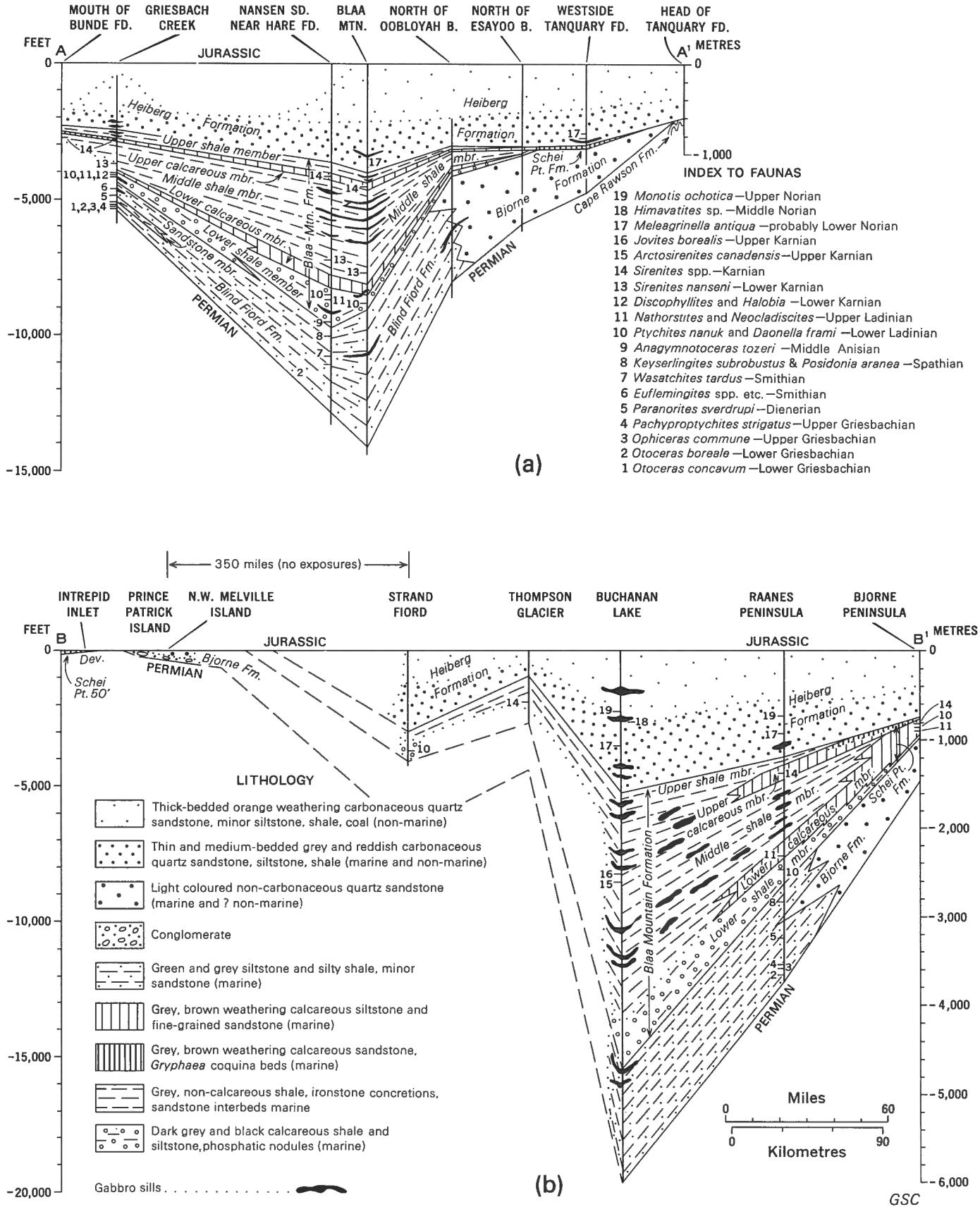

FIGURE X-10. Restored sections, Triassic formations of Arctic Archipelago; (a) along lines A–A', (b) along lines B–B' of Figure X-9 (by E. T. Tozer).

the extreme margins of the basin, for example in Canyon Fiord and northeast of Trold Fiord on Ellesmere Island, and on Melville Island. Bjorne rocks weather bright red, orange, yellow, brown, and white. Fossils are rare, but *Otoceras* (Griesbachian) occurs near the base at Trold Fiord, and poorly preserved younger Lower Triassic ammonoids are known from other localities. Marine fossils are unknown in the beds on the south margin of Sverdrup Basin. The Bjorne is overlain by Anisian strata and is thus fairly well dated as Lower Triassic. The contemporary rocks exposed on the western coast of Ellesmere Island and on eastern and northern Axel Heiberg Island are the Blind Fiord Formation, which consists of green and grey siltstone, fine-grained sandstone, and shale. The Blind Fiord beds are well dated by ammonoid faunas. On Raanes Peninsula, eastern Axel Heiberg Island, and near Hare Fiord, they span the Lower Triassic, from the Griesbachian to the Spathian. Around Otto Fiord and on northern Axel Heiberg Island the uppermost beds are Smithian and the boundary between the Lower and Middle Triassic lies within the overlying Blaa Mountain Formation. On northern Axel Heiberg Island a sandstone member of Dienerian age occurs within the Blind Fiord Formation. This member consists of hard green and red sandstone, with some conglomerate and beds of *Myalina* coquina. The sandstones of this member become increasingly fine grained towards the south.

The Schei Point and Blaa Mountain Formations are essentially contemporaneous and of Anisian, Ladinian, and Karnian age. Around Otto Fiord and on northern Axel Heiberg Island latest Lower Triassic (Spathian) beds are included within the Blaa Mountain Formation. Typical Schei Point rock is grey, brown weathering, highly calcareous siltstone and fine-grained sandstone, with bioclastic layers composed of brachiopod and bivalve shells in the upper part. Shale is the typical Blaa Mountain lithology. Two prominent tongues of Schei Point lithology interfinger with the Blaa Mountain shales throughout western Ellesmere Island and on northern Axel Heiberg Island. These tongues divide the Blaa Mountain Formation into five members, in ascending order: the Lower Shale, Lower Calcareous, Middle Shale, Upper Calcareous, and Upper Shale. The Lower and Upper Calcareous Members are tongues of Schei Point facies. The Lower Shale is commonly black and calcareous, with calcareous and phosphatic concretions. The higher shales are non-calcareous, commonly with red weathering clay ironstone concretions. Shales equivalent to the Lower and Middle Shale Members are recognizable in Schei Point sections at some localities; notably on Fosheim and Raanes Peninsulas.

Throughout central Ellesmere Island, the top of the Schei Point Formation is marked by the Upper Karnian "Gryphaea Bed": 100 feet of calcareous sandstone with coquinoid layers of *Gryphaea* and *Plicatula,* which on Raanes Peninsula and north of Greely Fiord can be traced laterally into the Upper Calcareous Member of the Blaa Mountain Formation. This bed is of unusual interest. In Tanquary Fiord it oversteps the Lower Triassic Bjorne Formation. Comparable beds are also present on the west side of the Sverdrup Basin, on Prince Patrick and Borden Islands. At Intrepid Inlet, Prince Patrick Island, the Schei Point *Gryphaea* bed rests unconformably upon Devonian rocks; on Borden Island it represents the oldest exposure. A similar and apparently contemporary *Gryphaea* bed in the Blaa Mountain Formation of northwestern Axel Heiberg Island occupies the position of the Upper Calcareous Member. Upper Karnian *Gryphaea* beds are thus widely distributed on the margins of the Sverdrup Basin and their occurrence probably indicates an interval of shoal water conditions along the border of Sverdrup Basin following uplift.

The Blind Fiord and Blaa Mountain Formations of the axial part of Sverdrup Basin were laid down some considerable distance from shore. In the Triassic there were two sources of sediment, one essentially continuous feeding sediment to the south and east margins of the basin, and the other an intermittent source (or sources) providing sediment to the north and northwest margins of the basin. Formation of the Dienerian sandstone member was perhaps related to uplift of the lower Paleozoic massif of northern Axel Heiberg Island. The Upper Karnian *Gryphaea* beds of northern Axel Heiberg and Borden Islands probably reflect uplift of the marginal sill of Sverdrup Basin. Beyond the sill there was probably open Arctic Ocean, because the presence of the cosmopolitan marine faunas that entered Sverdrup Basin throughout Triassic time demands a circum-Arctic migration route. The postulated shoal or sill on the northwest margin of Sverdrup Basin does not appear at any time to have constituted a barrier to the movement of these faunas.

Upper Triassic Deltaic Rocks

The youngest Triassic formation in Sverdrup Basin is the Heiberg, which consists mainly of non-marine, carbonaceous sandstone, with marine beds at several levels in the lower part. The lower marine beds with *Meleagrinella antiqua* are probably Lower Norian, the higher marine strata include Middle Norian beds and strata with the cosmopolitan Upper Norian *Monotis ochotica.* The beds above those with *Monotis* are entirely non-marine, with fossil plants and thin coal seams. They may be partly Jurassic, but if so, they are not younger than Sinemurian. The Heiberg normally overlies the Blaa Mountain Formation or the Schei Point Formation, but at the head of Tanquary Fiord the Heiberg overlaps Cape Rawson beds.

The Heiberg Formation indicates that in late Triassic time the level of sedimentation in Sverdrup Basin rose above strandline, at first intermittently, later continuously. From the Griesbachian to the Karnian the Triassic depocentre apparently lay on eastern Axel Heiberg Island and in latest Triassic time there were at least two such centres, one on eastern and the other on western Axel Heiberg Island. They were separated by a structure on strike with the Lower Paleozoic massif at the north of the island.

Jurassic

Triassic–Jurassic Boundary

As already mentioned, it is possible that beds of earliest Jurassic age occur in the upper, non-marine part of the Heiberg Formation, but for this there is no proof. If the uppermost Heiberg beds are Jurassic they are not yet separable by lithological criteria from the lower, undoubtedly Triassic, part of the formation. The top of the uppermost Heiberg sandstone commonly forms a prominent, easily recognized bedding plane throughout the Sverdrup Basin. The next younger beds are marine Jurassic strata; in places they are Sinemurian, the Borden Island Formation; elsewhere they are Toarcian, the Savik, Wilkie Point, and Jaeger Formations. Marine Hettangian rocks are unknown and are probably absent. Both Sinemurian and Toarcian rocks overstep locally and in places a basal conglomerate occurs beneath Toarcian beds. The contact between the Heiberg and succeeding Jurassic formations is evidently a paraconformity.

Lower Jurassic (Sinemurian) Marine Rocks

The oldest known Jurassic rocks, the Sinemurian Borden Island Formation, are exposed on Borden and Melville Islands, also on eastern Axel Heiberg and western Ellesmere Islands (Fig. X-11). The formation comprises up to 200 feet of grey and green glauconitic sand and is particularly characterized by thin beds of hard, red, ferruginous sandstone. On Ellesmere and Axel Heiberg Islands the Borden Island beds overlie the Heiberg Formation. In the western islands they overlap, resting on Schei Point strata on Borden Island and upon the Bjorne Formation on Melville Island. The Borden Island Formation has not been recognized in the axial part of the Sverdrup Basin. Possibly the lowermost beds of the Savik shale are Sinemurian and equivalent to the Borden Island sands, but for this there is no evidence. The distribution of Sinemurian beds in the Sverdrup Basin is unusual, for throughout the Mesozoic it is the general rule that all beds on the margins of the basin are represented in the axial part; for the Sinemurian this does not apply. Rocks of the next Jurassic stage, the Pliensbachian, are unknown throughout the archipelago. Possibly Pliensbachian time was one of gentle uplift and erosion, and the present Sinemurian beds are remnants of a former more extensive deposit.

Later Jurassic (Toarcian–Oxfordian) Marine Rocks

Toarcian sand and sandstone on the margins of Sverdrup Basin are represented by shale in the axial region. The facies interfinger on Mackenzie King and Ellef Ringnes Islands and in the Eureka Sound area. The Lower Wilkie Point sands on Prince Patrick and Melville Islands and the Jaeger beds of Cornwall and western Ellesmere Islands are quartzose, grey and yellow or glauconitic and green. Hard red ferruginous beds and phosphatic nodules are common features. The shales of the Savik Formation on Axel Heiberg Island are light grey, argillaceous, and charged with clay ironstone concretions. The Oxfordian shales are black and silty.

The régime of marginal marine sand and axial shale deposition persisted from the Toarcian through the Bajocian and Bathonian into the Lower Callovian. Ammonite faunas of all these stages occur in the sands on the south side of the basin, and in shale on western Axel Heiberg Island.

Middle (?) and Upper Jurassic Deltaic Rocks and Marine Equivalents

In the Middle (?) and Upper Jurassic non-marine beds are widely distributed in the Sverdrup Basin (Fig. X-11a). In some areas marine strata are intercalated and the non-marine beds can be accurately dated; elsewhere they have been dated only in terms of the age of the underlying and overlying marine rocks.

On Prince Patrick and Melville Islands non-marine quartzose sands of the Upper Wilkie Point Formation lie between the Lower Bathonian *Arcticoceras* beds and the Lower Volgian Mould Bay Formation. On Cornwall Island comparable non-marine strata are assigned to the Awingak Formation which overlies Lower Callovian marine beds. The mainly non-marine Awingak Formation of Ellesmere and Axel Heiberg Islands has some marine intercalations. The formation consists of alternating units of hard pale coloured quartzose sandstone, and dark recessive silty shale and sandstone. On eastern Axel Heiberg Island the Awingak is underlain by Lower Oxfordian beds of the Upper Savik. In several areas, *Amoeboceras* of Upper Oxfordian or Lower Kimmeridgian age occurs within the formation, and on Fosheim Peninsula and near Li Fiord Lower Volgian *Buchia* beds occur in the uppermost strata. In Bunde Fiord of northwest Axel Heiberg Island the Awingak Formation is absent but *Amoeboceras*-bearing shales suggest that equivalents are present. The Awingak delta probably was formed of sediment derived from the south and east and northwest Axel Heiberg Island evidently lay beyond the edge of the delta.

Deltaic rocks are absent in the Jurassic section of Mackenzie King Island and probably on Ellef Ringnes Island. Whether the Jurassic deltas extended to these islands is not known, for disconformities probably occur in these sections.

Middle (?) and Upper Jurassic Unconformities and Disconformities

On the margins of Sverdrup Basin the Jurassic sequence is locally interrupted by unconformities and disconformities. The most profound is on southern Prince Patrick Island where the Lower Volgian Mould Bay Formation progressively truncates beds of the Wilkie Point Formation and eventually rests unconformably upon Devonian rocks. This unconformity probably reflects elevation of the Prince Patrick Uplift. On Mackenzie King Island, Oxfordian or Kimmeridgian shales of the Lower Mould Bay Formation rest directly but apparently con-

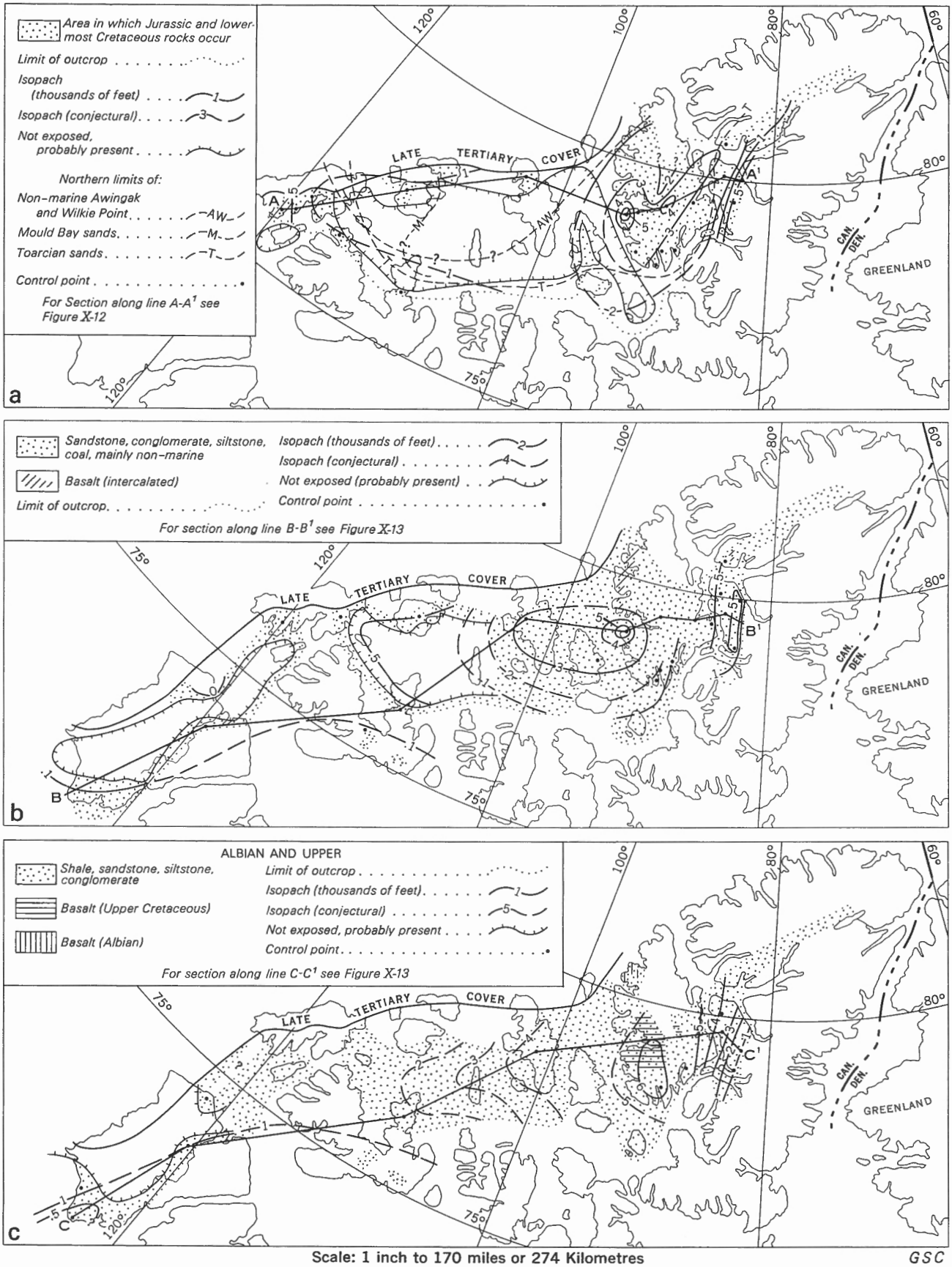

FIGURE X-11. Distribution of Jurassic and Cretaceous rocks in Arctic Archipelago; (a) Jurassic and lowermost Cretaceous, (b) Lower Cretaceous, Isachsen Formation, (c) Lower Cretaceous, Albian, and Upper Cretaceous (by E. T. Tozer).

formably upon Lower Bajocian beds of the Wilkie Point Formation. Upper Bajocian, Bathonian, and Callovian beds are unknown and apparently absent through disconformity.

On Ellef Ringnes Island no rocks are known that span the interval between the Lower Callovian and the Valanginian. Possibly such rocks are present, but the close proximity of Callovian sands to the Valanginian Deer Bay shale suggests that most or all of the Upper Jurassic is missing and that there is a disconformity.

At Buchanan Lake, eastern Axel Heiberg Island, the black, silty Upper Savik shales with Oxfordian *Cardioceras* rest directly on Lower Bajocian light grey clay shales of the Lower Savik. The lithological boundary between the Lower and Upper Savik is abrupt and probably indicates non-deposition during Upper Bajocian, Bathonian, and Callovian time. On Fosheim Peninsula the Lower Savik includes the Upper Bajocian or Lower Bathonian *Arkelloceras* beds, which have escaped truncation by the pre-Oxfordian disconformity that forms the lower boundary of the Upper Savik.

The unconformities presumably indicate positive activity on the margins of Sverdrup Basin. Probably the sections on southwestern Axel Heiberg Island are devoid of these unconformities as substantial thickness of unfossiliferous rock occurs between beds of Callovian and Volgian age.

Latest Jurassic and Earliest Cretaceous Marine Rocks

Sedimentation was widespread, the pattern of facies distribution (Fig. X-11a) resembling that for the Triassic and Jurassic. Sands and sandstone of the Mould Bay Formation occur on the southwest margin and grey shale and siltstone of the Deer Bay Formation in the axial area. Interfingering relationships are seen on Mackenzie King and southern Axel Heiberg Islands.

In the Deer Bay Formation, a homogeneous shale devoid of any evidence of physical breaks, ammonoid faunas of Upper Volgian, Berriasian, and Valanginian age indicate that sedimentation was continuous throughout the late Jurassic and early Lower Cretaceous. The Mould Bay Formation of Mackenzie King, Prince Patrick, and

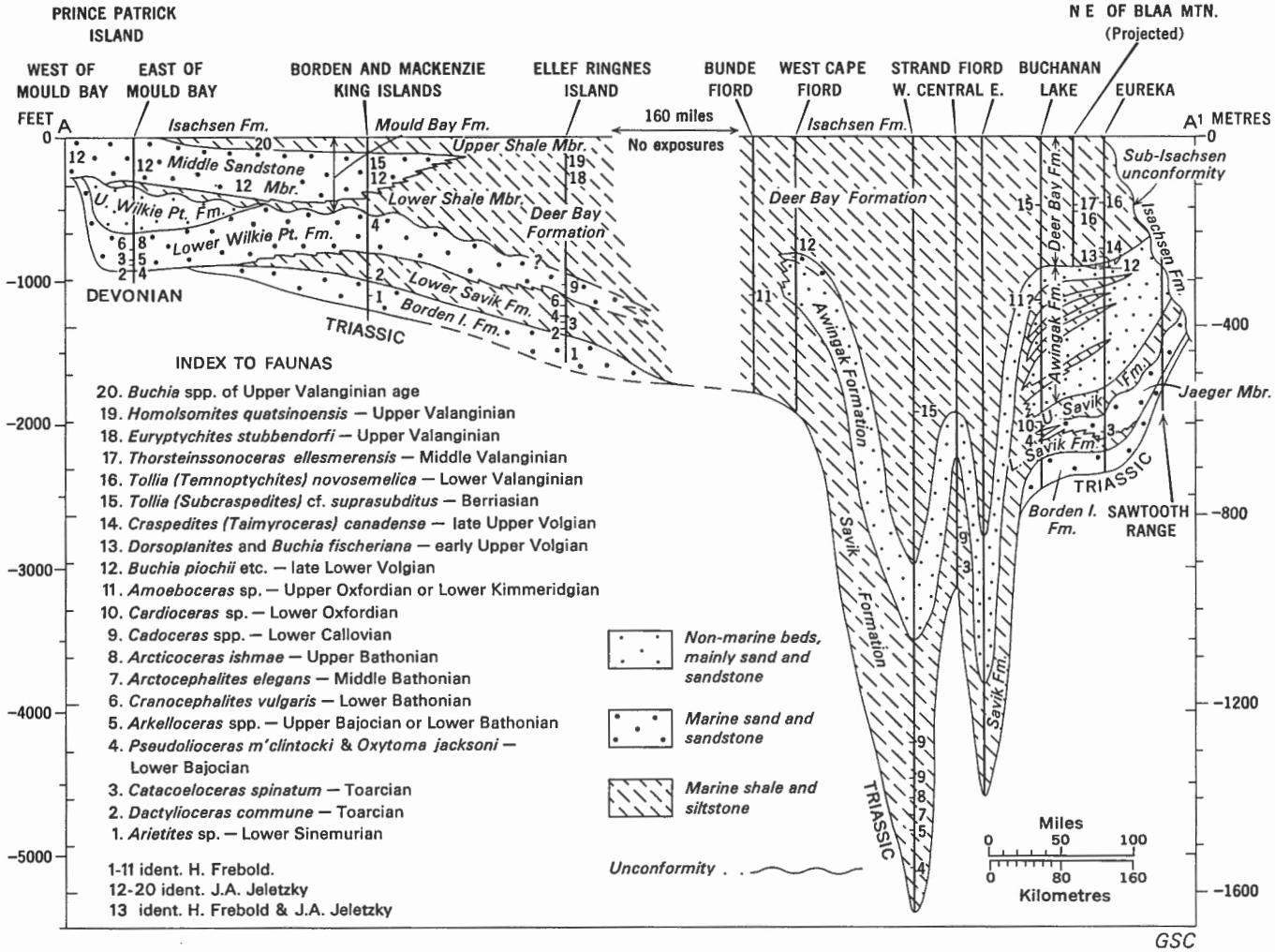

FIGURE X-12. Restored sections, Jurassic and lowermost Cretaceous rocks of Arctic Archipelago, along line A–A' of Figure X-11 (by E. T. Tozer).

Melville Islands includes strata older than any part of the Deer Bay Formation. In the south and southwest the beds are mainly sand with some conglomerate. Tongues of shale that thicken to the northeast also occur. The Lower Shale Member on Mackenzie King Island has Oxfordian or Kimmeridgian *Amoeboceras* and the Mould Bay beds of all areas contain *Buchia* species of Lower Volgian age. The fossil record for the Mould Bay Formation of Prince Patrick Island is far less complete than that for the Deer Bay shales for it includes only Lower Volgian and Upper Valanginian. Upper Volgian, Berriasian, and much of Valanginian time are not demonstrably represented. For this period, as in the earlier Mesozoic, the margins of the Sverdrup Basin were relatively resistant to subsidence and preserve an incomplete sedimentary record.

Cretaceous

Mesozoic rocks of Triassic to the lowermost Cretaceous are found only in Sverdrup Basin. Beginning with the Isachsen Formation, the later Cretaceous rocks are more widespread and occur not only in the basin but also as outliers unconformably overlying the folded Devonian rocks and as extensive tracts in western Arctic Lowlands, lying on rocks as old as Precambrian (Figs. X-11b and X-11c).

Lower Cretaceous

The mainly non-marine Isachsen Formation conformably overlies the Deer Bay Formation on Axel Heiberg and Ellef Ringnes Islands. The relationship is essentially transitional, as shown both by the concordance of the strata and the occurrence at Good Friday Bay and near Isachsen of marine Valanginian *Buchia* beds in the lower part. On the margins of Sverdrup Basin the Isachsen beds overlap and rest on various Paleozoic and Mesozoic rocks. On Ellesmere Island, on the west side of Fosheim Peninsula Isachsen beds overlie the Deer Bay shale abruptly and concordantly, but probably disconformably. Towards the east Isachsen beds rest on progressively older Mesozoic beds: on the Awingak Formation in Sawtooth Range; on the Heiberg in central Vesle Fiord and northwest of Trold Fiord; on the Schei Point at the head of Vesle Fiord and north of Trold Fiord; on the Bjorne Formation in Canyon Fiord. Finally, northeast of Vesle Fiord and around Strathcona Fiord, the Isachsen rests on Devonian and older Paleozoic rocks. Where the Isachsen rests on upper Paleozoic or Mesozoic beds the contact is structurally conformable. On the other hand, the Devonian and older rocks that underlie the Isachsen have been folded and the contact is one of angular unconformity. On the northwest edge of Sverdrup Basin the formation abruptly and probably disconformably overlies Mould Bay strata. Beyond Sverdrup Basin, Isachsen beds occur on Banks Island and southeast Melville Island. Coal-bearing beds that occur near Pond Inlet, northern Baffin Island, possibly represent an outlier of the Isachsen resting on Precambrian crystalline rocks.

The Isachsen Formation normally consists of light coloured, quartzose sandstone, grit and conglomerate, with interbeds of dark siltstone, shale, and coal. Crossbedding is common. Lower Cretaceous leaves and plant microfossils are known from several localities. A non-marine environment of deposition is indicated for all but the thin marine beds near the base. In northwestern Axel Heiberg Island basalts are interbedded. The exact age of the Isachsen is uncertain. In places the basalt beds are Valanginian and marine Albian, and possibly also Aptian beds overlie. The greater part is probably Hauterivian and Barremian, but Aptian beds may be present.

Following deposition of the Isachsen Formation, marine transgression led to deposition of the Christopher Formation within Sverdrup Basin and on southern Melville Island and Banks Island. Christopher rocks are mainly grey shale, with calcareous septaria and concretions and small amounts of siltstone and sandstone. On Elles-

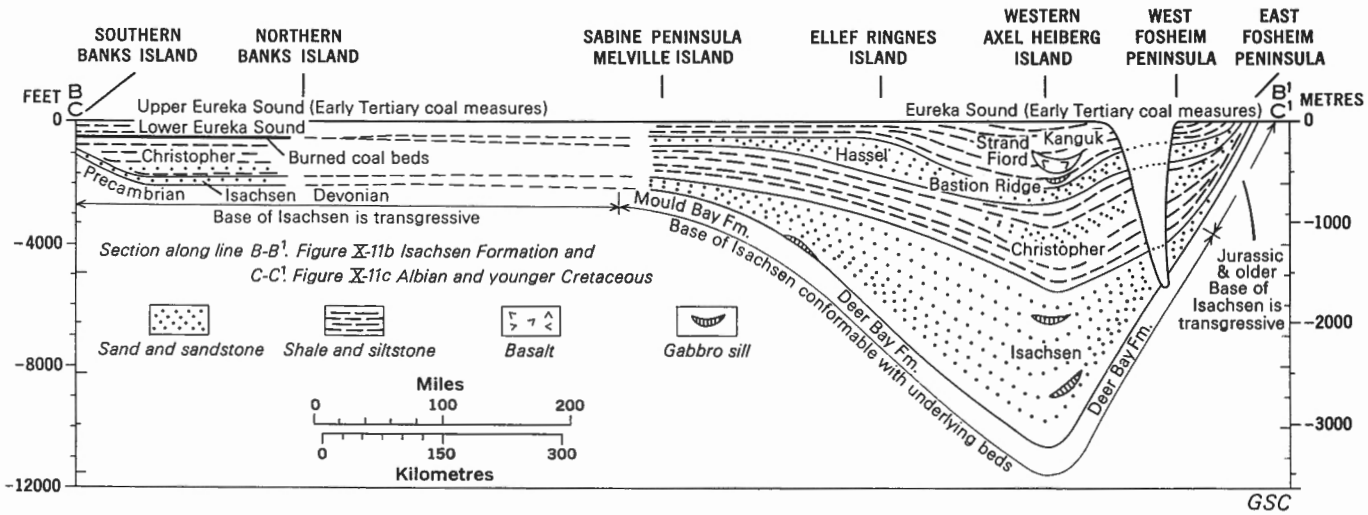

FIGURE X-13. Restored sections, Cretaceous rocks of Arctic Archipelago, along lines B–B′ and C–C′ of Figures X-11b, c (by E. T. Tozer).

mere and Axel Heiberg Islands a prominent member of greenish grey sandstone occurs near the middle. In northern Axel Heiberg Island basalt flows are intercalated. The lower part of the Christopher is commonly unfossiliferous and of uncertain age. *Tropaeum* of Aptian age has been tentatively identified from Mackenzie King Island. Higher beds above the sandstone member of Axel Heiberg and Ellesmere Islands contain the *Cleoniceras* aff. *subbaylei, Beudanticeras affine,* and *Gastroplites* zones of Albian age.

Throughout Sverdrup Basin, on Ellesmere, Axel Heiberg, Ellef Ringnes, and Melville Islands, the Christopher shales are followed by the Hassel Formation, which consists mainly of medium- to fine-grained quartzose sandstone and siltstone. Macrofossils are rare; the marine bivalve *Panopea* has been recorded from Axel Heiberg, *Spirophyton* and worm burrows are also known there. Microplankton tentatively dated as Cenomanian have been described from Graham and Ellef Ringnes Islands. These organisms evidently indicate a marine environment of deposition but much of the Hassel sandstone on Ellesmere Island is highly carbonaceous and has a non-marine appearance. The Hassel Formation probably formed on a delta covered at intervals by marine waters.

On Banks Island, in the western part of the Arctic Lowlands, the Christopher shales are overlain by 25 feet of coal-bearing beds, commonly burned brick-red, followed by about 400 feet of light grey shale and fine-grained sandstone. The latter contains marine microplankton, that are probably Upper Cretaceous and may be equivalent to the Hassel.

Upper Cretaceous

The Kanguk Formation consists of dark grey shale and siltstone with subordinate sandstone and some local thin bentonitic and tuffaceous beds. The Kanguk has been identified only in Sverdrup Basin, although equivalents may occur on Eglinton Island. Normally the Kanguk conformably overlies the Hassel Formation, but on western Axel Heiberg Island the Strand Fiord basalts intervene and a thin shale, the Bastion Ridge Formation, rests on Hassel sandstone. This shale is probably equivalent to basal Kanguk beds of other sections. *Inoceramus* cf. *pictus* of Upper Cenomanian age occurs on Graham Island in beds that probably represent the lower part. *Inocerami* of the *lobatus* species group, dated as Santonian or early Lower Campanian, are common in the upper beds. *Scaphites* species of two zones occur, one of Lower Santonian age and the other of late Lower or early Upper Santonian age. The Kanguk Formation is the youngest marine Mesozoic formation in the Arctic Archipelago.

Cretaceous Volcanic Rocks

Basalt is intercalated with the Cretaceous sediments of western Axel Heiberg Island and northern Ellesmere Island. The oldest are on the west side of Axel Heiberg Island where as many as six flows with columnar jointing may occur within the Isachsen Formation. Their aggregate thickness is not known, but may amount to several hundred feet. A flow about 50 feet thick occurs between the Isachsen and Christopher Formations north of Blaa Mountain, Ellesmere Island. Others intercalated in the Christopher Formation of northwest Axel Heiberg Island are probably more than 2,000 feet thick. The youngest basalts, known as the Strand Fiord Formation, are widely distributed in the vicinity of Strand and Expedition Fiords. They lie between the Bastion Ridge and Kanguk shales and are probably Upper Cretaceous. In the Strand Fiord area the thickness increases from east to west, from 280 feet at the head of Strand Fiord to 600 feet at the tip of Kanguk Peninsula. North of Strand Fiord it is probably greater.

The main area for Cretaceous volcanic activity lay near northwest Axel Heiberg Island. This Cretaceous volcanic activity affecting Sverdrup Basin was evidently an earlier event than the Tertiary, Thulean, volcanicity that affected the North Atlantic area.

Cretaceous (?) Gabbro Sills and Dykes

In the western part of Sverdrup Basin gabbro sills and dykes intrude Carboniferous and Permian rocks and all Mesozoic formations that underlie, or pre-date, the basalts of the Cretaceous Strand Fiord Formation. Within the basin no igneous intrusions are known to cut the Upper Cretaceous Kanguk shales or rocks of Cenozoic age.

The gabbro occurs as individual sills up to 300 feet thick, but most are thinner. Numerous thick sills intrude the Permian rocks of northwest Axel Heiberg Island, Triassic shales, and the lower Heiberg Formation of Eureka Sound area; the Savik and Deer Bay shales of western Axel Heiberg Island; and the Deer Bay of Ellef Ringnes Island. Sills also intrude the Bjorne, Blind Fiord, Awingak, Isachsen, and Bastion Ridge Formations.

Dykes are widespread, occurring as swarms on Cornwall Island and in many parts of Axel Heiberg Island. Near Flat Sound, a dyke intrudes the Blaa Mountain Formation and is bevelled beneath the overlying Cenozoic Eureka Sound beds. They are common on western Ellesmere Island but are scarce farther east. A few occur on the Ringnes Islands, and near Tingmisut Lake of Melville Island, but none is exposed on Prince Patrick, Brock, Borden, and Mackenzie King Islands.

Masses of gabbro are common in some of the anhydrite diapirs that intrude the Sverdrup Basin sequence. Some may be upthrust sills but others may be true ring dykes that have selected the circular anhydrite diapiric intrusion as a conduit.

The stratigraphic evidence suggests that all the gabbro intrusions are not younger than Cretaceous, and it appears probable that they may have been emplaced at about the same time as the extrusion of the Cretaceous basalts of Axel Heiberg Island. Four samples from Ellef Ringnes have provided whole rock K–Ar age determinations ranging from 102 to 110 m.y.

The Cenozoic in the Arctic Archipelago was a time of orogenic activity, non-marine sedimentation and, in limited areas, of volcanic activity. The principal sediments are the early Cenozoic Eureka Sound Formation and the late Cenozoic Beaufort Formation. Eureka Sound beds are found in Sverdrup Basin as outliers upon the folded rocks of the Franklinian Geosyncline and they underlie the western part of the Arctic Lowlands on Banks Island. The principal period of tectonic activity, herein termed the Eurekan Orogeny, involved folding, thrust faulting, and diapiric intrusion of Upper Carboniferous evaporites. The deformation was most intense on Ellesmere and Axel Heiberg Islands. In western Sverdrup Basin the effects of the orogeny were less intense and the Arctic Lowlands were unaffected. An early phase of orogenic activity is recognized locally as occurring before and during the deposition of the Eureka Sound Formation. Most orogenic activity, however, post-dates these beds and precedes the deposition of the Beaufort Formation. Volcanic rocks occur in the Eureka Sound Formation of Bathurst Island and on the east coast of Baffin Island.

The mid-Cenozoic was mainly a period of uplift, followed by peneplanation of the greater part of the archipelago. Parts of the Beaufort Formation on Arctic Coastal Plain probably represent alluvium produced during this interval of denudation. Cenozoic normal faulting occurred in many parts of the archipelago, notably on the Prince Patrick and Boothia Uplifts, and the northwest-trending faults that dislocate the Paleozoic rocks of Baffin Island are probably of this age. Many faults appear to have had a long history of Cenozoic movement. On the Prince Patrick Uplift most of the movement took place between the deposition of the Lower Cretaceous Christopher Formation and the late Cenozoic Beaufort Formation but some movement was later. The very form of the archipelago, with numerous stretches of linear coast, and elevated islands adjacent to the straits and channels, is probably due to late Cenozoic normal faulting, much of which probably took place after the deposition of the Beaufort beds.

Early Cenozoic Non-marine Rocks

The early Cenozoic beds, the Eureka Sound Formation, appear to be entirely non-marine, and locally include considerable conglomerate; they are in part synorogenic. The age and correlation of most of the formation are not precisely known. Plant microfossils have been obtained from many localities; most have been dated as Paleocene or Eocene.

In much of Sverdrup Basin, on Sabine Peninsula, Ellef Ringnes Island, western and southern Axel Heiberg Island, and western Fosheim Peninsula, the Eureka Sound rests with structural conformity upon the Upper Cretaceous Kanguk shales. These rocks consist mainly of sandstone and shale, with numerous coal seams up to 30 feet thick. Conglomerate is rare. About 10,000 feet of beds is present on Fosheim Peninsula and at Strand Fiord.

In northeastern Sverdrup Basin, on eastern Axel Heiberg Island, and in northern Ellesmere Island, the Eureka Sound beds are about 3,000 feet thick and rest unconformably upon rocks ranging in age from early Paleozoic to Lower Cretaceous. In addition to the typical sandstone, shale, and coal beds, these beds include much conglomerate of local derivation. At the head of Mokka Fiord, conglomerate occurs high in the section, above a sequence of coal and sandstone; thirty miles northwest it occurs at the base of the section. In these areas the Eureka Sound outcrops are commonly bounded by faults. The conglomerates occur adjacent to the faults and evidently represent syntectonic fanglomerates as the cobbles are formed from rocks of the upthrown side.

In central Ellesmere Island, the Eureka Sound beds rest disconformably on various Paleozoic formations. The formation is mainly sandstone, shale, and coal, several thousand feet thick. In this area considerable uplift and erosion but no appreciable folding took place during the interval between deposition of the Silurian and Cenozoic strata.

In the Boothia Uplift the Eureka Sound Formation on Bathurst, Cornwallis, western Devon, and Somerset Islands occurs in grabens, unconformably overlying Paleozoic and Precambrian rocks. On Cornwallis Island about 2,500 feet of sand, sandstone, and clay, with coal seams, is present, and on Bathurst Island, unlike the other areas, basalt is interbedded (Kerr and Temple, 1965).

Termination of Uniform Subsidence in Sverdrup Basin

The character of the Eureka Sound Formation shows that the advent of Cenozoic time marked a fundamental change in the pattern of sedimentation and subsidence in Sverdrup Basin. From the Upper Carboniferous to the Upper Cretaceous the basin was more or less continuously an area of subsidence and deposition, filled with sediments derived from lands beyond the basin and not from uplifts within the basin itself. With the deposition of the Eureka Sound Formation the basin ceased to act as a simple downwarp, receiving a concordant sequence of sediments. As already mentioned there are places where the Eureka Sound beds concordantly follow the Upper Cretaceous. From this evidence alone it might be supposed that the Eureka Sound beds simply represent one more deltaic formation, analogous to the Isachsen and Heiberg Formations. However the relations on eastern Axel Heiberg and northern Ellesmere Islands, where the Eureka Sound beds are definitely synorogenic, show that uplifted ridges, being subjected to denudation, were exposed on the site of Sverdrup Basin in early Cenozoic time. The tectonic disturbance indicated by these synorogenic beds marks the end of uniform subsidence within the Sverdrup Basin and the beginning of Eurekan Orogeny.



PLATE X-7

Eurekan folds with diapiric anticlines, western Axel Heiberg Island, south of Strand Fiord. Formations are: A—Carboniferous evaporites, C—Christopher, H—Hassel, B—Bastion Ridge, S—Strand Fiord, K—Kanguk, E—Eureka Sound, diapiric intrusion (X). Ice covered mountains are formed of anticlinally folded Triassic Heiberg Formation and constitute the drainage divide of the island.

Eurekan Orogeny

In mid-Cenozoic time folding and thrust faulting took place throughout much of Sverdrup Basin and the adjacent part of Ellesmere Island. In the axial part of the basin Upper Carboniferous evaporites were intruded as piercement domes and diapirs. The principal Eurekan fold axes, faults, and piercement structures are shown on the tectonic map (Map 1251A). The more intense movements affected much of Ellesmere Island and nearly all of Axel Heiberg Island (Pl. X-7). On Cornwall and the Ringnes Islands there was some Eurekan folding but farther west, in the western Queen Elizabeth Islands, the folds are of low amplitude. Although Eurekan structures are superimposed with concordant strike on older Ellesmerian structures the Eurekan movements were not confined to the area deformed in Paleozoic time, as on much of central Ellesmere Island between Baumann and Canyon Fiords. Eurekan structures on western Ellesmere Island commonly strike northeasterly, parallel to the Ellesmerian structures. Those of Axel Heiberg Island also seem to reflect the strike of older folds, for the northwesterly strike of the major anticlinorium extending from Rens Fiord to the head of Skaare Fiord conforms with the Paleozoic strike imposed on the eugeosynclinal rocks of the Rens Fiord area.

The United States Range and Victoria and Albert Mountains were deformed during the Eurekan Orogeny. Hazen Plateau which lies between these ranges was barely disturbed and apparently formed a rigid block, folded during the Ellesmerian Orogeny but unaffected by the Eurekan deformation.

The Eurekan structures include a variety of folds, thrust faults, and normal faults. In the axial part of the Sverdrup Basin piercement structures, involving the intrusion of Carboniferous evaporitic rocks, are intimately related to the Eurekan structures. The anticlines of southern Axel Heiberg Island south of the belt of piercement structures are mainly box folds, with long, straight axes, flat crests, and steeply inclined limbs. To the north, in west-central Axel Heiberg Island, the folds have short,

arcuate, steeply plunging axes. The cores of the anticlines are commonly pierced by Upper Carboniferous evaporites. In this area the mobile layer of evaporites has clearly affected the folding. A belt of relatively intense folding on both sides of Eureka Sound is bordered to the west by the piercement belt. Probably these folds indicate a *décollement* at the Upper Carboniferous evaporites. There are no piercement structures in this belt, probably because competent Upper Carboniferous and Permian limestones immediately overlie the evaporites.

Numerous Eurekan thrust faults, with hanging-walls of Cambrian to Devonian strata, displaced eastwards, occur between Baumann Fiord and Copes Bay, on Ellesmere Island. Many are inclined at relatively low angles at the surface, the fault plane following evaporites in the lower part of the Ordovician Cornwallis Group. Between Bay and Tanquary Fiords many faults affect the upper Paleozoic, Mesozoic, and Cenozoic rocks of the Sverdrup Basin sequence. Most displacements are towards the east, but some, as north of Blaa Mountain, are towards the northwest. On some faults the Upper Carboniferous evaporites (Otto Fiord Formation) form the hanging-wall. The large Whitsunday Bay fault of eastern Axel Heiberg Island, placing Carboniferous rocks adjacent to the Eureka Sound Formation, is also locally soled by these evaporites. This fault is bordered by Eureka Sound fanglomerates and has experienced the early Eurekan movements. Thrust faults, directed to the east, northeast, and west dislocate the upper Paleozoic and Mesozoic rocks between Strand and Bunde Fiords of western Axel Heiberg Island. The Upper Carboniferous evaporitic rocks commonly overlie the thrust plane.

Transcurrent and normal faults occur in the area affected by the Eurekan Orogeny, but their displacements are small.

Piercement bodies of Upper Carboniferous evaporites are characteristic of the axial part of Sverdrup Basin, extending for 400 miles in a northeasterly trending belt from Sabine Peninsula on Melville Island to Otto Fiord, Ellesmere Island. The material forming the piercement structures is part of Otto Fiord Formation, which lies near the base of the Carboniferous section in the axial part of the basin. The piercement bodies seem to be confined to areas where incompetent shales and siltstones of Hare Fiord Formation overlie the evaporites, as described in the part of this chapter dealing with the Carboniferous rocks. Some of the bodies are domes produced by geostatic loading rather than the stresses of the Eurekan Orogeny, but the domes are so intimately related to evaporite-cored diapirs produced during the orogeny that it is appropriate to describe these features together.

The piercement bodies occur in three main structural situations:

1. *In synclinal areas.* These are more or less circular domes, 6 to 7 miles in diameter, confined to the western part of Sverdrup Basin and typified by the domes of Sabine Peninsula, Melville Island, the Isachsen dome of Ellef Ringnes Island (Pl. X-8), and the South Fiord dome of Axel Heiberg Island. The outcropping rock is mainly gypsum and anhydrite with inclusions of Upper Carboniferous limestone and gabbro. Some of the gabbro may represent uplifted sills, but on Sabine Peninsula the igneous rock is in the form of a ring dyke possibly indicating that the dome was in existence, although not necessarily exposed, when the gabbro was intruded, probably in the Cretaceous. Negative gravity anomalies of 10 and 40 milligals above the domes of Ellef Ringnes Island suggest that these bodies are underlain by halite and have risen as a result of geostatic loading of evaporites of low density. The strata adjacent to the domes on Ellef Ringnes and Axel Heiberg Islands are steeply inclined. On Sabine Peninsula the surrounding beds appear to be nearly horizontal. Cretaceous beds become thinner when traced towards the Isachsen dome suggesting that it had formed, and started movement, prior to the Eurekan folding.

2. *In the cores of anticlines.* These bodies are usually narrowly ovoid, pear-shaped or linear, but locally, as on the north side of Mokka Fiord, parts assume circular form. These structures probably owe their origin to tectonic squeezing with subordinate elevation due to density differences. Sedimentary sections on the flanks of these diapiric folds are commonly thick suggesting that the diapirs, unlike the domes, were tectonically inactive during Mesozoic time. Near the diapir east of Cape Levvel small gypsum dykes and sills intrude the Jurassic and Cretaceous shales at a distance up to 3 miles from the main intrusion.

3. *Adjacent to thrust faults.* Numerous piercement bodies occur at or near the surface trace of the Swiss Range and Whitsunday Bay thrust faults of Axel Heiberg Island, and a large diapiric intrusion occurs near the south limit of the thrust fault north of Blaa Mountain, Ellesmere Island. These bodies, like those on the axes of anticlines, probably resulted partly from tectonic squeezing and partly through elevation due to density difference. Immature topography characterizes all the piercement structures. They project above the surrounding landscape suggesting recent uplift or that they may be undergoing uplift. There are, however, some interesting exceptions, notably that on eastern Axel Heiberg Island adjacent to the Whitsunday Bay thrust. For about 15 miles, there is an almost continuous belt of evaporitic rocks which includes three piercement bodies characterized by immature topography, and also an area where the evaporites are bedded and essentially *in situ*. In addition, however, there are evaporites which are apparently intrusive, but bevelled to a level concordant with that of the surrounding country. In this belt, by physiographic criteria, it would appear to be possible to distinguish between active diapiric bodies, characterized by immature topography, and inactive ones that have moved in the past but have not been rejuvenated.

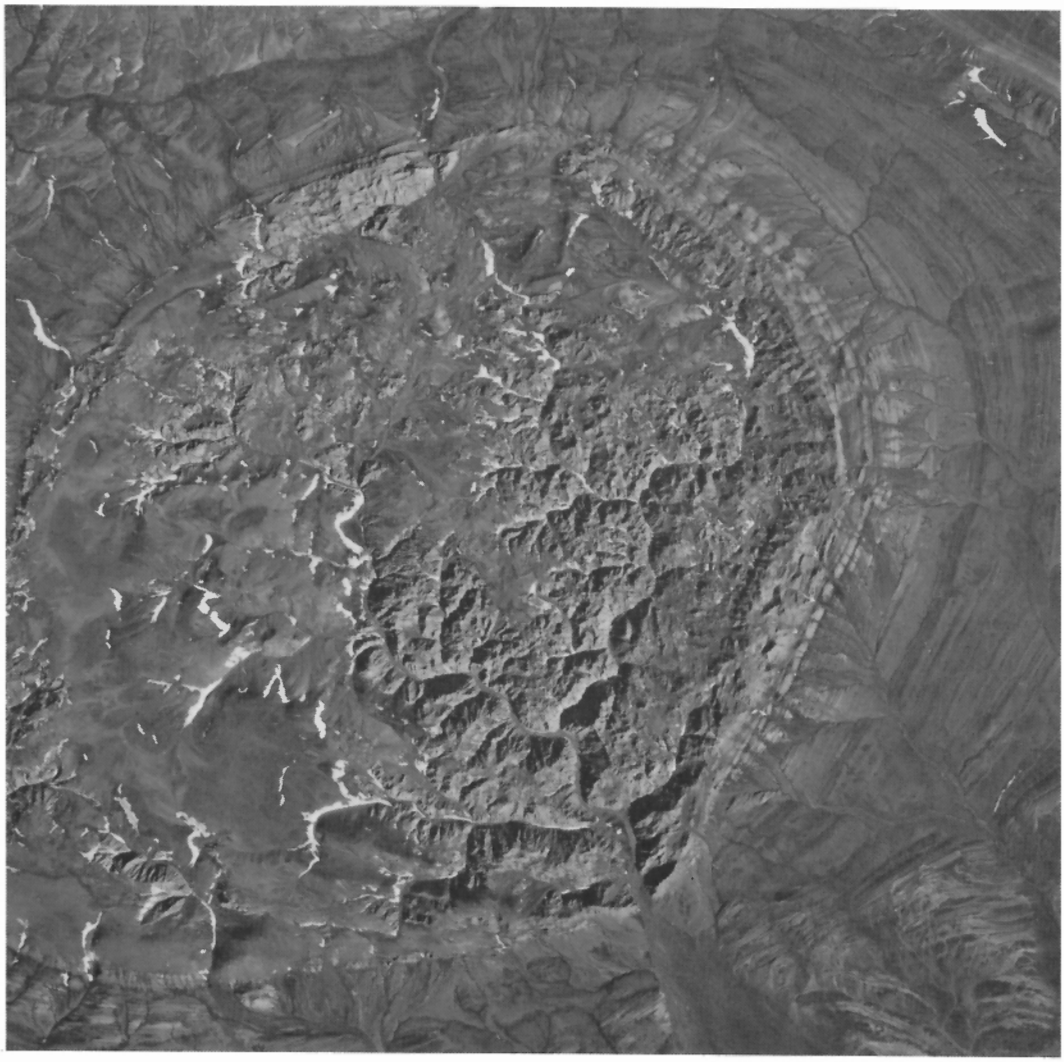

PLATE X-8
Isachsen dome, Ellef Ringnes Island. Core 4 miles in diameter of Upper Carboniferous evaporites, flanked by sharply inclined Cretaceous, Isachsen, and Christopher Formations.

Mid-Cenozoic Normal Faulting

North-trending normal faults characterize Prince Patrick Uplift on Banks and southern Prince Patrick Islands; and Boothia Uplift, on Bathurst, Cornwallis, Devon, and Somerset Islands. Northwesterly trending faults are present throughout much of Baffin Island. The time of fault movement cannot be precisely determined but is probably mid-Cenozoic. Some faults also show evidence for late Cenozoic movement, as noted below.

Late Cenozoic

Wood-bearing alluvial deposits, known as the Beaufort Formation, underlie the Arctic Coastal Plain from Meighen Island to the southwestern tip of Banks Island. In addition, isolated patches of similar deposits occur on hilltops to the east on Prince Patrick, Ellef Ringnes, Banks, and Melville Islands and even on the west side of Ellesmere Island. The Beaufort beds generally rest upon a surface of low relief and overlie unconformably rocks ranging in age from Precambrian to early Cenozoic. On Meighen Island, silts with poorly preserved bivalve shells, probably of marine genera, underlie the Beaufort beds. This occurrence suggests the possibility of other marine deposits beneath the Beaufort on the Arctic Coastal Plain and the Continental Shelf. Towards the east, where the basal beds are visible, the Beaufort is a non-marine, alluvial deposit. On Banks Island it attains a thickness of 400 feet; the total thickness to the north is not known, but exceeds 250 feet. The Beaufort Formation consists of unconsolidated crossbedded sand, gravel, and silt, with woody detritus, wood, and rare peat. The sands are well sorted and quartzose and the gravels contain an abundance of well-rounded resistant rocks derived from distant sources to the east and southeast. Logs of wood more than 10 feet long and up to a foot in diameter are common. The wood is normally unaltered spruce and pine and unlike the carbonized and mineralized wood from the older Cenozoic and Mesozoic formations. As the wood is driftwood, there is no evidence to indicate that it grew where found. Rock samples characteristically

contain pollen of spruce, pine, birch, alder, various herbaceous plants, and spores of ferns and mosses. Many samples also contain small amounts of pollen from hemlock and temperate climate hardwoods such as hazel, beech, elm, hornbeam, and oak. Some hardwood pollen grains may be secondary and derived from older Cenozoic strata.

The age of the Beaufort Formation is uncertain. It is probably Pliocene or early Pleistocene but preglacial. The continuity of the formation between islands now separated by straits and channels indicates that the beds were laid down before the archipelago assumed its present form. It seems to represent the thin inland part of a body of sediment (derived from sources to the southeast and east) that extends seaward beyond the present coast. Parts of the formation probably represent detritus produced during denudation of the areas affected by the Eurekan Orogeny.

Structure of the Arctic Coastal Plain and Continental Shelf

The youngest clearly defined structural features in the archipelago are those of the Arctic Coastal Plain. On outcrop the Beaufort beds appear to be horizontal but air photographs reveal that they are inclined (J. G. Fyles). On Prince Patrick Island, east of Discovery Point, the interfluves show chevron patterns that indicate a northwesterly dip at an angle greater than the gentle seaward slope of the plain. On Isachsen Peninsula, Ellef Ringnes Island, a series of low cuestas reveals a northeast dip, away from the Arctic Ocean.

On Prince Patrick Island the plain is traversed by several spectacular lineaments, up to 40 miles long, that represent the surface traces of faults. Northwest of Mould Bay most trend north and are aligned with pre-Beaufort faults. Northwest of Intrepid Inlet they trend northeast. In places these lineaments have virtually no topographic expression, but elsewhere they represent scarps with a relief of 20 to 30 feet. Northwest of Intrepid Inlet the scarps face the Arctic Ocean, suggesting that the northwest sides are downthrown. Although most movement on faults of Prince Patrick Uplift predates deposition of the Beaufort beds, these lineaments indicate later minor movements. Prince Patrick Island currently experiences more seismic activity than neighbouring regions, possibly indicating movement on these faults.

Formation of the Archipelago

Numerous straits and channels indicate a drowned topography, but the physiography of most individual islands shows that they have been uplifted relatively recently. In the areas affected by Eurekan Orogeny the uplift is clearly later than the peneplanation which followed the orogeny, probably the result of late Cenozoic movement on gravity faults.

As previously described, Cenozoic normal faults are widespread in the archipelago. In addition to those recognized on land there are many linear and arcuate stretches of coast, often with seacliffs up to 3,000 feet high, that probably indicate the presence of faults. Many islands seem to have been uplifted, to different degrees, by movement on the faults that border the coasts. Uplift along such faults would account for the different physiographic characters of adjacent islands that are geologically similar, for example, the Ringnes Islands compared with Axel Heiberg Island. These islands are formed of Mesozoic and early Cenozoic rocks that were folded during the Eurekan Orogeny. After the orogeny they experienced planation. The peneplane on Axel Heiberg Island is more elevated than that on the Ringnes Islands. The linear connection between the headlands on the west coast of Axel Heiberg Island indicates the probable position of the faults along which Axel Heiberg Island has been elevated compared with the Ringnes Islands. Some of the uplift on such faults probably took place after deposition of the Beaufort Formation and although a relatively recent event it should not be confused with the later widespread post-Pleistocene uplift. Areas that experienced substantial post-Pleistocene uplift commonly were not appreciably elevated during the earlier period and *vice versa*. The west sides of Axel Heiberg and Melville Islands, for example, show elevated peneplanes but no strandlines at high elevations. Southwest Cornwallis Island, in contrast, shows elevated strandlines but little evidence for the earlier period of uplift. These examples leave little doubt that two periods of uplift, one before the Pleistocene, the other after, have affected the archipelago. The earlier period was probably related to faulting and essentially responsible for the present form of the archipelago; the later one was simply due to isostatic rebound following deglaciation.

ECONOMIC GEOLOGY

With the exception of small amounts of coal for local use the Phanerozoic rocks of the Arctic Archipelago have not, as yet, yielded minerals of economic value. There has been some activity in searching for metallic minerals on the islands, but mostly within the Precambrian terrains dealt with in Chapter IV. From an economic standpoint the principal interest of the arctic islands lies in the petroleum possibilities.

Gypsum and anhydrite. Deposits of gypsum and anhydrite have been found in Cambrian, Ordovician, Silurian, Upper Carboniferous, and Permian formations. The most notable

are in the Ordovician and the Upper Carboniferous. Also noteworthy are the evaporites in the Hadrynian, Minto Inlet, Wynniatt, and Kilian Formations of the Shaler Group of Victoria Island.

Iron. Massive magnetite occurs in a gypsum diapir, northwest of the head of Strand Fiord, Axel Heiberg Island.

Sulphur. Native sulphur occurs in association with calcite in a diapir on western Axel Heiberg Island.

Lead and zinc. Lead and zinc sulphides have been found in brecciated dolomite of the Ordovician Cornwallis Formation on Little Cornwallis Island. A small amount of galena has been observed in the Blue Fiord limestone east of Weatherall Bay, Melville Island.

Copper. A few small flecks of native copper have been observed in the volcanic rocks of the Natkusiak Formation, of Hadrynian age, on Victoria Island, and reports from Eskimos suggest that there may be larger occurrences in this formation but this has not been confirmed.

Coal. Coal seams occur in formations of Devonian, Lower Carboniferous, Triassic, Jurassic, Lower Cretaceous, and Cenozoic age. The thickest, up to 30 feet, are in the Cenozoic Eureka Sound Formation. Seams up to 5 feet thick have been reported from the Lower Cretaceous Isachsen Formation. The seams in the older formations are thin, generally less than a few inches thick. A slacking, non-coking sub-bituminous coal, probably of Lower Cretaceous age, has been mined near the settlement of Pond Inlet on northern Baffin Island. Slightly more than 100 tons has been produced annually, mainly for use at the settlement.

Oil and gas. The great volume of Phanerozoic sediment and the presence of numerous structures in the arctic islands have attracted the attention of geologists seeking petroleum. Since 1959 geological parties from several oil companies have visited the islands and in 1961-62 the first exploration well was drilled at Winter Harbour on Melville Island. Since then wells have been drilled on Bathurst and Cornwallis Islands. No commercial production has been found, although a showing of gas was obtained from the Devonian rocks encountered in the Winter Harbour well. Bitumen deposits have been discovered in the Lower Triassic Bjorne Formation of Melville Island (Trettin and Hills, 1966). They are now uneconomic but are a direct indication of petroleum in Sverdrup Basin. Several potential petroleum provinces have been outlined (Douglas, *et al.*, 1963). Porous carbonate rocks, constituting potential reservoirs, occur at many levels in the Paleozoic, notably in the Ordovician, Silurian, Devonian, and Upper Carboniferous. Potential sandstone reservoirs occur in the Devonian, Upper Carboniferous, and Permian and throughout the Mesozoic succession. Marine shales, which may be regarded as potential source rocks, occur at all these levels. Anticlinal structures produced during the Ellesmerian Orogeny are a notable feature of the Parry Islands and Ellesmere Fold Belts. Structures within the northern part of the Boothia Uplift may also provide situations for the accumulation of hydrocarbons. Within the Sverdrup Basin the Eurekan folds and associated piercement domes, composed of upthrust Upper Carboniferous evaporitic rocks, provide possible structural traps for petroleum. Abrupt facies changes occur at many levels in the section, both in Franklinian Miogeosyncline and Sverdrup Basin and may have produced stratigraphic traps.

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XI. Biochronology: Standard of Phanerozoic Time

Introduction 592

Cambrian Faunas 593

Ordovician and Silurian Faunas 601

Devonian Faunas 614

Carboniferous and Permian Faunas 623

Marine Triassic Faunas 633

Marine Jurassic Faunas 641

Cretaceous Macrofaunas 649

Paleobotany 663

Marine Pleistocene Faunas 671



INTRODUCTION
by D. J. McLaren

Phanerozoic time is represented by rocks that contain abundant fossilized remains of the hard parts of Metazoan animals. It includes all geological periods from Cambrian to Recent and directly succeeds Proterozoic time during which organisms possessing hard parts either did not exist or were very rare. Proterozoic fossils include plant micro-organisms, stromatolitic algae, casts, trails, and imprints, but true body fossils are unknown and appear suddenly in Cambrian rocks.

Biochronology depends on the law of superposition and the continuous evolution of life. It affords a unique method of correlating widely separated events on the earth's crust and of erecting an empirical but remarkably accurate relative time-scale. For example, correlation between continents of events during the Cretaceous Period is effected by means of a faunal zonation based largely on ammonites and tested repeatedly in many parts of the world. Each zone has an average duration of some 1.6 to 1.8 million years, using an estimated duration for the period based on radioactive age determinations. During the Ordovician and Silurian, worldwide graptolite zones had a duration of about 2.5 million years using similar estimates. Within a continent or major region biostratigraphic zonation allows a more accurate temporal subdivision, and zones representing one million years or less are common. An interesting feature of the method is that accuracy does not decrease with time but is roughly constant, depending rather on the group of organisms chosen for such zonation. Thus the percentage error, in terms of absolute age, theoretically decreases with increasing age back to the Cambrian. Biostratigraphic zones based on stromatolites so far proposed for the Proterozoic are broad, each zone corresponding to several hundred million years. Such zonal schemes have not yet been tested in Canada.

All fossils, to a greater or lesser extent, are dependent on facies, and all fossils, similarly, are biochronologically meaningful. Thus trilobites are not found in freshwater sediments, and oysters are unknown in the Paleozoic. Fossils useful biochronologically are those groups in which evolution proceeded rapidly and produced recognizable morphological changes to the skeleton or shell and which were also widespread and found in a variety of rock types. Planktonic or nektonic organisms, such as graptolites, ammonites, or certain Foraminifera, frequently are the most useful, but in some facies such organisms may be rare. Benthonic fossils, such as brachiopods, bryozoans, or pelecypods, may be equally valuable provided facies restrictions are recognized. In many areas, particularly in dealing with Paleozoic rocks, it may be possible to erect a local or regional zonation that is more refined than the standard planktonic zonation used for intercontinental correlation. For example, in western Canada eleven trilobite zones are recognized in Lower Ordovician rocks, and eight brachiopod zones in the Upper Devonian. It appears increasingly likely that zonation by benthonic organisms may prove applicable to intercontinental correlation in many instances.

This chapter deals with the paleontological basis for correlation within the Phanerozoic rocks that are described elsewhere in this volume and which are named on the geotectonic correlation charts. The rock-units are arranged on these charts so that their relative position indicates geological age in terms of a standard succession. The standards shown are worldwide for the Mesozoic and Tertiary, and both North American and European for the Paleozoic. The paleontology of each period is discussed

XI

Biochronology: Standard of Phanerozoic Time

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Eurypterus lacustris Harlan—an arthropod from the Upper Silurian Bertie Formation of southern Ontario.

briefly and the guide fossils or faunal zonation are shown in the accompanying tables in relation to the standard sequence. Some of the more common or important fossils are figured. The discussion and charts are necessarily brief and highly selective, and do not give a true picture of the complexity and variety of past life. Nor is the treatment of each period necessarily uniform. The phyla and classes of animals and plants found most valuable in each period vary widely as, indeed, does the lithology of the rocks in which they are contained. This, in itself, is a reflection of the complexity of the biochronologic method and its necessary adaptation to local and regional requirements.

The geological time-scale is an empirical artefact that has grown haphazardly, depending on observations spread over nearly 200 years in areas determined by entirely extraneous factors that were largely social and economic. Real stratigraphic boundaries in regions where the hierarchy of time-rock terms was invented have to be searched for elsewhere in sequences of rocks which may represent a stratigraphic continuum or which may be entirely absent. Such correlations are made by fossils. It is not surprising that the correlation of boundaries is not always easy and, strictly in a physical sense, such boundaries may not represent real or easily established events. These comments apply in force to this country in which, over vast areas, rigorous stratigraphic and biochronologic observations date back hardly more than 20 years. It is a remarkable vindication of the methods of biochronology that such accurate time-scales may be erected for most periods, and it is these time-scales that are summarized and discussed here.

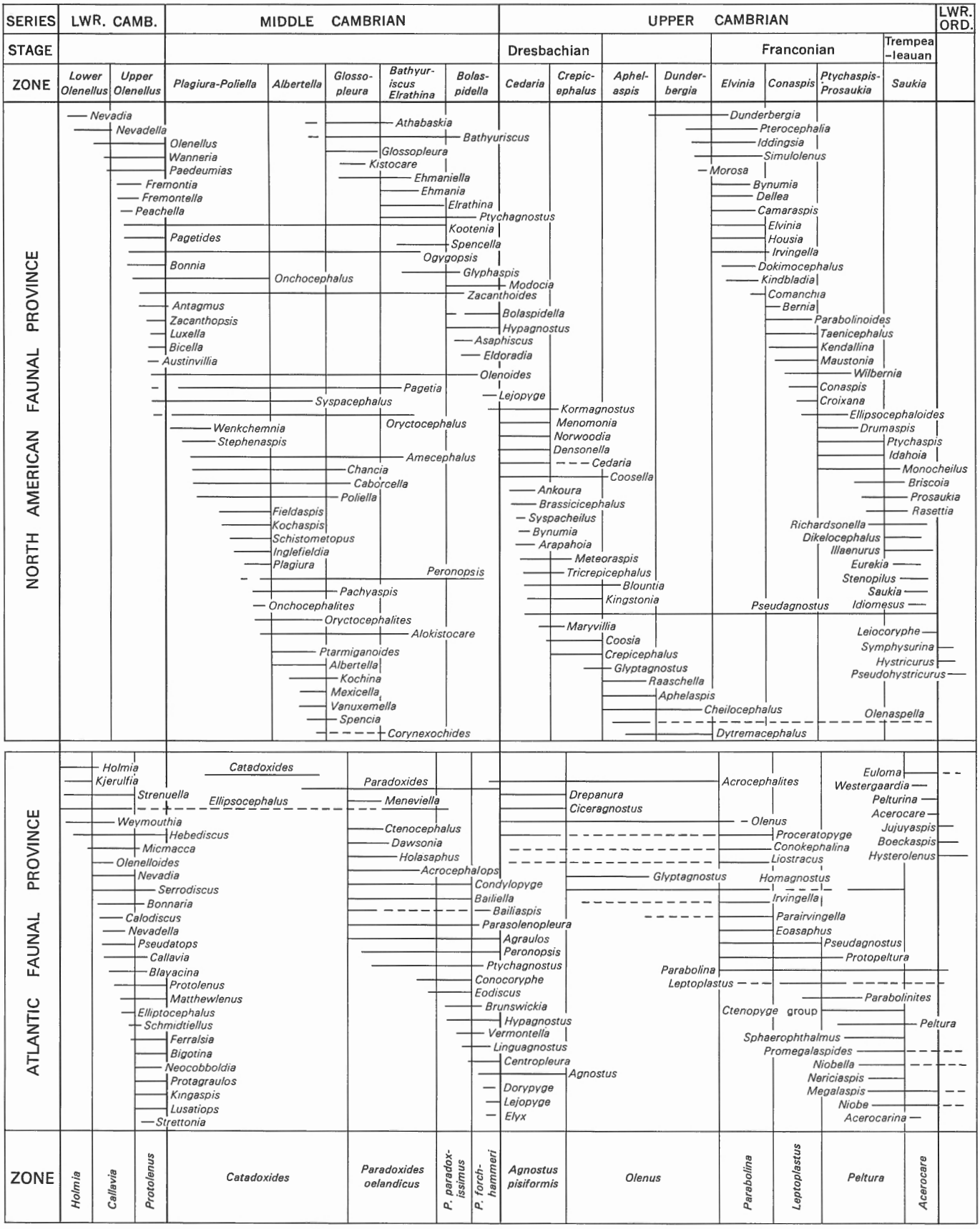
CAMBRIAN FAUNAS

by W. H. Fritz

The Cambrian is the oldest geological system containing numerous and diverse megafossils. Strata immediately predating the Cambrian appear to be unfossiliferous only because Precambrian organisms did not have the hard parts necessary for their preservation. Rarely, however, soft-bodied animals such as sea pens and jellyfish did make impressions as in the latest Precambrian strata of Australia (Glaessner, 1962), and so left a brief record of surprisingly advanced forms. Later, in Middle Cambrian time, soft-bodied worms, algae, sponges, holothuroids, jellyfish, and arthropods underwent unusual preservation in what is known as the Burgess Shale of British Columbia (Walcott, 1910–12). Fossils from both localities provide but a glimpse of a large segment of early life that otherwise perished unrecorded.

Although there is still some uncertainty as to the rate and order in which organisms first developed hard parts, there is no doubt that these parts provided an evolutionary advantage. Proof is the abundant skeletal material preserved in Cambrian to Recent strata. As might be expected, the number of fossils diminishes near the base of the Cambrian, and as yet there is no widely accepted agreement as to what faunal or stratigraphic horizon should represent the boundary between the Cambrian and the Precambrian. There has always been, however, a tendency to place the base of the Cambrian at or near

TABLE XI-1 | *Ranges of Cambrian trilobite genera in the North American and Atlantic Provinces*



GSC

the first occurrence of the remains of organisms containing hard parts. An important contribution towards resolving the boundary problem will come from additional study of the earliest skeletal-bearing animals such as the archaeocyathids (PL. I, FIG. 11), gastropods (PL. II, FIG. 14), hyolithids (PL. II, FIG. 2), sponges, and trilobites.

Of the various Cambrian fossils, trilobites are by far the most abundant and distinctive, and it is upon this group that the Cambrian biochronological framework is based. This framework for North America has been presented by Lochman-Balk and Wilson (1958), who show the life span or ranges of the trilobite genera and the manner in which their ranges are utilized to subdivide the Cambrian into zones. Each zone bears the name of one or more genera, but the zones are also based in part on additional genera that are considered to be of equal importance. Also, the name-bearing genus may have a range longer or shorter than the time-span of the zone it represents. A range chart generalized from the more extensive compilation by Lochman-Balk and Wilson is shown in Table XI-1.

Actually, not one but two charts are shown in Table XI-1 as the genera represented are from two distinct faunal provinces, the North American and the Atlantic Provinces. Within one province the relative range of various genera is well documented, and the provincial zonation has been used with reasonable success. It is difficult, however, to compare the age of genera or even zones across the provincial boundaries, as only a very small percentage of genera are common to both provinces.

Areally, the two faunal provinces found in North America are unequal. The smaller, the Atlantic Province, is confined to southeastern Newfoundland, eastern New Brunswick, Nova Scotia, and a small part of northeastern United States. Within it, trilobites bear a very close resemblance to those of the Atlantic Province in parts of Great Britain, Sweden, and various countries of Europe. Unique to the Atlantic Province are certain eyeless or nearly blind trilobites like *Bailiaspis* (PL. III, FIG. 16) and *Conocoryphe* (PL. III, FIG. 17) and gargantuan forms such as *Paradoxides* (PL. III, FIG. 18). Many trilobites from the Canadian part of the Atlantic Province have been described by Hutchinson (1962).

The North American Province includes all North America exclusive of the Atlantic Province. This is by far the more important of the two provinces in terms of areal extent and number of fossils collected in Canada, and therefore the paleontology of the North American Province is emphasized here.

The Cambrian fossils of the North American Province are discussed mainly with reference to the zones in which they occur. A broader discussion in terms of stages is not practical, as thus far only three Upper Cambrian stages have been erected. These three, the Dresbachian, Franconian, and Trempealeauan, have type areas in the upper Mississippi Valley where they were thought to represent all Upper Cambrian time. Modification of the three-stage concept is now needed, as a significant unconformity is known to be present at the Dresbachian–Franconian contact. Lochman-Balk and Wilson (1958) have suggested erecting a new stage elsewhere to represent strata and faunas missing in the Cambrian of the upper Mississippi Valley, but no action has yet been taken on their proposal.

Series names are rarely used in the Cambrian, but subdivisions of series magnitude, the Lower, Middle, and Upper Cambrian, have gained wide acceptance. These subdivisions in the North American Province approximately equal the Lower, Middle, and Upper Cambrian of the Atlantic Province. Thus it is the Lower, Middle, and Upper Cambrian and various Cambrian zones which are usually cited as a reference framework in discussing Cambrian fossils of the North American Province, and this procedure is followed here.

In the Lower Cambrian two zones are recognized, the lower and upper *Olenellus* Zones. Both zones characteristically contain olenellids which have large semicircular head shields and crescent-shaped eyes. In the lower *Olenellus* Zone, olenellids such as *Nevadella* (PL. I, FIG. 10) and *Nevadia* have an anteriorly tapered axial region on the head shield that does not reach the raised marginal border. In the upper *Olenellus* Zone, the axial region on olenellid head shields extends farther forward where it widens and has a subglobular shape (see *Olenellus,* PL. I, FIG. 1). Exceptions in the upper *Olenellus* Zone are a few species of *Olenellus* and most species of *Paedeumias* (PL. I, FIG. 2). Small ptychoparioid trilobites, such as *Onchocephalus* (PL. I, FIG. 4), are fairly common in the upper *Olenellus* Zone but are absent or rare in the lower *Olenellus* Zone. *Bonnia* (PL. I, FIGS. 8, 9) is common in the latest part of the upper *Olenellus* Zone.

In addition to trilobites, archaeocyathids (PL. I, FIG. 11) and *Salterella* (PL. I, FIG. 12) are common at some horizons in the Lower Cambrian. More prevalent are hyolithids (PL. II, FIG. 2) and inarticulate brachiopods (PL. II, FIG. 9), both of which range through the Cambrian and into younger systems. Articulate brachiopods (PL. I, FIG. 3) are rare in the Lower Cambrian and do not become common until mid-Upper Cambrian. Ostracods (PL. II, FIG. 13) and gastropods (PL. II, FIG. 14) are known from strata as old as Lower Cambrian, but they are rare, although gastropods increase in number by the late Upper Cambrian.

By Middle Cambrian time the olenellids were extinct and ptychoparioid trilobites with small tails were common. Trilobites belonging to the family Zacanthoididae, such as the genus *Albertella* (PL. II, FIG. 19), are prevalent in the early Middle Cambrian and decrease in numbers in younger strata. Genera that are widespread and prevalent, but restricted in time to the early Middle Cambrian, are *Plagiura* (PL. II, FIGS. 15, 16) in the late part of the *Plagiura–Poliella* Zone, and two genera, *Albertella* and *Glossopleura* (PL. II, FIGS. 4, 5) which are restricted to the zones bearing their names. *Bathyuriscus* (PL. II, FIG.

12) is an example of a genus that is most prevalent in, but overlaps, the zone bearing its name. Still longer ranging genera, such as *Ogygopsis* (PL. II, FIG. 8) and *Oryctocephalus* (PL. II, FIG. 3) are of special interest even though they have little value for dating purposes. These genera are abundant in rocks of a certain lithology, which suggests they may prove to be important paleoenvironmental indicators. Near the close of the Middle Cambrian, in the *Bolaspidella* Zone, ptychoparioid trilobites with convex, smooth central lobes on the head shields and small tails (see *Modocia,* PL. III, FIGS. 19, 20) increase in number. Trilobites belonging to the families Zacanthoididae and Bathyuriscidae are rare and become extinct within the *Bolaspidella* Zone. *Hypagnostus* (PL. III, FIGS. 21, 22) is locally abundant in this zone.

In the earliest Upper Cambrian, trilobites of the *Cedaria* (PL. III, FIG. 24) and *Crepicephalus* (PL. III, FIG. 23) Zones also have ptychoparioid-like heads, but the tails are much larger, being almost equal to the heads in size. These trilobites are accompanied by numerous genera of medium to small size that have low, smooth heads and tails. Also common in the *Cedaria* Zone is the agnostid *Kormagnostus* (PL. III, FIGS. 14, 15).

In the succeeding three Upper Cambrian zones, the *Aphelaspis, Dunderbergia,* and *Elvinia,* numerous genera appear that are characterized by rectangular lobes in the centre of the head shield. The truncated front part of this lobe differs from the rounded front part seen on trilobite head shields in the *Cedaria* and *Crepicephalus* Zones (compare *Aphelaspis,* PL. III, FIGS. 12, 13, and *Dytremacephalus,* PL. III, FIG. 11, with *Cedaria* and *Crepicephalus).* Other trilobites are quite different and distinctive, such as *Irvingella* (PL. III, FIG. 10) which occurs in the *Elvinia* Zone and is one of the best index fossils for worldwide Cambrian correlation.

Near the end of the Cambrian, in the *Conaspis, Ptychaspis–Prosaukia,* and early *Saukia* Zones, there is an increase in the number of large-sized trilobites, many of which have large, fan-shaped tails and rectangular lobes in the middle of the head shield (see *Briscoia,* PL. III, FIGS. 1, 2). In the latest Upper Cambrian, near the end of *Saukia* Zone, the number of large trilobites diminishes. *Eurekia* (PL. III, FIGS. 7, 8) is a distinctive genus in this zone. *Leiocoryphe* (PL. III, FIGS. 5, 6) is locally abundant and is perhaps more typical of the youngest Cambrian genera.

Trilobite genera on either side of the Cambrian–Ordovician boundary are small and diverse in outline, and thus an abrupt change is not apparent. Among the most common that become extinct in latest Cambrian time are *Eurekia, Idiomesus, Illaenurus,* and *Leiocoryphe.* Among the most common that originate in earliest Ordovician time are *Hystricurus* and *Symphysurina.* Slightly younger, larger, and thus more conspicuous Lower Ordovician genera are *Bellefontia, Kainella, Leiostegium,* and *Xenostegium.*

Perhaps the greatest change near the Cambrian–Ordovician boundary is the explosive increase in the number and type of graptolites and conodonts, which are rare and often overlooked in the Upper Cambrian.

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PLATE I. Lower and Middle Cambrian Fossils

Figure 1 *Olenellus* sp. Head x2, GSC No. 21348, upper *Olenellus* Zone, Peyto Formation, Chaba River, Alberta.

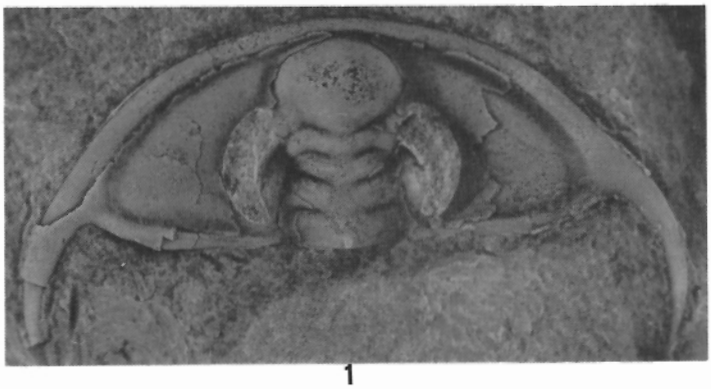
Figure 2 *Paedeumias* sp. Latex cast of head x2, GSC No. 21349, upper *Olenellus* Zone, Eager Formation, near St. Eugene Mission, B.C.

Figure 3 *Nisusia* sp. Latex cast of brachial and pedicle valves x2, GSC No. 21350, *Bathyuriscus-Elrathina* Zone, Burgess Shale, between Mount Field and Mount Wapta, B.C.

Figure 4 *Onchocephalus* sp. Part of head x5, GSC No. 21351, upper *Olenellus* Zone, unnamed formation near June Lake, District of Mackenzie.

Figure 5 *Fremontia?* sp. Latex cast of exoskeleton x2, GSC No. 21352, Lower Cambrian, Mahto Formation, Munn Peak, Alberta.

Figures 6, 7 *Wanneria* sp. Head x5 with part of area x10 to show reticulate surface pattern, GSC syntype No. 414b, Lower Cambrian, L'Anse au Loup, Labrador.



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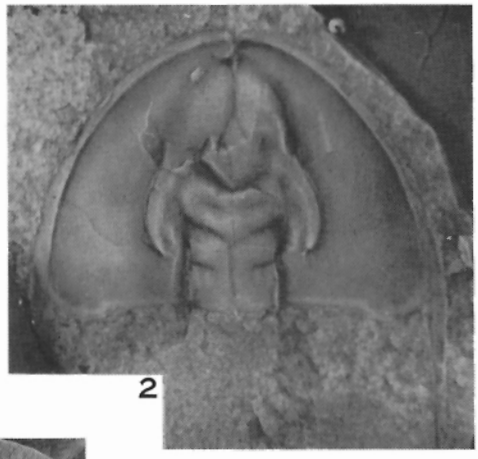

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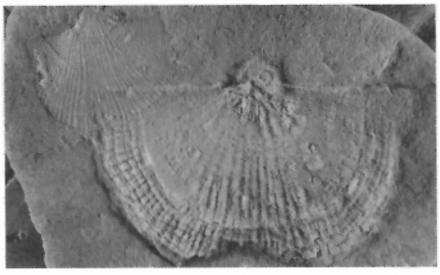

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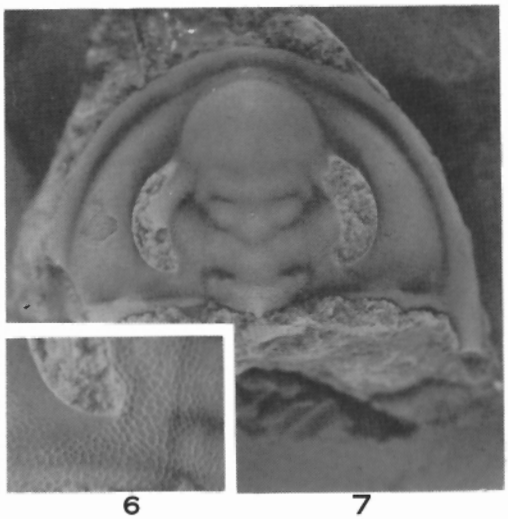

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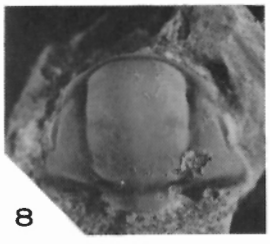

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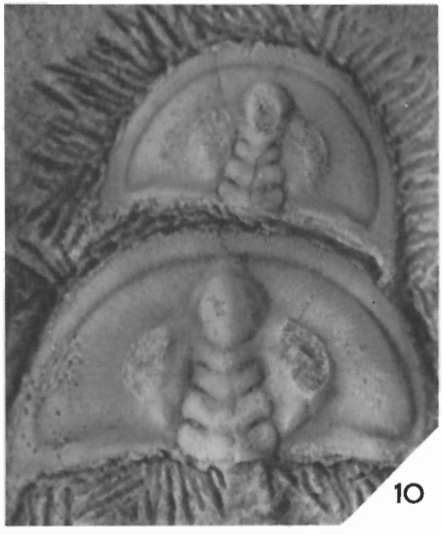

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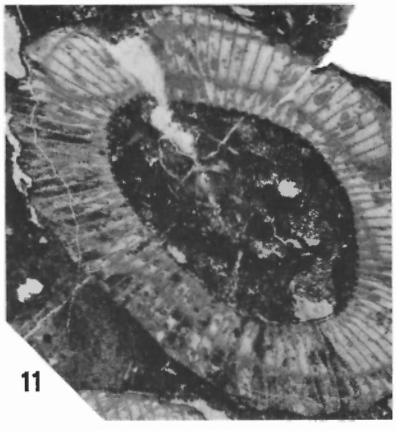

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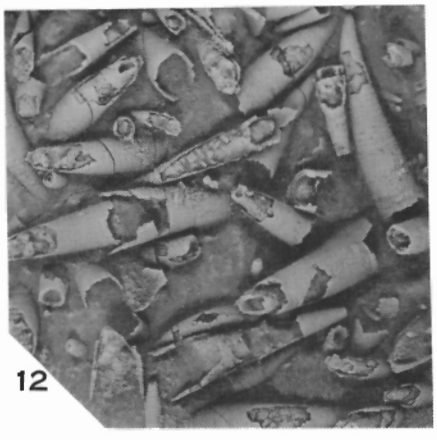

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Figures 8, 9 *Bonnia* sp. 8, Part of head x3, GSC No. 21353. 9. Tail x6, GSC No. 21354. Upper *Olenellus* Zone, Peyto Formation, Chaba River, B.C.

Figure 10 *Nevadella* sp. Two heads x2, GSC No. 21355, lower *Olenellus* Zone; formation and locality unknown.

Figure 11 *Coscinocyathus* sp. Transverse slice in thin section x2, GSC No. 13328, Lower Cambrian, Atan Group, near mile post 702, Alaska Highway, Yukon Territory.

Figure 12 *Salterella* sp. Conchs x2, GSC No. 410, Lower Cambrian, Levis Conglomerate, Point Levis, Quebec.

PLATE II. Middle Cambrian Fossils

Figure 1 *Elrathina* sp. Exoskeleton x3, GSC No. 21356, *Bathyuriscus-Elrathina* Zone, *Ogygopsis* Shale, Mount Stephen, B.C.

Figure 2 *Hyolithes* sp. Impression of shell x1, GSC No. 21357, *Albertella* Zone, Cathedral Formation, Mount Bosworth, Alberta-British Columbia.

Figure 3 *Oryctocephalus* sp. Exoskeleton x2, GSC No. 21358, *Bathyuriscus-Elrathina* Zone, *Ogygopsis* Shale, Mount Stephen, B.C.

Figures 4, 5 *Glossopleura* sp. 4, Part of head x3, GSC No. 21359. 5, Tail x2, GSC No. 21360. Cathedral Formation, Mount Assiniboine, Alberta, British Columbia.

Figures 6, 7 *Peronopsis* sp. 6, Head x9, GSC No. 21361. 7, Tail x9, GSC No. 21362. *Bathyuriscus-Elrathina* Zone?, formation unknown, Mount Field, B.C.

Figure 8 *Ogygopsis* sp. Exoskeleton, GSC No. 21363, *Bathyuriscus-Elrathina* Zone, *Ogygopsis* Shale, Mount Stephen, B.C.

Figure 9 *Micromitra* sp. Brachial valve x5, GSC No. 21364, *Bathyuriscus-Elrathina* Zone?, formation unknown, Mount Field?, B.C.

Figures 10, 11 *Pagetia* sp. 10, Head x8, GSC No. 21365. 11, Tail x8, GSC No. 21366. *Bathyuriscus-Elrathina* Zone?, formation unknown, Mount Field, B.C.

Figure 12 *Bathyuriscus* sp. Latex cast of exoskeleton x3, GSC No. 21395, *Bathyuriscus-Elrathina* Zone, *Ogygopsis* Shale, Mount Stephen, B.C.

Figure 13 *Bradoria?* sp. Left valve x6, GSC No. 21367, either late *Albertella* or early *Glossopleura* Zone, Gordon Shale, Flathead Range, Alberta.

Figure 14 *"Helcionella"*. Shell x7, GSC No. 21368, *Bolaspidella* Zone, unnamed formation, near June Lake, District of Mackenzie.

Figures 15, 16 *Plagiura* sp. 15, Part of head x2, GSC No. 21369. 16, Latex cast of tail x2, GSC No. 21370. *Plagiura-Poliella* Zone, Mount Whyte Formation, Mount Weed, Alberta.

Figures 17, 18 *Vanuxemella* sp. 17, Part of head x5, GSC No. 21371. 18, Tail x4, GSC No. 21372. *Albertella* Zone, Cathedral Formation, Mount Thompson, Alberta.

Figure 19 *Albertella* sp. Part of head and tail x5, GSC No. 21373, Cathedral Formation, Mount Thompson, Alberta.

Figure 20 *Amecephalus* sp. Latex cast of incomplete exoskeleton x3, GSC No. 21374, *Plagiura-Poliella* or *Albertella* Zone, Mount Assiniboine, Alberta-British Columbia.

PLATE III. Middle and Upper Cambrian Fossils

Figures 1, 2 *Briscoia* sp. 1, Part of head x2, GSC No. 21375. 2, Tail x1, GSC No. 21376. *Ptychaspis-Prosaukia* or lower *Saukia* Zone, McKay Formation, 3 miles west of Donald, B.C.

Figures 3, 4 *Ptychaspis* sp. 3, Latex cast of part of head x2, GSC No. 21377. 4, Tail x3, GSC No. 21378. *Ptychaspis-Prosaukia* Zone, Bison Creek Formation, Chaba River, Alberta.

Figures 5, 6 *Leiocoryphe* sp. Head in plan and rear view x5, GSC No. 21379, *Saukia* Zone, Mistaya Formation, Mount Wilson, Alberta.

Figures 7, 8 *Eurekia* sp. 7, Part of head x3, GSC No. 21380. 8, Tail x3, GSC No. 21381. *Saukia* Zone, Mistaya Formation, Mount Wilson, Alberta.

Figure 9 *Drumaspis* sp. Part of head x6, GSC No. 21382, *Ptychaspis-Prosaukia* Zone, Bison Creek Formation, Chaba River, Alberta.

Figure 10 *Irvingella* sp. Part of head x3, GSC No. 21383, *Elvinia* Zone, McKay Formation, Hughes Range, B.C., Alberta.

Figure 11 *Dytremacephalus* sp. Part of head x5, GSC No. 21384, *Aphelaspis* or *Dunderbergia* Zone, Lyell Formation, Chaba River, Alberta.

Figures 12, 13 *Aphelaspis* sp. 12, Part of head x2, GSC No. 21385. 13, Tail x3, GSC No. 21386. *Aphelaspis* Zone, Highway 1, mile post 15, west of Field, B.C.

Figures 14, 15 *Kormagnostus* sp. 14, Head x4, GSC No. 21387. 15, Tail x10, GSC No. 21388. *Cedaria* or *Crepicephalus* Zone, Waterfowl-Sullivan formational contact zone, Mount Brussilof, B.C.

Figure 16 *Bailiaspis?* sp. Head x3, GSC paratype No. 12033, Middle Cambrian, Highland Cove, Trinity Bay, Newfoundland.

Figure 17 *Conocoryphe?* sp. Head x2, GSC No. 12025, Middle Cambrian, Highland Cove, Trinity Bay, Newfoundland.

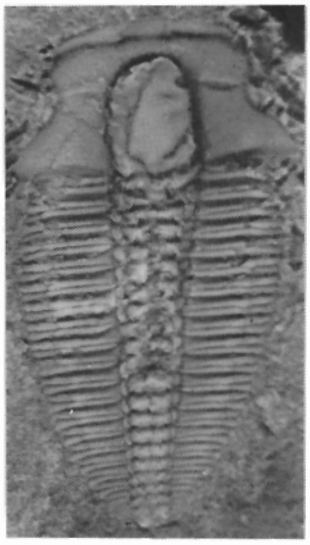
Figure 18 *Paradoxides* sp. Part of head x2, GSC No. 13069, Middle Cambrian, first small bay east of McLeod Point, Trinity Bay, Newfoundland.

Figures 19, 20 *Modocia* sp. 19, Part of head x4, GSC No. 21389. 20, Tail x6, GSC No. 21390. *Bolaspidella* Zone, Pika Formation, Beauty Creek, Jasper Park, Alberta.

Figures 21, 22 *Hypagnostus* sp. 21, Head x9, GSC No. 21391. 22, Tail x6, GSC No. 21392. *Bolaspidella* Zone, unnamed formation, near June Lake, District of Mackenzie.

Figure 23 *Crepicephalus* sp. Latex cast of part of head and tail x2, GSC No. 28393, *Crepicephalus* Zone, Sullivan? Formation, Ram Range, Alberta.

Figure 24 *Cedaria* sp. Part of head and tail x2, GSC No. 21394, *Cedaria* Zone, Upper Chancellor Formation or lower Ottertail Formation, Wolverine Pass, B.C.



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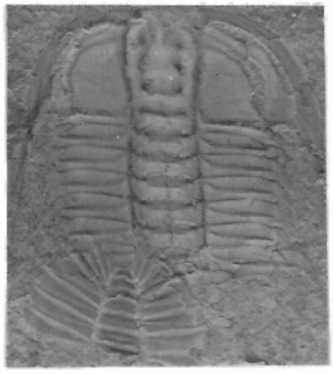

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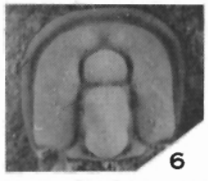

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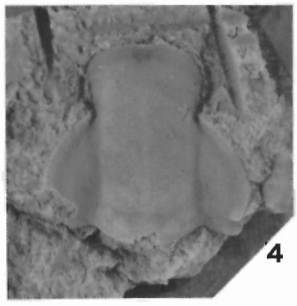

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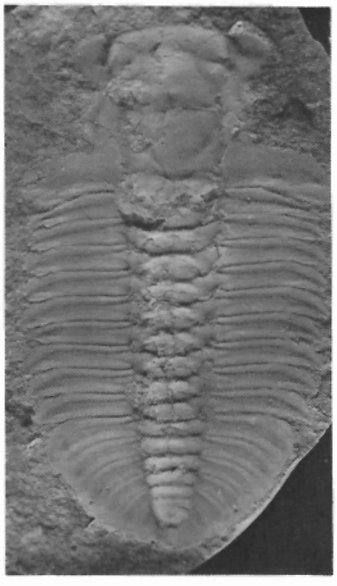

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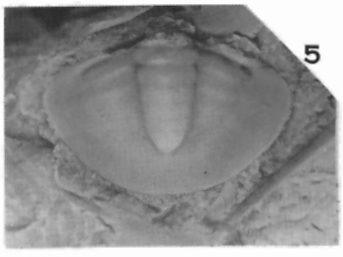

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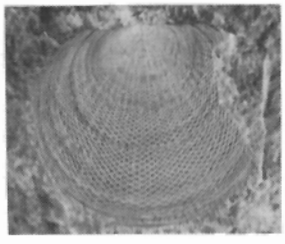

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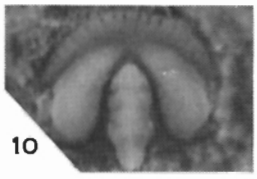

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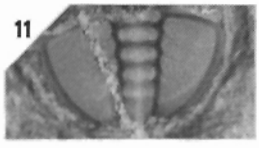

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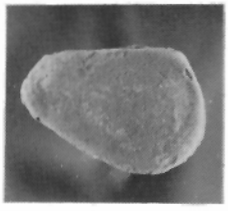

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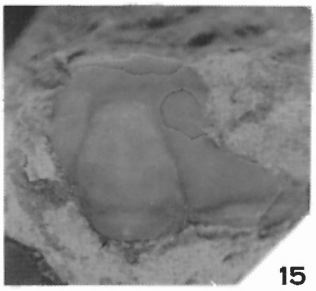

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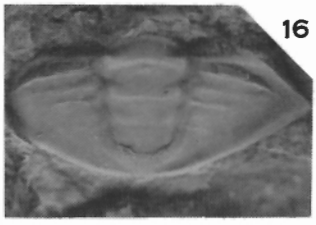

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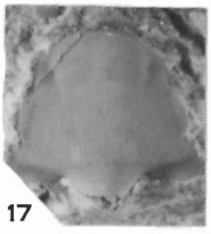

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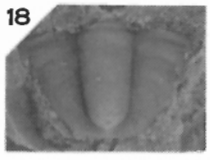

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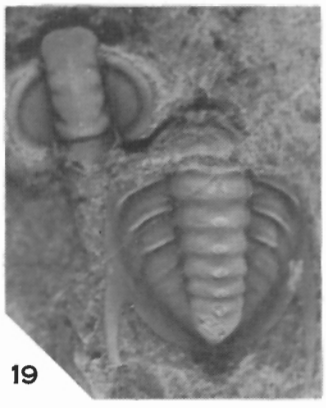

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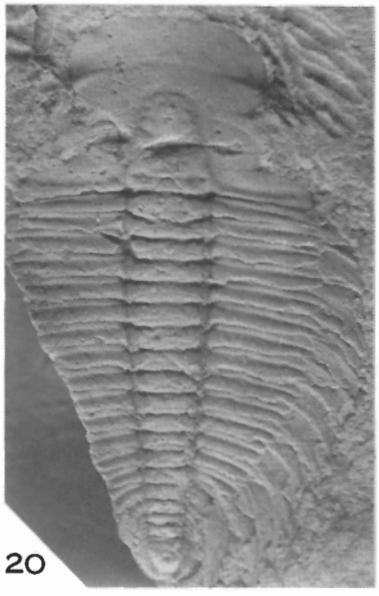

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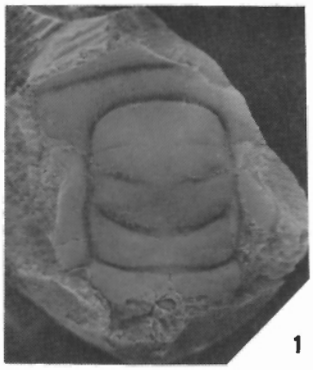
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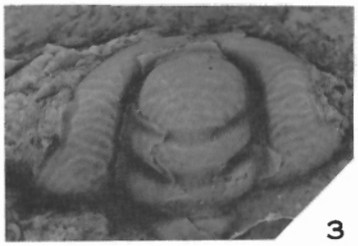
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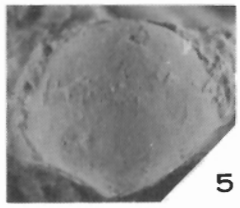
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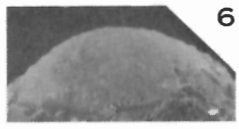
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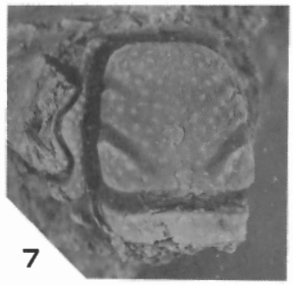
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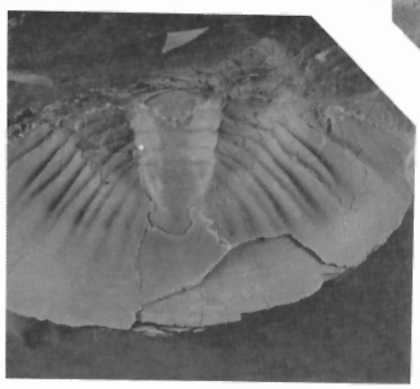
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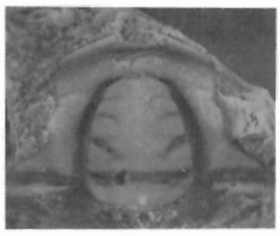
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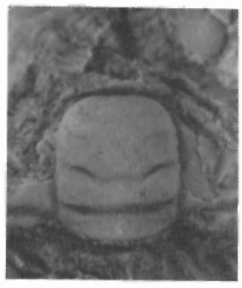
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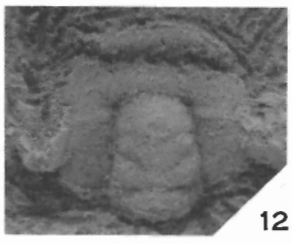
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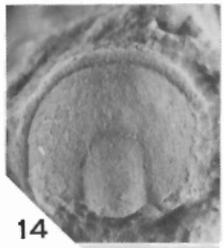
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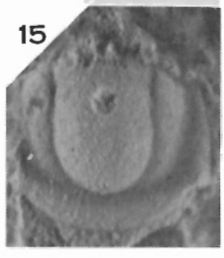
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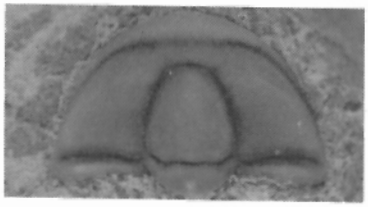

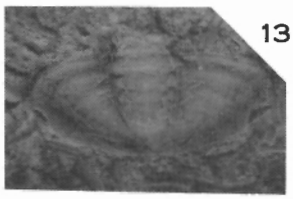
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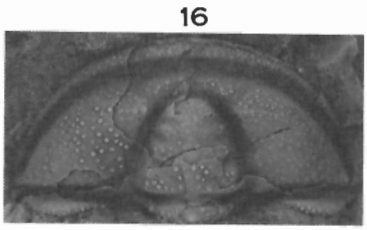
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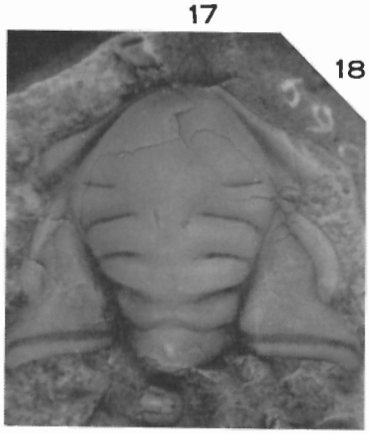
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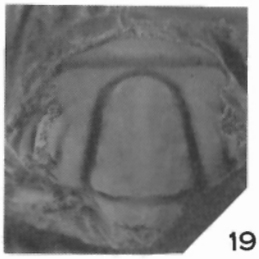
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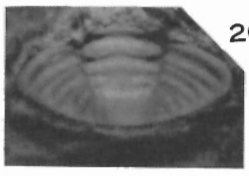
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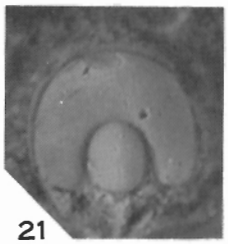
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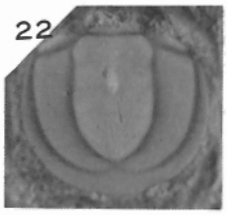
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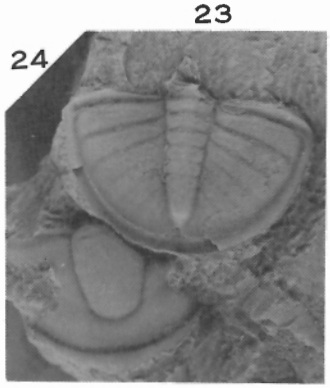
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ORDOVICIAN AND SILURIAN FAUNAS

by B. S. Norford, T. E. Bolton, M. J. Copeland,
L. M. Cumming, and G. W. Sinclair

Pelagic Faunas of the Ordovician and Silurian

Worldwide correlation of the Ordovician and Silurian Systems is based on graptolites. Graptolites were pelagic animals; their remains are most abundant and best preserved in rocks deposited in deep water. Rocks of graptolitic facies accumulated seaward of the belts of platform carbonates, sandstones, and siltstones that bordered the North American landmass in early Paleozoic time. Some sequences of graptolitic rocks seem to represent almost continuous deposition throughout the Ordovician and Silurian. Radioactive dating suggests that the duration of the combined Ordovician and Silurian interval was about 100 million years. This interval has been divided into thirty-nine graptolite zones.

The composite sequence of zones shown on Tables XI-2 and XI-3 is based on successions of graptolite faunas in various parts of Canada (Jackson, 1964; Jackson and Lenz, 1962; Thorsteinsson, 1958). The zonal successions are best developed in the Road River, Kechika, and Glenogle units of the Cordillera, and in the Ibbett Bay and Cape Phillips Formations of the arctic islands (Fig. XI-1); all within geosynclines but close to regions of platform sedimentation. Sparse occurrences of graptolites are known from the platform regions and aid in correlation of associated benthonic faunas with the standard graptolite zonation. A few localities in the Rocky Mountains, the northern Yukon, and the arctic islands show interfingering of graptolitic rocks with rocks bearing benthonic faunas.

Bulman (1958) has shown that the sequence of graptolite zones can be grouped into four successive gross faunas that are characterized by generic assemblages, in contrast to the assemblages of species that form the zones. *Anisograptus, Bryograptus, Clinograptus, Dictyonema, Staurograptus,* and *Triograptus* are the dominant genera of the Anisograptid fauna of Tremadoc age. The Arenig Dichograptid fauna consists primarily of the genera *Dichograptus, Didymograptus, Isograptus, Oncograptus, Phyllograptus,* and *Tetragraptus.* The Diplograptid fauna lasted from Llanvirn to early Llandovery and is characterized by a diversity of biserial forms. Common genera include *Amplexograptus, Climacograptus, Cryptograptus, Dicellograptus, Dicranograptus, Didymograptus, Diplograptus, Glossograptus, Glyptograptus, Leptograptus, Nemagraptus,* and *Orthograptus.* In contrast, the Monograptid fauna has one dominant genus, *Monograptus.* Subordinate genera include *Cyrtograptus, Linograptus, Rastrites,* and *Retiolites.* The fauna ranged from early Llandovery to early Devonian time.

Benthonic Faunas of the Ordovician and Silurian

The benthonic faunas of the carbonates and detrital rocks of the platform consisted primarily of corals, echinoderms, brachiopods, molluscs, trilobites, stromatoporoids, bryozoans, and ostracods. Graptolites were extremely rare. Corals were not present in Canadian and Whiterock faunas but were very important in younger Ordovician and Silurian faunas.

Benthonic Ordovician of Eastern Canada

In Ontario and western Quebec the Canadian Series is represented by dolomites with few fossils. The gastropod *Ophileta* is present. A large and varied Canadian fauna does occur in fossiliferous limestones and shales farther east in Quebec (Eastern Townships, Lévis, and the Mingan Islands), and in western Newfoundland. Common and characteristic forms are the brachiopods *Clarkella* and *Finkelnburgia,* the gastropods *Ceratopea* and *Euconia,* the cephalopods *Kirkoceras* and *Piloceras,* and the trilobites *Leiostegium, Pilekia,* and *Shumardia.*

Stages for the interval between the Canadian and Wilderness have recently been revised in the United States and their precise representation in Canada has yet to be established. The Table Head sequence of western Newfoundland is of Whiterock age and contains the trilobites *Ectenonotus, Eorobergia, Galbagnostus, Miracybele,* and *Xystocrania.* The Chazyan formations are considered to be of Marmor age (Cooper, 1956). These rocks are exposed near Ottawa and Montreal, on the Mingan Islands, and in Newfoundland. The first appearance of large numbers of bryozoans, echinoderms, and corals in the geological column is in the Marmor Stage. Common genera are the echinoderms *Blastoidocrinus, Bolboporites,* and *Malocystites,* the corals *Billingsaria* and *Lamottia,* the brachiopods *Mimella* and *Rostricellula,* the gastropod *Scalites,* and the conulariid *Conularina.* Some strata in the Mingan Islands and in Newfoundland that previously have been considered Chazyan may be of post-Marmor age.

Beds of Wilderness age occur throughout southern Ontario and Quebec and are commonly very fossiliferous. The earliest fauna is found in the Gull River Formation of Ontario and includes *Cryptophragmus* and abundant but virtually unstudied molluscs. The fauna of the Lowville Formation and its equivalents is characterized by *Bathyurus extans* and many cephalopod species and is known throughout southern Ontario, and near Montreal and Lake St. John. Later Wilderness faunas are present

in the Chaumont and Rockland Formations and their equivalents throughout the Ottawa–St. Lawrence Valley from Lake Timiskaming to Murray Bay, in the Saguenay Valley, and in Newfoundland. These faunas include the echinoderms *Amecystis, Carabocrinus huronensis,* and *Foerstediscus;* the coral *Lambeophyllum;* the brachiopods *Doleroides, Hesperorthis, Opikina,* and *Strophomena filitexta;* the trilobites *Raymondites ingalli* and *Thaleops;* the ostracods *Eoleperditia* and *Pteroleperditia;* and the sponge *Receptaculites.*

The Barneveld is the best known and most fossiliferous of the North American Ordovician stages, containing as it does the formations of the Trenton Group of old usage. Characteristic fossils are the coral *Protaraea vetusta;* the sponges *Brachiospongia* and *Steliella;* the echinoderms *Cleiocrinus, Edrioaster, Glyptocrinus, Lebetodiscus,* and *Pleurocystites;* the brachiopods *Parastrophina, Platystrophia, Rafinesquina trentonensis, Sowerbyella "sericea",* and *Vellamo;* the gastropods *Fusispira, Hormotoma trentonensis,* and *Trochonema;* the cephalopod *Endoceras proteiforme;* the trilobites *Calliops, Ceraurus pleurexanthemus, Cryptolithus tessellatus,* and *Flexicalymene senaria;* and the bryozoans *Arthroclema* and *Prasopora.*

The boundary between the Eden and Maysville Stages is not firmly established in Canada, and lumping of the combined interval may prove to be useful. In eastern Canada, rocks correlated as Eden are mainly shales and argillaceous limestones (e.g., Hillier, Collingwood, and Terrebonne units) with such fossils as the echinoderm *Cheirocrinus;* the brachiopods *Cyclospira, Oxoplecia,* and *Plectatrypa;* the cephalopods *Ephippiorthoceras* and *Probillingsites;* and the trilobites *Pseudogygites, Triarthrus eatoni, T. glaber,* and *T. spinosus.* In contrast the Maysville is mostly represented by sandy beds with abundant pelecypods such as *Byssonychia, Rhytimya,* and *Whitella,* the brachiopods *Leptaena moniquensis* and *Zygospira modesta,* and trilobites (including *Proetus chambliensis*).

The Richmond Stage is known from widely scattered areas of eastern Canada: Manitoulin Island, southern Ontario, Eastern Townships, Lake St. John, Gaspé, and Anticosti Island. Common fossils include the coral

TABLE XI-2 *Ordovician faunal successions*

EUROPEAN STAGES	BENTHONIC ZONES Western & Northern Canada	PELAGIC ZONES	COMMON BENTHONIC FOSSILS Eastern Canada, north and west of Logan's Line	NORTH AMERICAN STAGES
ASHGILL	*Bighornia - Thaerodonta*	*Dicellograptus complanatus ornatus*	*Grewingkia rusticum* *Schuchertoceras* *Rhombopora quadrata* *Catazyga headi* *Billingsites* *Beatricea*	RICHMOND
		Orthograptus n. sp. A		
CARADOC	Red River fauna	*Orthograptus quadrimucronatus*	*Leptaena moniquensis* *Whitella complanata* *Byssonychia radiata* *Cryptolithus bellulus* *Isotelus maximus* *Proetus chambliensis*	MAYSVILLE
			Cyclospira bisulcata *Lambeoceras* *Triarthrus spinosus* *Paleschara beani* *Pseudogygites* *Probillingsites primus*	EDEN
			Platystrophia Vellamo *Onniella whittakeri* *Lebetodiscus dicksoni* *Prasopora selwyni* *Ceraurus dentatus* *Cryptolithus tessellatus*	BARNEVELD
	Faunas of the Sunblood Formation probably span this interval	*Orthograptus truncatus intermedius*	*Sowerbyella curdsvillensis* *Amecystis laevis* *Doleroides Opikina* *Hesperorthis tricenaria* *Gonioceras* *Tetradium cellulosum* *Receptaculites occidentalis* *Bathyurus extans* *Raymondites ingalli*	WILDERNESS
		Climacograptus bicornis		
		Nemagraptus gracilis		PORTERFIELD
LLANDEILO		*Glyptograptus euglyphus*		ASHBY
			Lamottia *Pliomera* *Mimella* *Sigmacystis* *Rostricellula plena* *Scalites* *Blastoidocrinus* *Malocystites* *Bolboporites americanus*	MARMOR
LLANVIRN	*Anomalorthis*	*Paraglossograptus etheridgei*	*Miracybele* *Xystocrania* *Ampyxoides* *Eoleperditia "bivia"*	WHITEROCK
	Orthidiella			
ARENIG	*Hesperonomia*	*Isograptus caduceus*		
	J *(Pseudocybele)*			
	I *(Presbynileus)*	*Didymograptus protobifidus*		
	H *(Trigonocerca)*			
	G_2 *(Protopliomerella)*	*Tetragraptus fruticosus*	*Finkelnburgia, Tritoechia* *Lingula quebecensis* *Ecclioptéris planidorsalis* *Ellesmeroceras Piloceras* *Shumardia Euconia* *Ceratopea Ophileta*	CANADIAN
		Tetragraptus approximatus		
TREMADOC	G_1 *(Hintzea)*	*Adelograptus-Clonograptus*		
	F *(Rossaspis)*			
	E *(Tesselacauda)*			
	D *(Leiostegium-Kainella)*			
	C *("Paraplethopeltis")*			
	B *(Bellefontia-Xenostegium)*			
	A *(Symphysurina-Euloma)*	*Staurograptus*		
		Dictyonema		

GSC

Faunas poorly known or unknown

Grewingkia; the brachiopods *Catazyga headi* and *Strophomena fluctuosa;* the gastropod *Lophospira beatrice;* and the cephalopods *Apsidoceras, Manitoulinoceras,* and *Schuchertoceras.*

Benthonic Ordovician of Western and Northern Canada

Many fossiliferous rock-units are present in western and arctic Canada but few Ordovician and Silurian fossils have been described. Most of the trilobite zones that have been established for the Canadian Stage of Utah and Nevada can also be recognized in the McKay, Survey Peak, and Outram units of the southern Rocky Mountains of Canada. Some of these zones are also present in the Chushina Formation near Jasper and near Mount Robson, in the Kechika Group of northern British Columbia, in unnamed rocks in the northern Yukon and southwest District of Mackenzie, and probably also in the Mingo River and Nadlo Point Formations of Devon Island.

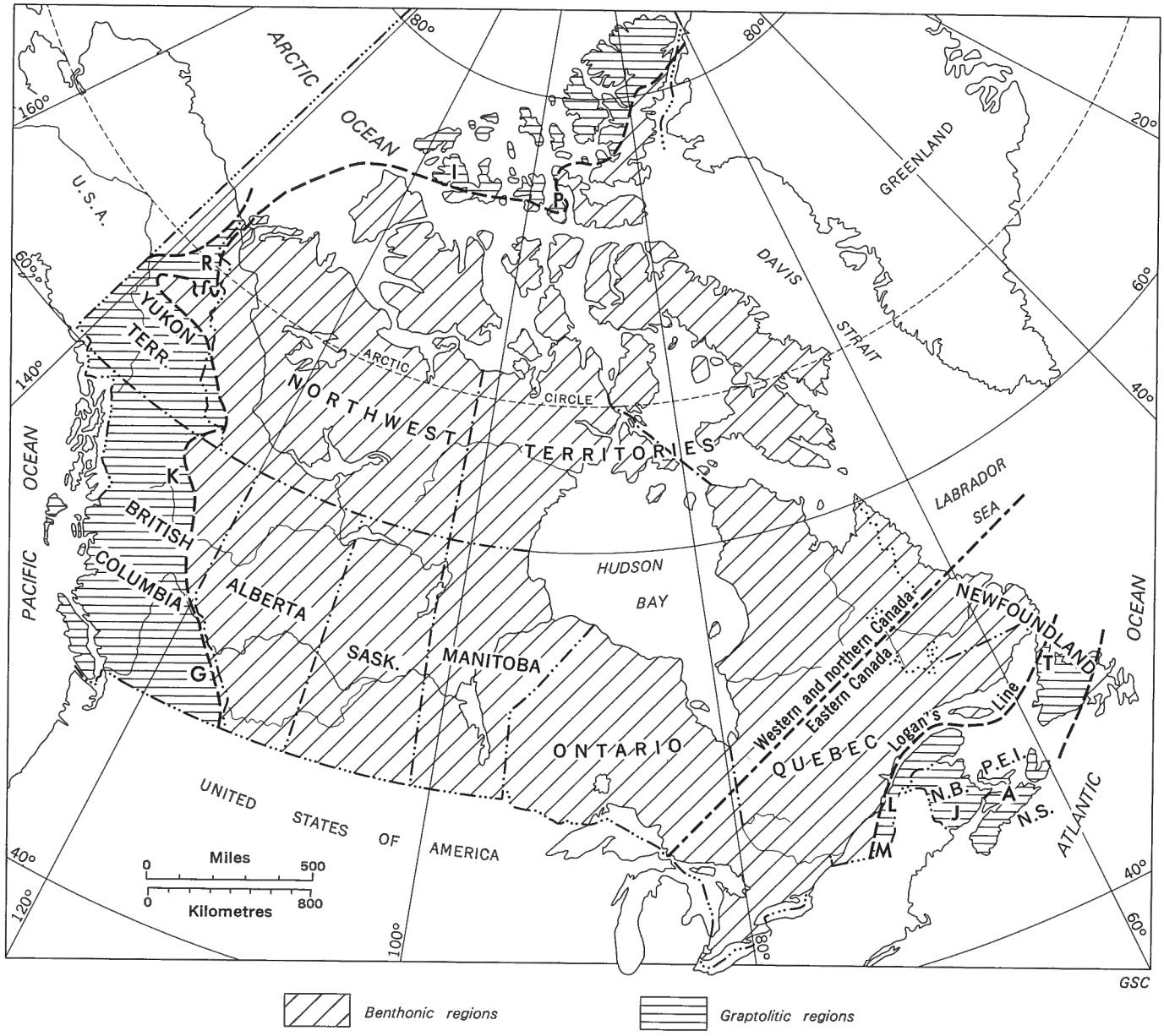
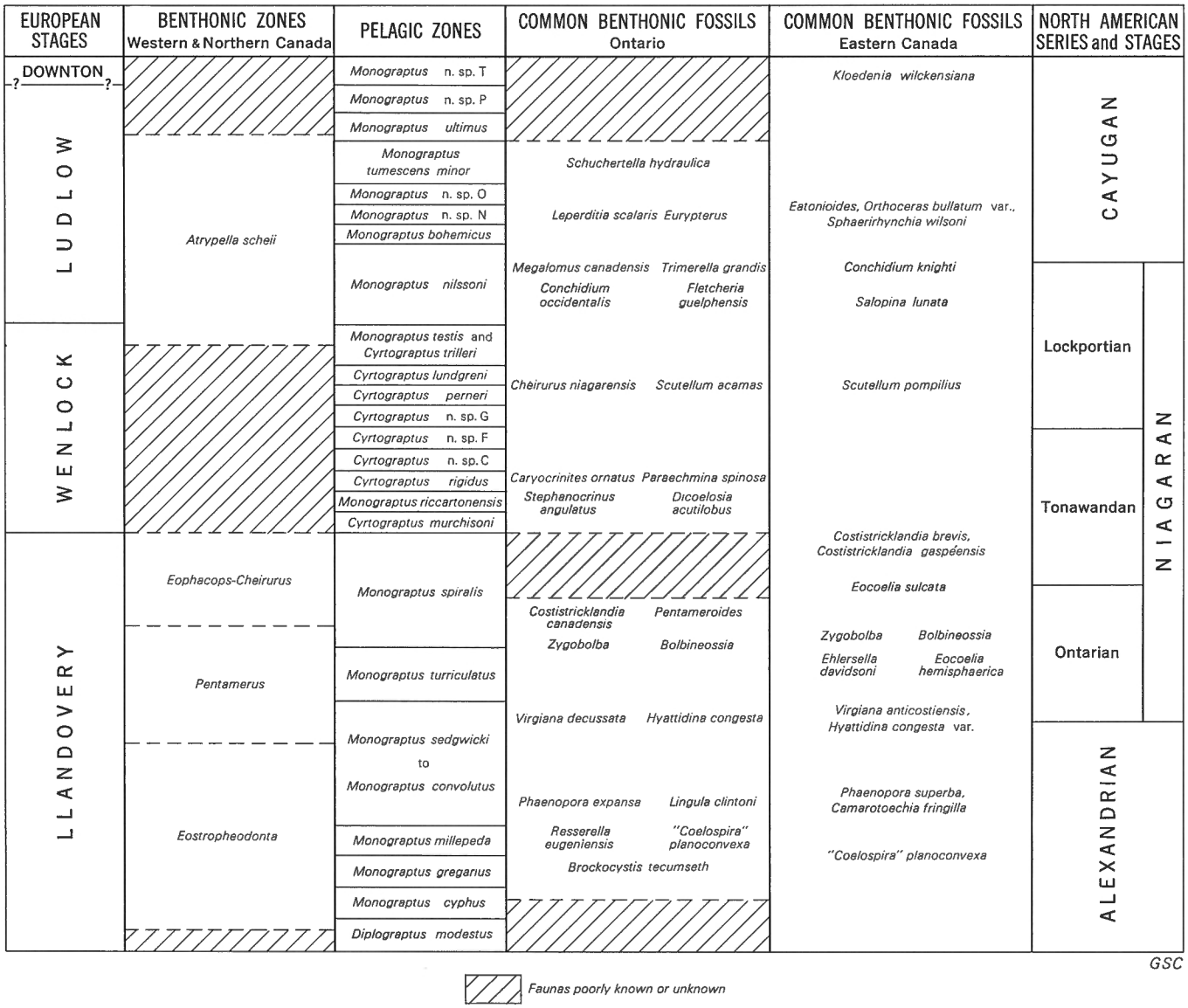


FIGURE XI-1. Distribution of Ordovician and Silurian faunas.

TABLE XI-3 *Silurian faunal successions*



The uppermost Canadian *Hesperonomia* and the Whiterock *Orthidiella* and *Anomalorthis* brachiopod assemblage zones are well developed in the Skoki Formation of southwest Alberta. The *Orthidiella* Zone is probably also present in eastern British Columbia at Cecilia Lake and near Peace River. The *Hesperonomia* Zone includes the brachiopods *Diparelasma, Leptella, Syntrophopsis, Taffia,* and *Tritoechia.* Species of *Diparelasma, Orthambonites,* and *Petroria* are found in the *Orthidiella* Zone. Genera in the *Anomalorthis* Zone include *Receptaculites,* the gastropods *Maclurites* and *Palliseria,* and the brachiopod *Syntrophopsis.*

Fossiliferous Marmor to Maysville rocks are rare in western Canada. The most complete sequence seems to be in the southwest District of Mackenzie where the Sunblood Formation contains a succession of well-preserved brachiopod and trilobite faunas. This succession probably spans the entire interval but the faunas are virtually unstudied. Faunules from high in the Kechika Group of northern British Columbia, from the Admiralty and Cornwallis units of the arctic islands and from the Bad Cache Rapids and Red River units of Manitoba probably correlate with faunules in the upper Sunblood. Many faunules in arctic Canada have been referred to "the Arctic Ordo-

vician fauna"; most of these faunules correlate with the fauna of the Red River Formation but some suggest the younger *Bighornia–Thaerodonta* Zone. The Red River fauna includes species of *Receptaculites;* the corals, *Calapoecia, Catenipora, Tollina, Palaeofavosites, Palaeophyllum, Grewingkia,* and *Favistina;* the brachiopods *Rafinesquina,* and *Rhynchotrema;* the gastropods *Fusispira, Hormotoma,* and *Maclurites,* and the cephalopods *Diestoceras, Cyrtogomphoceras, Lambeoceras,* and *Winnipegoceras.*

Richmond faunas are widely distributed in western Canada and can mostly be referred to the *Bighornia–Thaerodonta* assemblage zone that is well developed low in the Beaverfoot Formation of the southern Rockies. Other typical genera are the corals *Catenipora, Favistina, Lobocorallium, Palaeofavosites, Palaeophyllum, Sarcinula,* and *Streptelasma,* and the brachiopods *Dinorthis* and *Rhynchotrema.* Such faunas are present in the lower part of the Sandpile Group of northern British Columbia; in the Chedabucto Lake and lower Mount Kindle Formations of the District of Mackenzie; in unnamed rocks in northeast British Columbia; the Yukon, and the District of Mackenzie; and in the Stony Mountain, Stonewall, and Churchill River units of Manitoba.

Benthonic Silurian of Eastern Canada

In Ontario, the Alexandrian Series consists of the Whirlpool, Power Glen, Manitoulin, Cabot Head, and Grimsby Formations. Characteristic species include the coral *Palaeofavosites asper;* the echinoderm *Brockocystis tecumseth;* the brachiopods *Resserella eugeniensis, "Coelospira" planoconvexa,* and *Lingula clintoni;* the bryozoans *Helopora fragilis* and *Phaenopora expansa;* the trilobite *Liocalymene clintoni;* and the trail *Arthrophycus alleghaniensis.* On Anticosti Island, 15 feet of the Becscie Formation underlies the *Virgiana* beds and may be the eastern representative of the Alexandrian. The fauna includes *"Coelospira" planoconvexa* and *Phaenopora superba.* Alexandrian faunas are present in the Beechhill Formation (with the brachiopod *Eostropheodonta*) and the lower part of the Ross Brook Formation of Arisaig, in Gaspé, and in the Cabano Formation (also with *Eostropheodonta*) and possibly also in the overlying Pointe-aux-Trembles Formation that carries *Eocoelia quebecensis.*

Characteristic fossils of the Ontarian Stage include the brachiopods *Virgiana, Hyattidina, Pentameroides, Camarotoechia winiskensis, Plectatrypa lowi, Costistricklandia canadensis, Eocoelia hemisphaerica,* and *Ehlersella davidsoni;* and the ostracods *Zygobolba* and *Bolbineossia.* Such fossils are found in the Reynales, Dyer Bay, Wingfield, St. Edmund, Wabi, and Thornloe Formations of Ontario; the Becscie, Gun River, and Jupiter Formations of Anticosti Island; part of the Whitehead Formation at Percé; the Awantjish, Burnt Jam Brook, and Clemville Formations of Gaspé; the Quoddy Formation of New Brunswick; the middle member of the Ross Brook Formation of Nova Scotia; and the Pike Arm Formation of Newfoundland.

The brachiopods *Dicoelosia acutilobus, D. biloba, Eocoelia sulcata, Costistricklandia brevis,* and *C. gaspéensis;* the echinoderms *Stephanocrinus angulatus, Caryocrinites ornatus,* and the ostracod *Paraechmina spinosa* are present in faunas of the Tonawandan Stage, found in formations such as the Rochester of southern Ontario; the Val Brillant, La Vieille, and Laforce of Gaspé; the Long Reach of New Brunswick; and in the upper member of the Ross Brook Formation of Nova Scotia.

Benthonic species of the Lockportian Stage include the coral *Fletcheria guelphensis;* the pelecypod *Megalomus canadensis;* the brachiopods *Salopina lunata, Trimerella grandis, Conchidium knighti,* and *C. occidentalis;* and the trilobites *Scutellum acamas, S. pompilius,* and *Cheirurus niagarensis.* Lockportian faunas are found in the Lockport, Amabel, and Guelph Formations of Ontario; in the Sayabec, Gascons, Bouleaux, West Point, Robitaille, Mont Wissick, and Sargent Bay Formations of Quebec; in the Jones Creek and Pembroke Formations of New Brunswick; and in the Kentville Formation and the upper member of the McAdam Formation of Nova Scotia.

Faunas of the Cayugan Series are not widespread and generally consist of few species. Typical forms include the ostracod *Leperditia scalaris,* the brachiopod *Schuchertella hydraulica,* and eurypterids in the Bertie–Akron sequence of Ontario; the brachiopods *Eatonioides* and *Sphaerirhynchia wilsoni* and a variety of the cephalopod *Orthoceras bullatum* in the Moydart Formation, and the ostracod *Kloedenia wilckensiana* in the Stonehouse Formation, both of Nova Scotia.

Benthonic Silurian of Western and Northern Canada

A sequence of Llandovery benthonic zones is present in the Beaverfoot Formation of the southern Rocky Mountains. The *Eostropheodonta* zone contains the brachiopod *Hesperorthis;* the corals *Catenipora, Favosites,* and *Streptelasma;* and the trilobites *Eophacops* and *Encrinurus.* Genera of the *Pentamerus* zone include the corals *Cystihalysites, Favosites,* and *Streptelasma;* and the ostracods *Aparchites, Bythocypris,* and *Tubulibairdia.* The *Eophacops–Cheirurus* zone contains *Encrinurus,* the brachiopods *Glassia* and *Atrypa,* and *Favosites.* Faunas corresponding to the interval *Pentamerus* to *Eophacops–Cheirurus* are widespread in western Canada: Silurian formations of northern British Columbia; much of the Allen Bay Formation of the arctic islands; the *Aulacopleura socialis* beds of the Road River Formation of the northern Yukon; most of the Silurian formations of the Hudson Bay and James Bay Lowlands; the Cedar Lake and East Arm Formations of southern Manitoba. Faunas of the Fisher Branch to Atikameg interval in Manitoba and of the Port Nelson and Severn River limestones of the Hudson Bay Lowland may correlate with the *Eostropheodonta* zone. The Attawapiskat Formation of the Hudson Bay Lowland may be high Llandovery or low Wenlock. Upper

Llandovery and lower Wenlock fish are present in the arctic islands.

Rocks of Wenlock and Ludlow age are missing from most of western Canada, probably as a result of Late Silurian and Early Devonian erosion. Wenlock and Ludlow rocks are found in the Yukon and the Northwest Territories but their benthonic faunas are poorly known. The *Atrypella scheii* zone has been reported from the Read Bay, Douro, Gossage, and Cape Rawson units, and from unnamed rocks near the Alaska–Yukon Boundary. Genera of this zone include the brachiopods *Conchidium* and *Howellella;* the corals *Favosites, Parafavosites, Striatopora,* and *Syringopora;* the trilobites *Encrinurus* (*Frammia*) and *Hemiarges;* the clam *Megalomus,* and eurypterids. Ludlow fish have been found in the Read Bay Formation and Ludlow and Downton fish and graptolites in the Delorme Formation.

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PLATE IV. Ordovician Fossils of Eastern Canada

(All figures natural size unless otherwise stated)

Figure 1 *Bathyurus superbus* Raymond. Wilderness; Ontario; 7422a.
Figure 2 *Tetradium clarki* Okulitch. x5.5; Wilderness; Ontario; 7397c.
Figure 3 *Ophileta compacta* (Salter). Canadian; Ontario; 17088.
Figure 4 *Receptaculites occidentalis* Salter. Wilderness; Ontario; 1125p.
Figure 5 *Malocystites murchisoni* Billings. Marmor; Quebec; 1012a.
Figure 6 *Selenoharpes ottawaensis* (Billings). Barneveld; Ontario; 329.
Figure 7 *Hesperorthis tricenaria* (Conrad). Wilderness; Ontario; 1151.
Figure 8 *Lingula cobourgensis* Billings. Eden; Ontario; 1635a.
Figure 9 *Amecystis laevis* (Raymond). Wilderness; Ontario; 7936.
Figure 10 *Onniella whittakeri* (Raymond). x2; Barneveld; Ontario; 3240a.
Figure 11 *Pleurocystites* cf. *P. filitextus* Billings. Barneveld; Ontario; 17696.
Figure 12 *Cupulocrinus jewetti* (Billings). Wilderness; Ontario; 7784z.
Figure 13 *Edrioaster* sp. x2; Barneveld; Ontario; 17087.
Figure 14 *Probillingsites cobourgensis* Flower. Eden; Ontario; 11092.
Figure 15 *Fusispira angusta* Ulrich and Scofield. Barneveld; Quebec; 13247.
Figure 16 *Flexicalymene granulosa* (Foerste). x4; Maysville; Ontario; 18667.
Figure 17 *Catazyga erratica* (Hall). Maysville; Ontario; 18670.
Figure 18 *Cryptolithus bellulus* (Ulrich). Eden; Ontario; 18669.
Figure 19 *Platystrophia clarksvillensis* Foerste. Richmond; Ontario; 18672.
Figure 20 *Glyptorthis insculpta manitoulinensis* Foerste. Richmond; Ontario; 6784.
Figure 21 *Pseudogygites latimarginatus* (Hall). Eden; Ontario; 18666.
Figure 22 *Rafinesquina sardesoni* Salmon. Eden; Ontario; 3256.
Figure 23 *Triarthrus spinosus* Billings. Eden; Ontario; 18668.
Figure 24 *Byssonychia grandis* Ulrich. Richmond; Ontario; 2120.

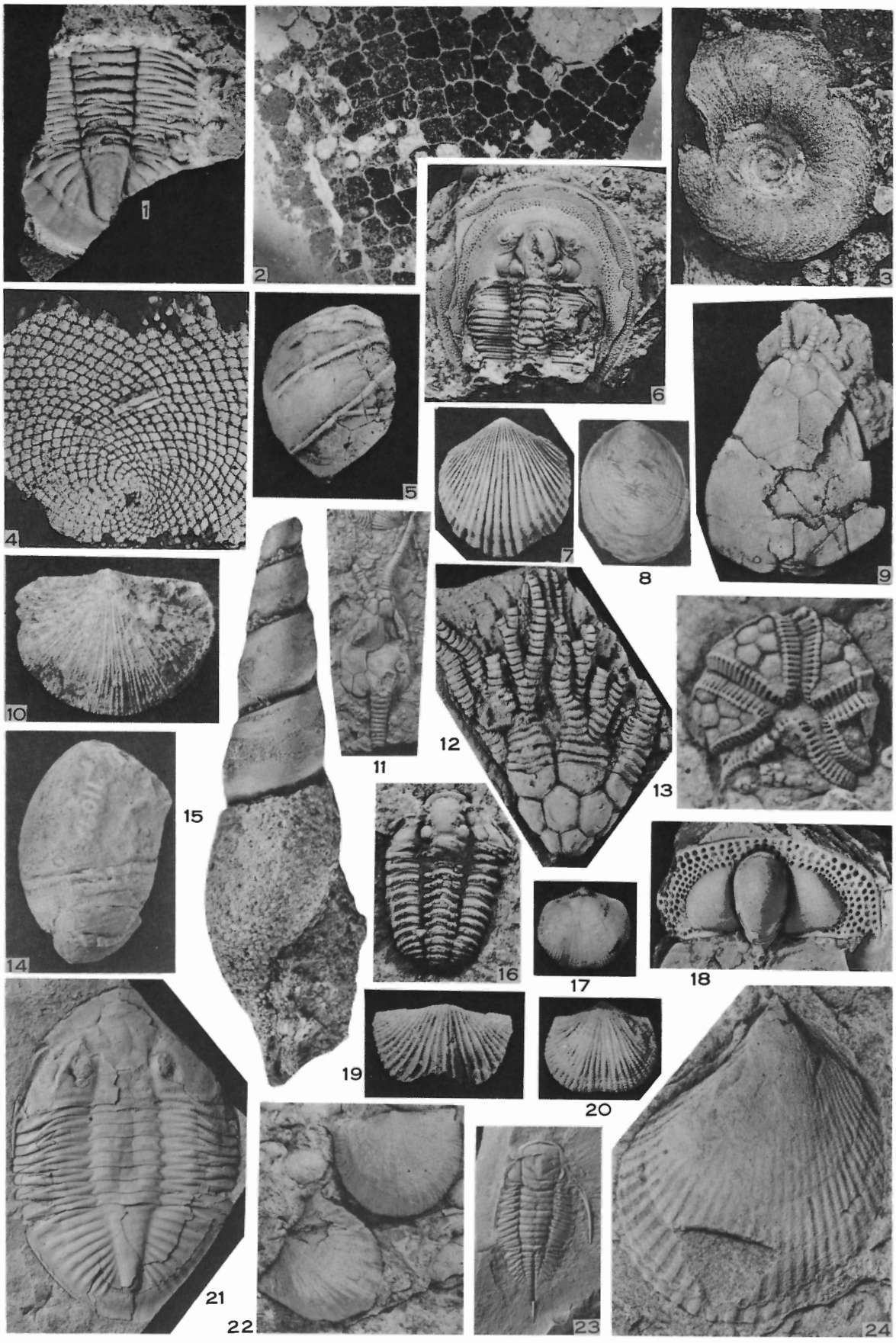
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PLATE V. Ordovician Fossils of Western and Northern Canada

(All figures natural size unless otherwise stated)

Figure 1 *Lobocorallium prolongatum* (Wilson). Richmond; Alberta; 16917.

Figure 2 *Dinorthis rockymontana* Wilson. Richmond; Alberta; 16885.

Figures 3, 11 *Bighornia parva* Duncan. Richmond; Wyoming; USNM 127574 (x3) and 124801.

Figures 4, 10 *Dinorthis columbia* Wilson. Richmond; Alberta; 16892.

Figures 5, 8, 18 *Rhynchotrema kananaskia* Wilson. Richmond; British Columbia; 6749.

Figures 6, 7, 9 *Rhynchotrema windermeris* Wilson. Richmond; Alberta; 16895 and 16896.

Figures 12, 13, 14 *Thaerodonta* sp. aff. *T. saxea* (Sardeson). Richmond; Alberta; 16887, 16888, and 16889.

Figures 15, 16 *Calyptaulax* sp. Wilderness; Mackenzie; 16872 and 16873.

Figure 17 *Tollina* sp. Richmond; Mackenzie; 16915.

Figures 19, 20 *Palaeophyllum* sp. Richmond; Mackenzie; 16910 and 16911.

Figures 21, 22 *Leiostegium* sp. x2; Canadian; Mackenzie; 20432 and 20433.

Figure 23 *Oistodus lanceolatus* Pander. x50; Whiterock; Alberta; 20434.

Figure 24 *Catenipora robustus* (Wilson). Richmond; Alberta; 16907.

Figures 25, 26 *Opikina* sp. Wilderness; Mackenzie; 16876 and 16875.

Figures 27, 28 *Maclurites* sp. cf. *M. magnus* Lesueur. Whiterock; Mackenzie; 16883 and 16884.

Figures 29, 30 *Bellefontia* sp. Canadian; British Columbia; 20435 and 20436.

Figures 31, 32 *Kainella billingsi* Walcott. Canadian; British Columbia; USNM 70334 and 70336.

PLATE VI. Ordovician and Silurian Graptolites

(All figures natural size unless otherwise stated)

Figure 1 *Monograptus priodon* (Bronn). x2; Llandovery; Scotland; figure from Bulman (1955).

Figure 2 *Diplograptus* cf. *D. modestus* Lapworth. x5; Llandovery; Iowa; figure from Ross (1964).

Figure 3 *Monograptus turriculatus* (Barrande). x2; Llandovery; Bohemia; figure from Bulman (1955).

Figure 4 *Linograptus* sp. x3; Ludlow; Silesia; figure from Bulman (1955).

Figures 5, 6 *Monograptus bohemicus* (Barrande). x4; Ludlow; Illinois; figures from Ross (1962).

Figure 7 *Climacograptus scharenbergi* Lapworth. x6; Llandeilo-Caradoc; Scotland; figure from Bulman (1955).

Figure 8 *Cyrtograptus murchisoni* Carruthers. x2; Wenlock; Bohemia; figure from Bulman (1955).

Figure 9 *Monograptus convolutus* (Hisinger). x2; Llandovery; Sweden; figure from Bulman (1955).

Figure 10 *Climacograptus bicornis* (Hall). x2; Caradoc; Scotland; figure from Bulman (1955).

Figure 11 *Retiolites* (*Stomatograptus*) *grandis grandis* Suess. x2; Llandovery; Bohemia; figure from Bouček and Münch (1943).

Figure 12 *Didymograptus* sp., pendent species. Llanvirn; Wales; figure from Bulman (1955).

Figure 13 *Dicellograptus morrisi* Hopkinson. x4; Caradoc; Sweden; figure from Bulman (1955).

Figure 14 *Dicranograptus nicholsoni* Hopkinson. x4; Caradoc; Scotland; figure from Bulman (1955).

Figure 15 *Monograptus spiralis spiralis* (Geinitz). Llandovery; Bohemia; figure from Přibyl (1944).

Figure 16 *Phyllograptus angustifolius* Hall. Ventral view, x4; Arenig; Sweden; figure from Bulman (1955).

Figure 17 *Didymograptus spinosus* Ruedemann. x2; Llanvirn; Yukon; figure from Jackson (1964).

Figure 18 *Tetragraptus quadribrachiatus* (Hall). Arenig; Quebec; figure from Bulman (1955).

Figure 19 *Nemagraptus gracilis* (Hall). Caradoc; Scotland; figure from Bulman (1955).

Figure 20 *Phyllograptus typus* Hall. Arenig; Quebec; figure from Bulman (1955).

Figure 21 *Staurograptus dichotomus* Emmons. Tremadoc; New York; figure from Bulman (1955).

Figure 22 *Glyptograptus dentatus* (Brongniart). x4; Arenig; Sweden; figure from Bulman (1955).

Figure 23 *Tetragraptus fruticosus* (Hall). Arenig; Quebec; figure from Bulman (1955).

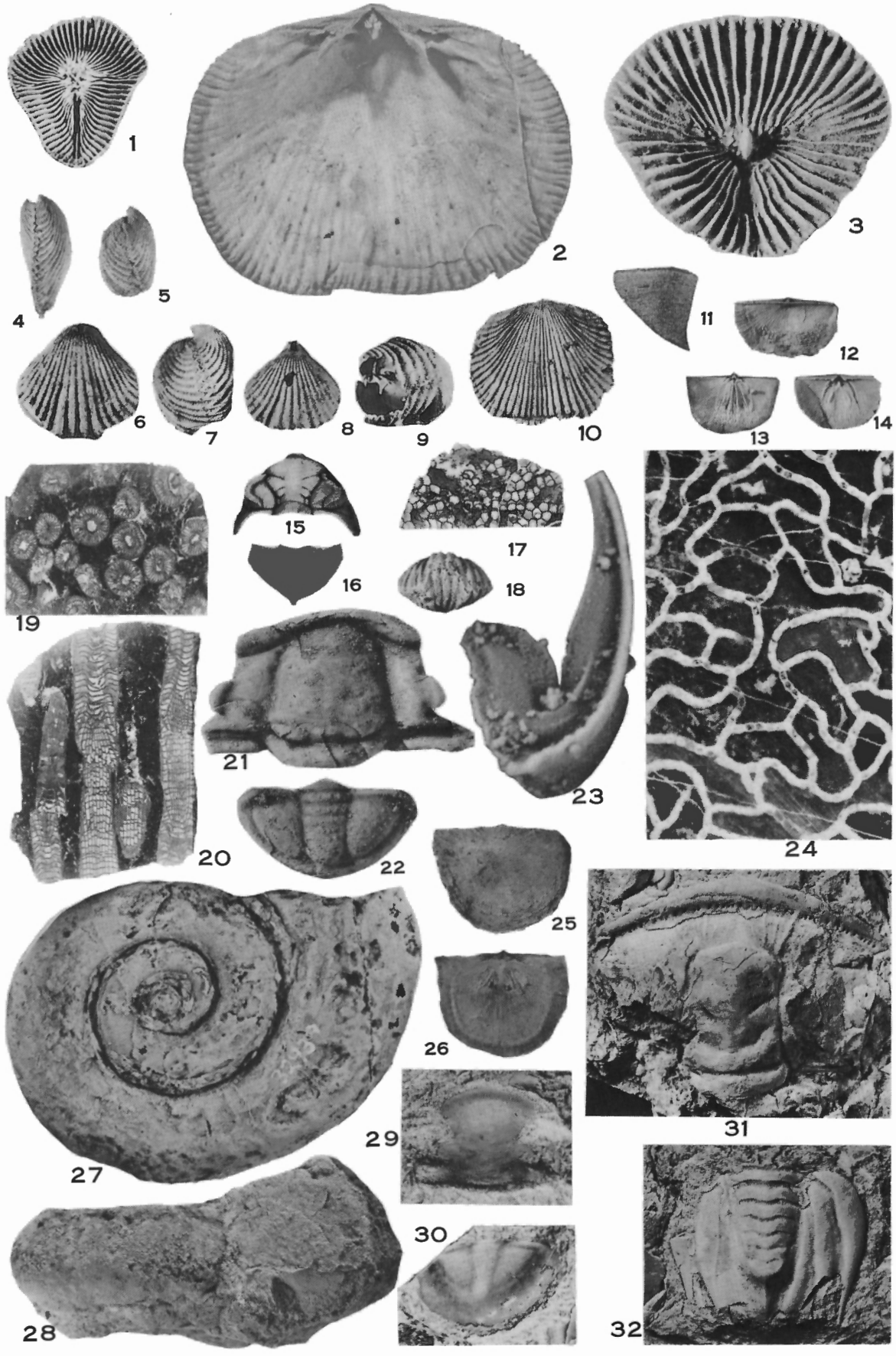
Figure 24 *Didymograptus extensus* (Hall). Arenig; Quebec; figure from Bulman (1955).

Figure 25 *Isograptus caduceus maximo-divergens* Harris. x2; Arenig; British Columbia; figure from Jackson (1964).

Figure 26 *Bryograptus* sp. Tremadoc; Sweden; figure from Bulman (1955).

Figure 27 *Dictyonema flabelliforme* (Eichwald). Tremadoc; England; figure from Bulman (1955), dashed lines show approximate zones of branching.

Figure 28 *Clonograptus rigidus* (Hall). x½; Arenig; Quebec; figure from Bulman (1955).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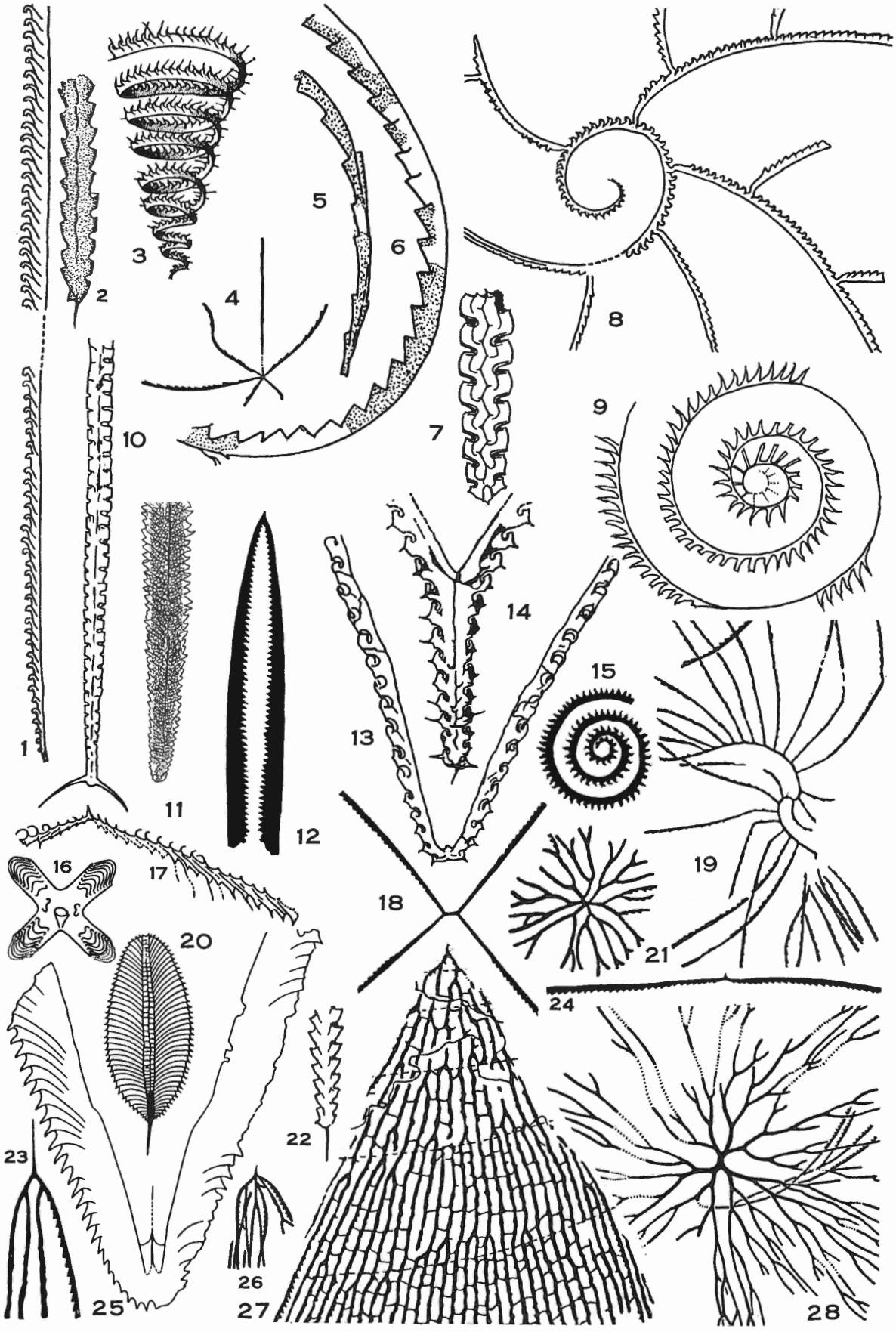
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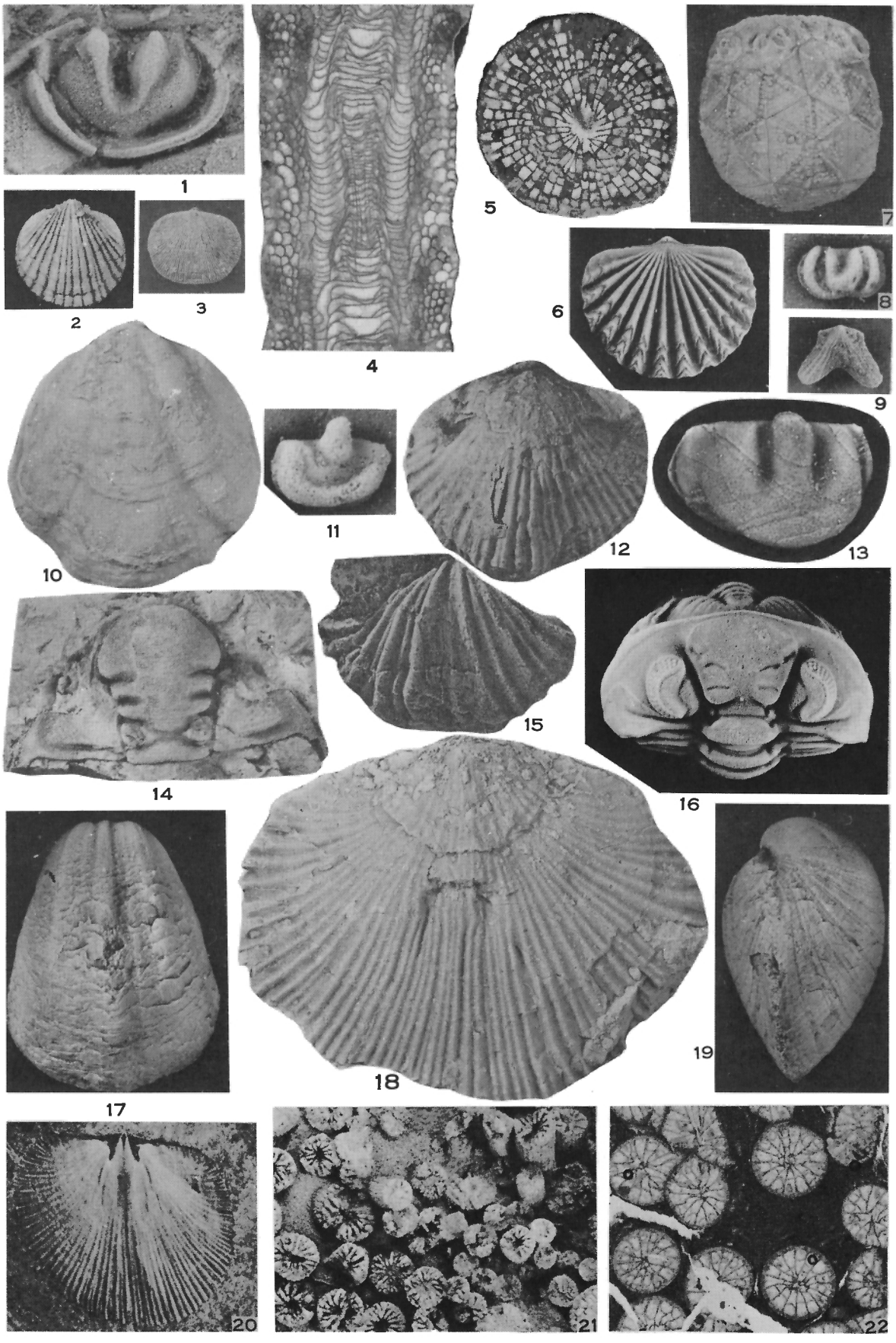
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PLATE VII. Silurian Fossils of Eastern Canada

Figure 1 *Zygobolba decora* (Billings). x16; Niagaran; Quebec; 2547.

Figure 2 *'Coelospira' planoconvexa* (Hall). x1; Alexandrian; Ontario; 17959.

Figure 3 *Mendacella* sp. x1; Alexandrian; Ontario; 17955.

Figures 4, 5 *Entelophyllum pennanti* (Billings). x4; Niagaran; Quebec; 3040m.

Figure 6 *Eocoelia hemispherica* (Sowerby). x3; Niagaran; Quebec; 17090.

Figure 7 *Caryocrinites ornatus* Say. x1; Tonawandan; Ontario; 2656d.

Figure 8 *Dizygopleura symmetrica* (Hall). x15; Tonawandan; Ontario; 15196b.

Figure 9 *Dicoelosia acutilobus* (Ringueberg). x5; Tonawandan; Ontario; 17965.

Figure 10 *Pentameroides* sp. x1; Ontarian; Ontario; 17973.

Figure 11 *Paraechmina spinosa* (Hall). x15; Tonawandan; Ontario; 15192a.

Figure 12 *Costistricklandia gaspéensis* Billings. x1; (after Northrop, 1939); Niagaran; Quebec.

Figure 13 *Kloedenia wilckensiana* (Jones). x15; Downtonian; Nova Scotia; 14514.

Figure 14 *Cheirurus niagarensis* (Hall). x1; Lockportian; Ontario; 17969.

Figure 15 *Janius* sp. x2; (after Boucot, 1963); Niagaran; New Brunswick; 15152.

Figure 16 *Eophacops orestes* (Billings). x3; Niagaran; Quebec; 17091.

Figures 17, 19 *Virgiana barrandei* (Billings). x1; Niagaran; Quebec; 2372, b.

Figure 18 *Costistricklandia canadensis* Billings. x1; Ontarian; Ontario; 17972.

Figure 20 *Salopina lunata* (Sowerby). x2; (after Boucot, *et al.*, 1960); Niagaran; New Brunswick; 14962.

Figure 21 *Palaeophyllum williamsi* Chadwick. x4; Alexandrian; Ontario; 4508.

Figure 22 *Palaeophyllum multicaule* (Hall). x4; Lockportian; Ontario; 17092.

PLATE VIII. Silurian Fossils of Western Canada

(All figures natural size unless otherwise stated)

Figures 1, 4 *Conchidium alaskense* Kirk and Amsden. Ludlow; southeast Alaska; USNM 116563 and 116566; figures from Kirk and Amsden (1952).

Figures 2, 3 *Atrypella scheii* (Holtedahl)? Ludlow, southeast Alaska; USNM 116578 and 116573; figures from Kirk and Amsden (1952).

Figures 5, 6 *Eurypterus fischeri* Eichwald. Upper Wenlock or lower Ludlow; Cornwallis Island; 13997 and 14003.

Figures 7, 8 *Atrypa gabrielsi* Norford. Upper Llandovery; British Columbia; 16918a and 16918b.

Figure 9 *Scutellum borealis* (Poulsen). Upper Llandovery; Yukon Territory; 15411.

Figures 10, 13, 17 *Glassia variabilis* Whiteaves. x2; upper Llandovery; British Columbia; 16900.

Figure 11 *Favosites discoideus* (Roemer). Upper Llandovery; British Columbia; 15822.

Figure 12 *Favosites favosus* (Goldfuss). Upper Llandovery; British Columbia; 15850.

Figure 14 *Encrinurus* cf. *E. princeps* Poulsen. Upper Llandovery; Yukon Territory; 16903.

Figure 15 *Halysites occidens* Norford. Upper Llandovery; British Columbia; 14484.

Figure 16 *Dicoelosia* cf. *D. biloba* (Linnaeus). x2; upper Llandovery; British Columbia; 15788.

Figure 18 *Cystihalysites magnitubus* (Buehler). Longitudinal section; upper Llandovery; British Columbia; 15757c.

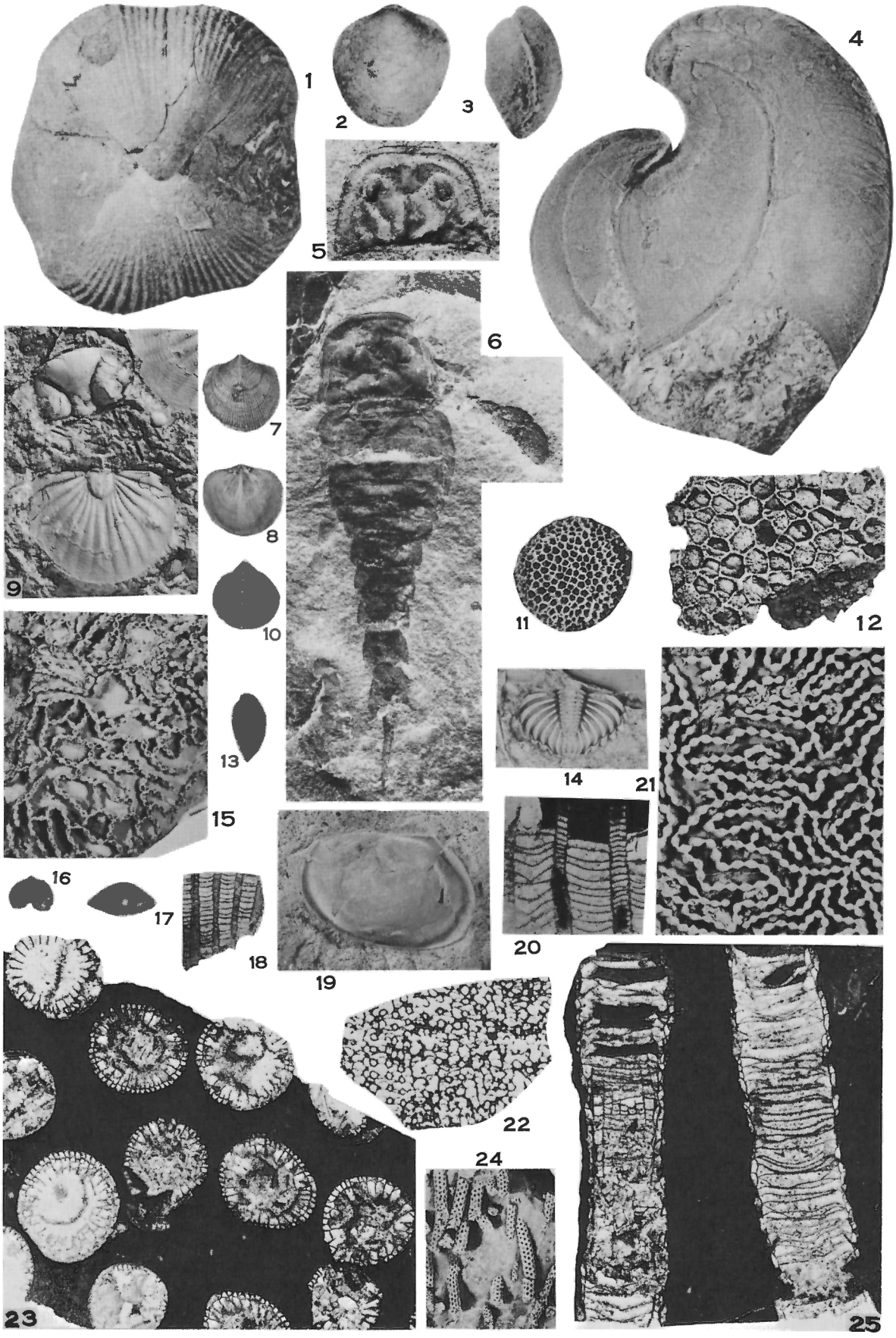
Figure 19 *Dihogmochilina latimarginata* (Jones). Llandovery; Manitoba; 17089.

Figures 20, 21 *Halysites nitida* Lambe. x4 (longitudinal section) and x1; upper Llandovery; British Columbia; 15825a and 15825.

Figure 22 *Multisolenia tortuosa* Fritz. Longitudinal section x3; Llandovery; Manitoba; 10485.

Figures 23, 25 *Columnaria columbia* Norford. Transverse and longitudinal sections x2; Llandovery; British Columbia; 16524a and 16524b.

Figure 24 *Coenites rectilineatus* (Simpson). Upper Llandovery, British Columbia; 15821.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DEVONIAN FAUNAS

by D. J. McLaren, A. W. Norris, and L. M. Cumming

In Canada the Devonian was characterized by increasing marine transgression throughout the period. By the Upper Devonian most of the land surface was submerged under shallow seas carrying rich and varied faunas. There is a marked difference between the faunas found in the rocks in western and arctic Canada and those of eastern Canada. Various subprovinces also exist within these two divisions that are dependent largely on rock facies.

Pelagic fossils, especially ammonoids, occur only at widely scattered localities and allow rare correlation with the European standard stages. Conodonts hold out hope of accurate long-range correlation, but their study is still at an early stage. Graptolites from Lower Devonian rocks are now known from northern, western, and eastern Canada, but are largely unstudied. Shelly benthonic faunas, typically corals, brachiopods, and trilobites were widespread in shallow seas with carbonate banks and reefs, especially during the Middle and early Upper Devonian. Zonal schemes have been worked out in some western areas, and in some instances allow intercontinental correlation. In Ontario rocks of similar facies carry very different coral and brachiopod faunas and correlation is not easy with the west or arctic although some overlapping of faunas occurs in Manitoba. Sandstone facies in the Appalachian and parts of arctic Canada again carry different faunas, both marine and non-marine.

Table XI-4 lists some common fossils that are restricted in time and are therefore of value in correlation. They are shown in relation to the standard European stages and to American ammonoid zones, although precise correlation has not always been established. Some typical Devonian fossils are illustrated on PLATES IX to XI.

Western Canada

Lower Devonian

Lower Devonian fossils have been recovered from only a few localities in western Canada, mainly because beds of this age, where present, are generally dolomitized or evaporitic. In the southern plains of Alberta and Saskatchewan only a few ostracods dated as either Lower or Middle Devonian have been found in the lower part of the evaporitic Elk Point Group. From the Delorme Formation on Cathedral Mountain, southern Mackenzie Mountains, brachiopods have been identified by G. A. Cooper (Lenz and Jackson, 1964) including *Gypidula pelagica, Atrypa* aff. *A. reticularis nieczlawiensis, Cyrtina praecedens, Camarotoechia* cf. *C. squamifera,* and others. This fauna is tentatively dated as early Lower Devonian (Gedinnian).

In northern Yukon *Monograptus yukonensis* is present in the upper part of the Road River Formation commonly associated with *M.* cf. *M. praehercynicus* and *Nowakia* cf. *N. acuaria.* The *M. yukonensis* fauna ranges in age from late Siegenian to early Emsian. A shelly fauna including brachiopods and corals occurs at a few places in limestone beds interbedded with the graptolitic shales. A unit of interbedded shale and carbonates unconformably overlying the Road River Formation in the west-central part of Yukon Territory contains a rich shelly fauna characterized by *Dechenellurus* sp. B that is tentatively dated as late Lower Devonian (Emsian) on the basis of trilobites and goniatites.

Middle Devonian

The ostracod *Moelleritia canadensis* occurs in the upper half of a unit of dolomitic carbonates generally otherwise barren of fossils. The unit is roughly equivalent to the lower four fifths of the Bear Rock Formation of the central Mackenzie River area. *M. canadensis* is widely distributed in northern Yukon Territory and occurs also on Victoria Island, District of Franklin. It indicates an Eifelian age.

Distinctive echinoderm ossicles with double and cross-like axial canals occur in a variety of facies in beds immediately above the *Moelleritia canadensis* zone but also overlapping this zone, and below the Hume Formation or equivalent beds. These ossicles occur in the Manetoe Formation and equivalent beds near Virginia Falls, and in the first canyon of South Nahanni River, in the lower part of a limestone in northeastern British Columbia, and elsewhere. They are commonly associated with a wide variety of fossils and are taken to indicate a mid-Eifelian to late Emsian age.

Two widely distributed associations of probable early Givetian age, characterized by *Schuchertella adoceta* and *Carinatina dysmorphostrota,* occur in the lower and upper beds respectively of the Hume Formation of the central Mackenzie region. A form resembling *Carinatina dysmorphostrota* occurs in dolomite beds about the middle of the evaporitic Chinchaga Formation west of Slave River suggesting a correlation with the upper Hume Formation of the central Mackenzie region.

Stringocephalus, the guide fossil for the Givetian, is known from the Burgess Mountain area of Yukon where it occurs throughout a thousand feet of section, and from a number of localities in the central Mackenzie region. In the latter area it occurs mainly in the Kee Scarp Formation, but has been reported also in the uppermost beds of the Hume Formation. Species of *Stringocephalus* occur in the Winnipegosis and Dawson Bay Formations of Manitoba, in the Methy Formation of northeastern Alberta, in the upper two thirds of the Pine Point and in the lower part of the Sulphur Point and Presqu'ile Formations of the Great Slave Lake area, in the McDame Group of northern British Columbia, and elsewhere.

The uncoiled gastropod *Mastigospira* is in places associated with *Stringocephalus* and occurs in the Dawson Bay Formation of Manitoba with other bizarre molluscs, in the Methy (lower part), La Butte and Little Buffalo Formations of northeastern Alberta. *Desquamatia arctica* and closely related forms occur abundantly in the Elm Point Formation of Manitoba, in the lower part of the Methy Formation of Alberta, in the lower part of the Pine Point Formation of the Great Slave Lake area, and in the Hume and Hare Indian Formations of the Mackenzie region.

Characteristic brachiopods of the upper two thirds of the Pine Point Formation include *"Chonetes" aurora, Devonoproductus* sp. E, *Emanuella meristoides, Warrenella* cf. *W. franklini, Warrenella kirki, Leiorhynchus awokanak, Leiorhynchus castanea,* and *Hadrorhynchia sandersoni.* Some of these forms are present also in the Horn River Formation on the northwest side of the Great Slave Lake, in the lower part of the Besa River Formation of northeastern British Columbia, and in the Hare Indian Formation of central Mackenzie region. *Leiorhynchus castanea* generally occurs in the basal beds of the Hare Indian Formation, but at some localities, as at Carcajou Ridge, it may range as high as middle Kee Scarp.

Post-*Stringocephalus* Middle Devonian megafaunas are with few exceptions not abundant in western Canada.

TABLE XI-4 *Distribution and succession of Devonian faunas*

SERIES	EUROPEAN STAGES	NORTH AMERICAN AMMONOID ZONES (House, 1962, p. 265)	YUKON TERRITORY AND CENTRAL MACKENZIE BASIN	ROCKY MOUNTAINS AND PLAINS	ARCTIC ISLANDS	ONTARIO	APPALACHIAN MOUNTAINS (Invertebrates)	APPALACHIAN MOUNTAINS (Vertebrates and plants)
UPPER DEVONIAN	FAMENNIAN	? *Cyrtoclymenia strigata* ? *Falciclymenia bowsheri* *Platyclymenia americana* *Sporadoceras milleri* (?) *Sporadoceras pompeckji* *Cheiloceras amblylobum*	*Basilicorhynchus* *Cyrtospirifer*	*Gastrodetoechia utahenis* *Evanescirostrum seversoni* *Eoparaphorhynchus walcotti*	*Acanthatia* *Basilicorhynchus* and *Cyrtospirifer*			
	FRASNIAN	*Manticoceras cataphractum* *Manticoceras rhynchostoma* *Probeloceras lutheri* *Manticoceras styliophilum* *Ponticeras perlatum* *Pharciceras amplexum*	*Theodossia scopulorum* *Calvinaria albertensis* *Ponticeras*	*Theodossia scopulorum* *Adolphia* *Calvinaria albertensis* *Calvinaria insculpta* *Calvinaria athabascensis* *Allanaria allani* *Ladogioides kakwaensis*	*Cyrtospirifer* *Eleutherokomma* *Allanaria* *Devonoproductus*	Conodonts, and *Tasmanites huronensis*		*Archaeopteris gaspensis,* and *Bothriolepis canadensis*
MIDDLE DEVONIAN	GIVETIAN	*Wedekindella brilonense* or *Tornoceras uniangulare* *Sobolewia virginiana* (?) *Cabrieroceras plebeiforme*	*Emanuella* sp. F *Stringocephalus* *Leiorhynchus castanea* *Carinatina dysmorphostrota*	*Eleutherokomma implana* *Emanuella* sp. F *Stringocephalus* *Leiorhynchus castanea* *Carinatina dysmorphostrota*	*"Camarotoechia princeps"* *Emanuella* *Leiorhynchus castanea*	*Spinocyrtia carinata* *Leiorhynchus laura,* *Mucrospirifer thetfordensis* *Fimbrispirifer divaricatus,* *Gallipleura nobilis* *Mucrospirifer arkonensis,* *Leiorhynchus kelloggi*		Spores in Gaspé and Cobequid Mountains areas.
	EIFELIAN	*Foordites buttsi*	*Schuchertella nevadensis* Echinoderm ossicles with double and cross-like axial canals *Moellerita canadensis*	Echinoderm ossicles with double and cross-like axial canals	*Cortezorthis,* *"Reticularia curvata"* and *Indospirifer*	*Martiniopsis maia* *Rhipidomella vanuxemi* *Paraspirifer acuminatus* *Prosserella lucasi*		
LOWER DEVONIAN	EMSIAN		*Dechenellurus* sp. B		*Cortezorthis* and *Dalejina*	*Amphigenia* cf. *A. curta* *Aemulophyllum exiguum*	*Amphigenia* *Etymothyris-Eodevonaria*	*Prototaxites,* and *Drepanophycus spinaeformis*
	SIEGENIAN		? *Monograptus yukonensis*		*Monograptus hercynicus* and *M.* n. sp. A	*Costispirifer arenosus* *Hipparionyx proximus*	*Rensselaeria* *Costispirifer arenosus,* *Murchisonia,* and *Meristella* *Globithyris*	
	GEDINNIAN		?	*Gypidula* cf. *pelagica*	*Monograptus uniformis* and *M. angusitidens*		*Tubulibairdia,* and *Zygobeyrichia Podolella,* and *Howellella*	*Traquairaspis*

GSC

Atrypa cf. *A. independensis, Emanuella* sp. C, and *Emanuella* sp. F are the main forms other than stromatoporoids present in the Slave Point Formation of the Great Slave Lake area, northeastern Alberta, and the Rocky Mountains of northeastern British Columbia. A unique coral and brachiopod fauna of late Givetian age occurs in the Horn Plateau Formation, a local reef development northwest of Great Slave Lake. Characteristic brachiopods include *Pholidostrophia?* sp., *Longispina whittakeri, Hypothyridina cameroni, Atrypa nasuta, Spinatrypa hornensis, Eleutherokomma implana,* and *Trematospira* sp. Important corals in this fauna include *Cylindrophyllum gruensis, Heliophyllum borealis,* and *Sociophyllum redactum.* Two of the above brachiopods, *Hypothyridina cameroni* and *Atrypa nasuta,* occur in beds elsewhere closely associated with *Stringocephalus* sp.

Upper Devonian

The Frasnian rocks of western Canada may be divided, on the basis of brachiopods, into a series of zones that are remarkably widespread and consistent. The lowest beds of the Waterways Formation contain a fauna markedly different from that of the underlying Middle Devonian, including productellids, *Ladogioides pax, Leiorhynchus russelli, Cyrtina billingsi,* and *Atrypa.* A similar fauna marks the base of transgressive Upper Devonian beds at scattered points in the Alberta Rocky Mountains. Succeeding faunas include the coral *Tabulophyllum athabascense* and the brachiopod genera *Devonoproductus, Allanaria,* and *Eleutherokomma.*

The middle Frasnian is marked by the incoming of abundant rugose corals, *Cyrtospirifer,* and, after a brief overlap, by the disappearance of *Eleutherokomma.* In the argillaceous sequence of the Hay River region and Alberta Rockies, the rhynchonellid zones, in ascending order, *Calvinaria athabascensis, C. insculpta,* and *C. albertensis* are the temporal equivalents of reef and massive carbonate bank developments with stromatoporoids, tabulate, and rugose corals, including species of *Hexagonaria, Phillipsastraea,* and *Thamnophyllum.* The latest Frasnian carbonates contain different species of corals and the brachiopod genera *Hypothyridina* and *Theodossia.*

Regression and erosion occurred at the end of the Frasnian which is marked by the disappearance of strophedontids, pentamerids, atrypids, and a considerable reduction of stromatoporoids. The lower and middle Famennian beds are typified by distinctive productellids, rhynchonellids, and *Cyrtospirifer* species. Typical rhynchonellids, in ascending order, include *Eoparaphorhynchus walcotti, Evanescirostrum seversoni,* and *Gastrodetoechia utahensis.* No late Famennian shelly faunas are known.

Arctic Islands

Many Lower Devonian rocks in the arctic islands have previously been considered Silurian on the basis of graptolites. Revision of the correlation between graptolitic and shelly fauna sequences in Europe suggests that several Devonian graptolite zones are present in the islands. On Bathurst Island, the upper part of the Cape Phillips Formation contains *Monograptus ultimus* succeeded by *M. uniformis* generally accepted as marking the base of the Gedinnian. The overlying Bathurst Island Formation contains *M. hercynicus* and a characteristic new species of *Monograptus* now believed to be of Siegenian age.

Lower Devonian shelly faunas which commonly overlie the graptolitic rocks are not widespread, and occur in highly variable successions not easy to correlate in detail between islands. On Bathurst Island, the Stuart Bay Formation contains brachiopods that suggest a Siegenian to Emsian age. Other faunas elsewhere have not yet been adequately studied. A large and varied fish fauna occurs in Lower Devonian rocks on Somerset and Prince of Wales Islands.

The Middle Devonian rocks are richly fossiliferous and the Blue Fiord Formation predominantly of Eifelian age contains abundant rugose corals, the brachiopods *Cortezorthis, "Reticularia* cf. *R. curvata", Indospirifer,* and a large variety of trilobites in which dechenellids predominate. In some districts the Blue Fiord may extend downwards into the Emsian Stage. The succeeding Bird Fiord Formation, of Givetian age, has a relatively restricted fauna; corals are rare but may be abundant locally; brachiopods include *"Camarotoechia princeps", Emanuella,* and abundant *Atrypa;* trilobites are present in large numbers. Brackish-water shales and sandstones upwards contain rich pelecypod–gastropod faunas with few brachiopods and fewer trilobites. On Melville Island, in a shale–sandstone sequence, the common Mackenzie Valley Givetian brachiopod *Leiorhynchus castanea* is found. The absence of the Givetian genus *Stringocephalus,* presumably due to lack of carbonate rocks, is noteworthy.

The Upper Devonian rocks of the eastern arctic islands are predominantly non-marine clastics with plant fossils, spores, and fish fragments. Scattered localities on Banks, Melville, and Prince Patrick Islands have yielded marine faunas including basal Frasnian rocks with *Eleutherokomma, Allanaria,* and *Spinatrypa,* succeeded by beds with *Cyrtospirifer.* Coral reefs developed in the early middle Frasnian on northeastern Banks Island. A widespread brachiopod fauna occurs in the early Famennian of Melville and Bathurst Islands and includes *Acanthatia, Basilicorhynchus,* and *Ptychomaletoechia.*

Eastern Canada

Ontario

The Lower Devonian Oriskany Sandstone, with the brachiopods *Costispirifer arenosus* and *Hipparionyx,* occurs sporadically in Niagara Peninsula. It is succeeded by a thin unit below the Onondaga, probably of Lower Devonian age, with the corals *Acrophyllum oneidense, Aemulophyllum exiguum,* and *"Cystiphyllum",* and the

brachiopods *Amphigenia* cf. *A. curta, Centronella,* and *Eodevonaria.*

The overlying beds from Niagara to the Woodstock area, equivalent to the Onondaga of New York, include the Detroit River Group, with the brachiopods *Prosserella lucasi* and *Paraspirifer acuminatus.* Corals, abundant throughout this interval, include *Acinophyllum, Bethanyphyllum, Blothrophyllum, Cystiphylloides, Eridophyllum, Siphonophrentis,* and *Synaptophyllum.*

The Delaware Limestone follows disconformably and contains the brachiopods *"Martiniopsis" maia* and *Rhipidonella vanuxemi,* and the pelecypod *Paracyclas proava.* The succeeding calcareous shales of the Hamilton Group are rich in solitary corals and brachiopods, including *Heliophyllum, Cystiphylloides, Mucrospirifer, Fimbrispirifer divaricatus, Leiorhynchus laura,* and *Spinocyrtia carinata,* and the trilobites *Greenops* and *Phacops.* The Kettle Point Shale rests with disconformity on the underlying beds and represents much of Upper Devonian time. Concretions within the shale contain plant spores, radiolarians, conodonts, and fish bones.

At a few places in the Hudson Bay Lowland, clastic deposits of the Sextant Formation immediately overlie Precambrian rocks and contain plant remains, including *Drepanophycus spinaeformis* Geoppert, dated as Lower Devonian. The overlying Abitibi River Formation, mainly limestone, contains numerous corals, stromatoporoids, bryozoans, brachiopods, pelecypods, cephalopods, and trilobites of Lower and Middle Devonian ages. The Williams Island Formation consisting of limestones, shales, sandstones, and gypsum contains Hamilton fossils in the lower part and *Hypothyridina* sp. in the uppermost bed suggesting correlation with the Tully of New York. Dark shales of the Long Rapids Formation are the youngest Devonian rocks in the Hudson Bay Lowland containing plant spores similar to those in the Kettle Point Formation of southwestern Ontario.

Appalachian Mountains

Lower Devonian fossils are widely distributed in Newfoundland, Nova Scotia, New Brunswick, and Quebec, and several distinct facies occur.

A Gedinnian marine faunule characterized by the brachiopods *Howellella* and *Podolella* occurs in the Nictaux–Torbrook region of Nova Scotia. This nonlimestone facies is associated with quartzites and sedimentary iron-formation and is similar to the European Rhenish facies. An Old Red Sandstone facies, terrestrial in origin and containing *Traquairaspis* occurs in the Arisaig region, 175 miles to the east.

In central Gaspé Peninsula of Quebec, a marine marginal-basin facies characterized by the zonal brachiopods *Meristella, Rensselaeria, Etymothyris,* and *Amphigenia* represents a typical benthonic Appalachian limestone facies of Siegenian and Emsian ages. This abundant fauna occurs in the Gaspé Limestone and lower calcareous part of the overlying Gaspé Sandstone. Ostracods (e.g., *Tubulibairdia* and *Zygobeyrichia*) and corals characterize the limy marine Dalhousie shale of New Brunswick.

Also in Gaspé, a transitional facies, characterized by *Globithyris* and *Rhenorensselaeria,* is laterally associated with a plant-eurypterid-fish-bearing sandstone with *Psilophyton, Drepanophycus,* and *Prototaxites.* The change in Lower Devonian sedimentation from limestone to sandstone deposition in Gaspé reflects the Acadian Orogeny the main uplift lying south of the subsiding marginal-basin in Gaspé. Because of widespread uplift of the eastern Appalachians during orogeny, marine faunas of Middle and Upper Devonian ages are unknown in the Canadian Appalachians. Middle Devonian plant material is probably present in the upper part of the Gaspé Sandstone and in other terrestrial deposits in Cobequid Mountains. Upper Devonian deposits with *Bothriolepis* and *Archaeopteris* are best known at Escuminac, Quebec. Similar post-orogenic sandstones and shales of terrestrial origin occur at Terranceville, Newfoundland, and in Charlotte county, New Brunswick.

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PLATE IX. Devonian Fossils, Yukon Territory and Northern Interior Plains

Figures 1, 2 *Monograptus yukonensis* Jackson and Lenz. x3; upper Road River Formation; Siegenian; Yukon Territory (from Jackson and Lenz, 1963).

Figure 3 *Moelleritia canadensis* Copeland. Eifelian; Yukon Territory; 20690.

Figures 4, 5 *Gasterocoma? bicaula* Johnson and Lane. x10; Eifelian; Yukon Territory; 20691,a.

Figures 6, 7 *Carinatina dysmorphostrota* Crickmay. Upper Hume Formation; Givetian; Yukon Territory; 16695.

Figure 8 *Spinatrypa coriacea* Crickmay. Upper Hume Formation; Givetian; District of Mackenzie; 16694.

Figures 9, 10 *Desquamatia arctica* (Warren). Upper Hume Formation; Givetian; District of Mackenzie; 16691.

Figures 11, 12 *Mesophyllum* (*Digonophyllum*) *rectum* (Meek). x2; upper Hume Formation; Givetian; District of Mackenzie; 16529.

Figures 13, 14 *Utaratuia laevigata* Crickmay. x2; Nahanni Formation; Givetian; District of Mackenzie; 16541.

Figures 15, 16 *Grypophyllum* cf. *G. wedekindi* Middleton. x2; Kee Scarp Formation; Givetian; District of Mackenzie; 16532.

Figures 17, 18 *Sociophyllum glomerulatum* (Crickmay). Middle Devonian Limestone Unit; Givetian; Yukon Territory; 16551.

Figures 19, 20 *Xystriphyllum hyperbolicum* Crickmay. x2; McDame Group; Givetian; British Columbia; 16539.

Figures 21, 22, 23 *Leiorhynchus castanea* (Meek). Upper Hume Formation; Givetian; District of Mackenzie; 15361.

Figures 24, 25 *Hadrorhynchia sandersoni* (Warren). Kee Scarp Formation; Givetian; District of Mackenzie; 15331.

Figure 26 *Stringocephalus* sp. Kee Scarp Formation; Givetian; District of Mackenzie, 20683.

Figure 27 *Cyrtospirifer thalattodoxa* Crickmay. Hay River Formation; Frasnian; District of Mackenzie; 16563.

Figures 28, 29 *Basilicorhynchus basilicus* (Crickmay). Famennian; District of Mackenzie; Crickmay holotype, 1952.

(Plates IX–XI—all figures natural size unless otherwise stated)

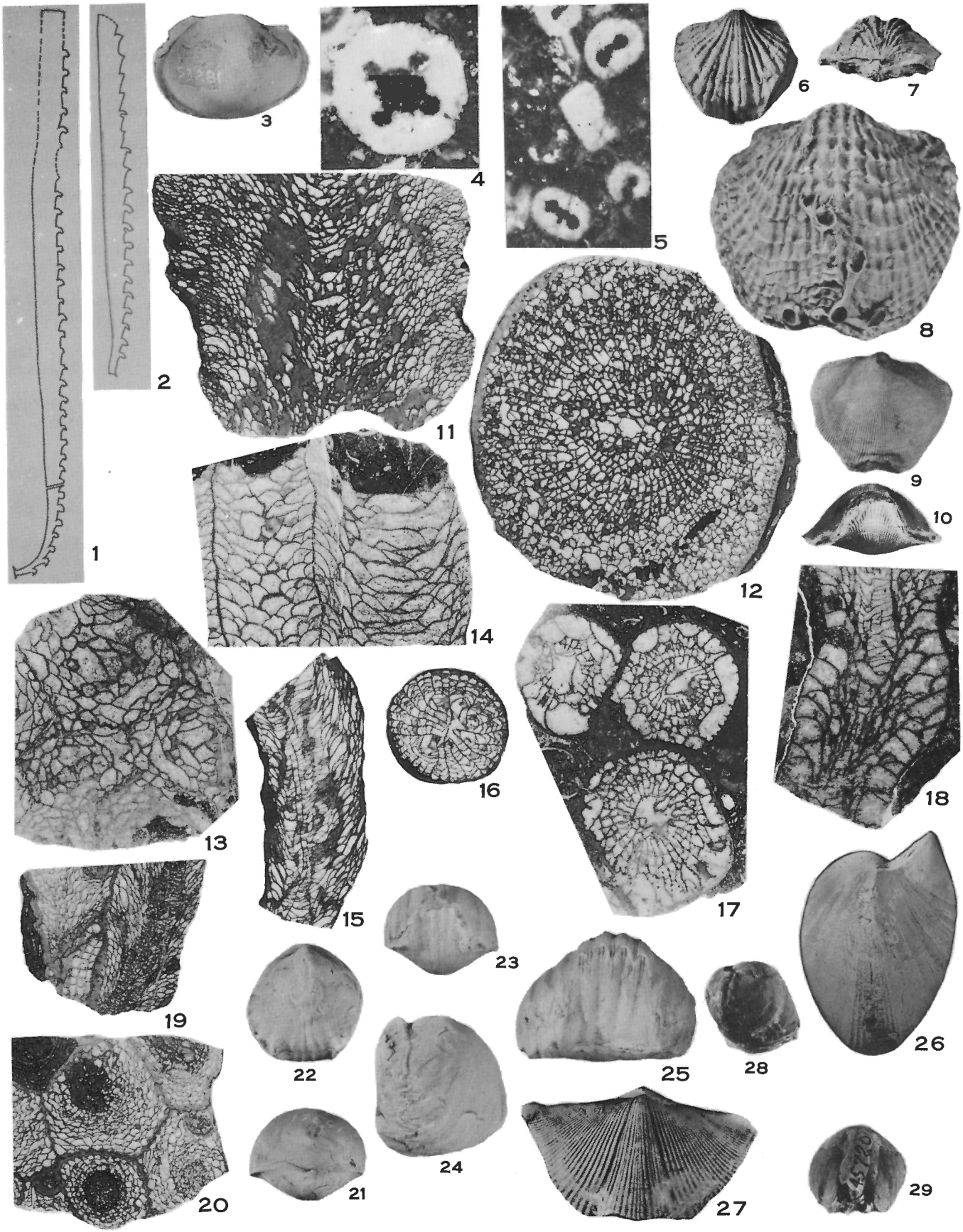
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PLATE X. Devonian Fossils, Rocky Mountains, Southern Interior Plains, and Ontario

ROCKY MOUNTAINS AND INTERIOR PLAINS

Figures 1, 2 *Gypidula pelagica* (Barrande); Delorme Formation; Lower Devonian; Cathedral Mountain, District of Mackenzie; 16688.

Figure 3 *Atrypa* cf. *A. reticularis nieczlawiensis* Kozlowski; same horizon and locality as above; 16689.

Figures 4, 5, 6 *Emanuella meristoides* (Meek); Pine Point Formation; Givetian; Great Slave Lake, District of Mackenzie; 16701.

Figures 7, 8 *Dendrostella trigemme* (Quenstedt); Nahanni Formation; Givetian; Nahanni Range, District of Mackenzie; GSC loc. 32125.

Figure 9 *Mastigospira alata* (Whiteaves); Dawson Bay Formation; Givetian; Lake Winnipegosis, Manitoba; 4099a.

Figure 10 *Emanuella* sp. F; x2; Sulphur Point Formation; Givetian; Great Slave Lake, District of Mackenzie; 16707.

Figures 11, 12 *Eleutherokomma implana* Norris; Horn Plateau Formation; Givetian; Horn Plateau, District of Mackenzie; 16106.

Figures 13, 14 *Ladogioides kakwaensis* (McLaren); Flume Formation; Frasnian; British Columbia; 13819.

Figures 15, 16 *Calvinaria variabilis athabascensis* (Kindle); Maligne Formation; Frasnian; Jasper National Park; 11232.

Figure 17 *Polygnathus asymmetrica ovalis* Ziegler and Klapper; x45; Moberly? Member of Waterways Formation; Frasnian; Alberta; 20693.

Figures 18, 19 *Allanaria allani* (Warren); Moberly Member of Waterways Formation; Frasnian; Alberta; 16714.

Figure 20 *Eleutherokomma* cf. *E. hamiltoni* Crickmay; Moberly Member of Waterways Formation; Frasnian; Alberta; 16715.

Figure 21 *Ancyrodella gigas* Youngquist; x45; Moberly Member of Waterways Formation; Frasnian; Alberta; 20694.

Figures 22, 23, 24 *Calvinaria variabilis insculpta* (McLaren); Cairn Formation; Frasnian; Alberta; 14939.

Figures 25, 26 *Calvinaria albertensis albertensis* (Warren); Mount Hawk Formation; Frasnian; Alberta; 11237.

Figures 27, 28 *Indospirifer orestes* (Hall and Whitfield); Mount Hawk Formation; Frasnian; Alberta; 13811.

Figures 29, 30 *Pterorrhiza proteus* (Smith); Hay River Formation; Frasnian; District of Mackenzie; 16540.

Figures 31, 32, 33 *Theodossia scopulorum* (Crickmay); Kakisa Formation; Frasnian; District of Mackenzie; 16562.

Figures 34, 35 *Tabulophyllum mcconnelli* (Whiteaves); Hay River Formation; Frasnian; District of Mackenzie; 16531.

Figures 36, 37 *Evanescirostrum seversoni* (McLaren); Palliser Formation; Famennian; Alberta; 10016.

ONTARIO

Figures 38, 39, 40 *Rensselaeria ovalis* Hall; Oriskany Formation; Lower Devonian; Ontario; 3366.

Figures 41, 42 "*Martinia*" *maia* (Billings); Onondaga Formation; Eifelian; Ontario; 3496b.

Figures 43, 44 *Aemulophyllum exiguum* (Billings); Bois Blanc Formation; Lower Devonian; Ontario; 3424.

Figures 45, 46, 47 *Leiorhynchus laura* (Billings); Hamilton Group; Givetian; Ontario; 3705h.

Figures 48, 49 *Macrospirifer thedfordensis* (Shimer and Grabau); Hamilton Group, Givetian; Ontario; 20689.

PLATE XI. Devonian Fossils, Arctic Islands and Appalachian Mountains

ARCTIC ISLANDS

Figure 1 *Spyroceras thoas* (Hall); x2; Eids Formation; Emsian or Eifelian; Ellesmere Island, GSC loc. 57730.

Figures 2, 3 *Indospirifer* sp.; Blue Fiord Formation; Eifelian; Ellesmere Island; 20688.

Figures 4, 5 *"Reticularia"* ex gr. *curvata* (Schlotheim); Blue Fiord Formation; Eifelian; Ellesmere Island; 20686.

Figures 6, 7 "*Eatonia medialis* var." Meyer 1913; Blue Fiord Formation; Eifelian; Ellesmere Island; 20687.

Figures 8, 9, 10 *Cupularostrum?* sp; Bird Fiord Formation; Givetian; Bathurst Island; 20685.

Figures 11, 12 *Gypidula recurrens* (Meyer); Blue Fiord Formation; Eifelian; Ellesmere Island; 20684.

APPALACHIAN MOUNTAINS

Figure 13 *Octonaria foordi* Copeland; x30; upper Dalhousie beds; Gedinnian; New Brunswick; 14525.

Figure 14 *Octonaria* cf. *O. typicus* (Bassler); x30; upper Dalhousie beds; Gedinnian; New Brunswick; 14524.

Figure 15 *Zygobeyrichia dalhousiensis* Copeland; x15; upper Dalhousie beds; Gedinnian; New Brunswick; 14536.

Figure 16 "*Amphissites* (?)" *concentricus* (Ulrich and Bassler); x15; lower Dalhousie beds; Gedinnian; New Brunswick; 14530d.

Figure 17 *Dizygopleura chaleurensis* Copeland; x15; upper Dalhousie beds; Gedinnian; New Brunswick; 14531c.

Figure 18 *Kloedenia retifera* Ulrich and Bassler; x15; upper Dalhousie beds; Gedinnian; 14540f.

Figure 19 *Mesomphalus magnificus* Copeland; x15; upper Dalhousie beds; Gedinnian; New Brunswick; 14537k.

Figure 20 *Billingsastraea affinis* (Billings); Grande Grève Formation; Siegenian; Gaspé Bay, Quebec; 3270.

Figures 21, 22 "*Spirifer*" *arenosus* var. *unicus* Hall; Grande Grève Formation; Siegenian; Gaspé Bay; 3301.

Figure 23 *Strophonella planulata* (Hall); Gaspé Limestone Group; Gedinnian; Quebec; 3210.

Figure 24 *Rensselaeria ovoides* var. *gaspensis* Clarke and *Spirifer gaspensis* Billings; internal casts; Gaspé Sandstone Group; Siegenian; Gaspé Bay; 3307.

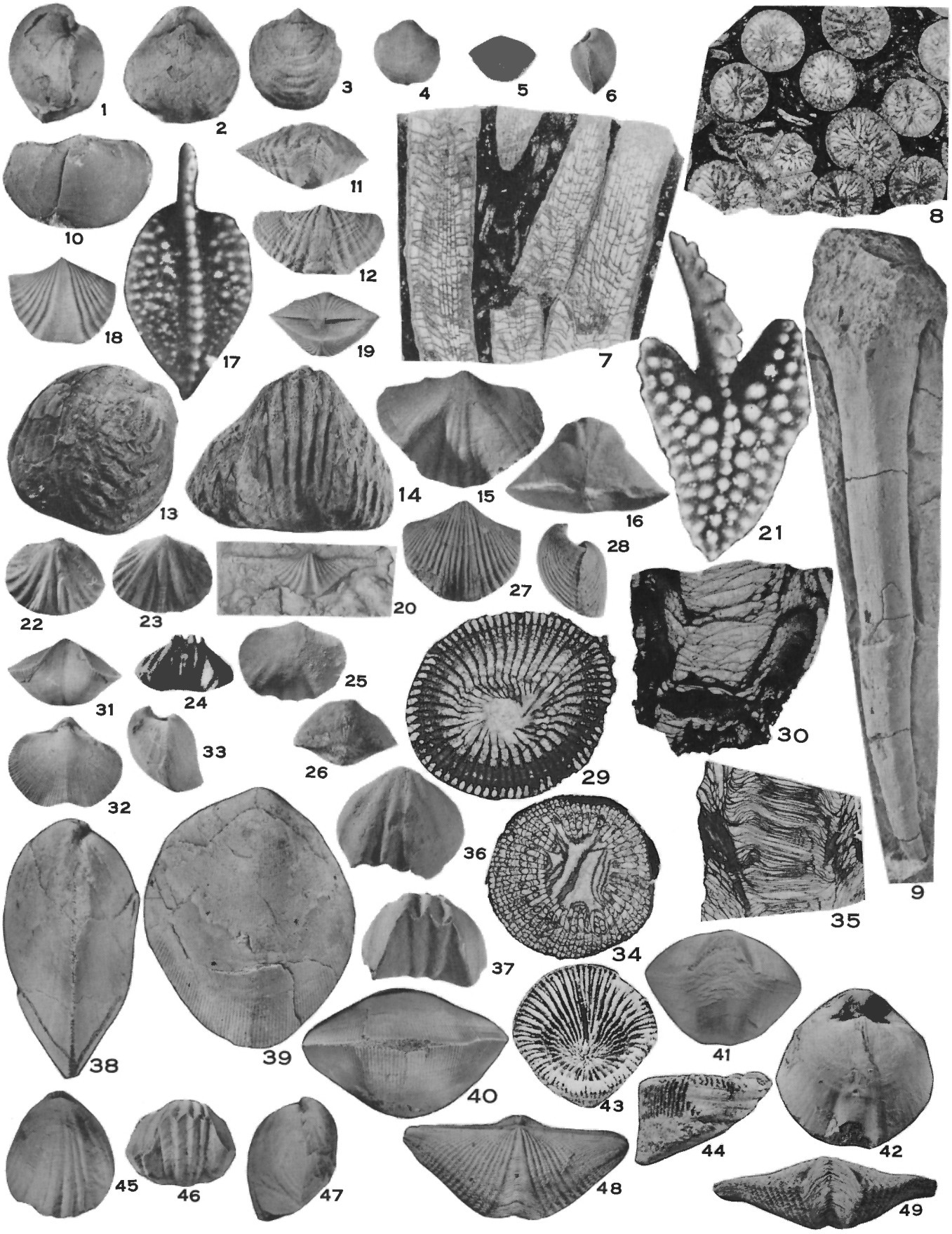
Figure 25 *Proetus phocion* Billings; Grande Grève Formation; Siegenian; Gaspé Bay; 3337.

Figures 26, 27, 28 *"Spirifer" cyclopterus* Hall; Grande Grève Formation; Siegenian; Quebec; 3303.

Figure 29 *Leptostrophia magnifica* var. *tullia* (Billings); Mount Joli Formation; Gedinnian; Quebec; 3285.

Figure 30 *Proschizophoria* sp.; Torbrook Formation; Gedinnian; Nova Scotia; 14761b.

Figure 31 *Etymothyris gaspensis* Boucot; York River Formation; Siegenian; Gaspé; 20692.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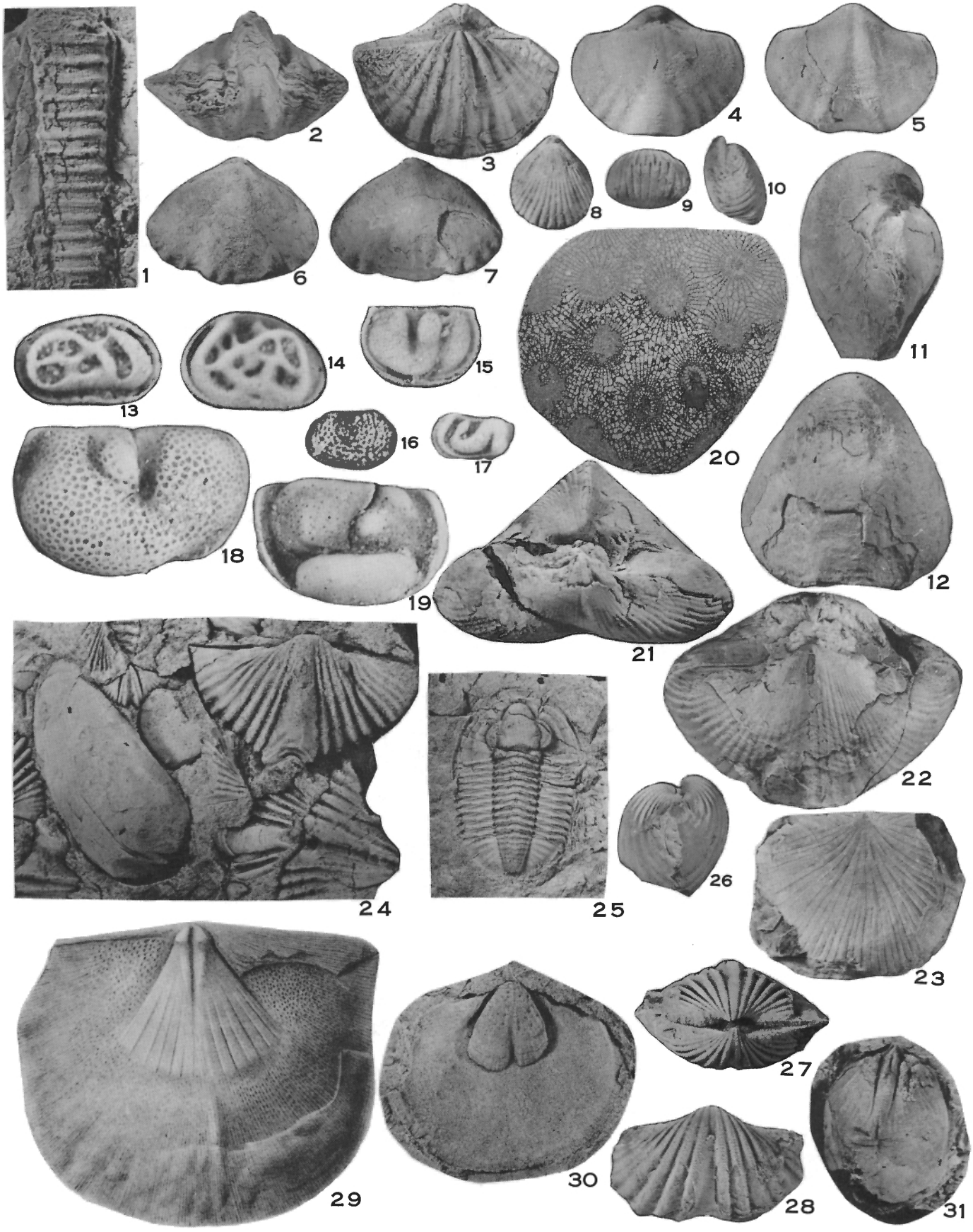
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CARBONIFEROUS AND PERMIAN FAUNAS

by E. W. Bamber and M. J. Copeland

Most major groups of marine invertebrate fossils are present in Carboniferous and Permian rocks. The characteristic fossils of these systems are productoid and spiriferid brachiopods, fenestrate bryozoans, rugose corals, and foraminifers.

Western and Northern Canada

Mississippian

The main emphasis for dating Mississippian rocks in Canada has been on brachiopods, corals, rare ammonoids, and more recently on foraminifers and ostracods. The remainder of the fauna is too rare or too poorly known to be stratigraphically useful. The ammonoids, which are useful for intercontinental correlation, occur mainly in the black shale facies. It is therefore difficult to integrate the dates afforded by ammonoids with those from other faunas that occur mainly in calcareous rocks.

Several zonal schemes have been proposed for the eastern Cordillera, based mainly on brachiopods, corals, and foraminifers (Nelson, 1961; McKay and Green, 1963). Only the macrofauna is discussed here. The most complete faunal sequence is found in the Rocky Mountains from the Banff area south to the United States border. These macrofossils have closer affinities with those from the Mississippi Valley region and the western United States and northern Mexico, than with the European Lower Carboniferous fauna. Their age is therefore given in terms of the standard Mississippi Valley sequence. Several species are very widespread and useful for correlation within North America. Table XI-5 gives the ranges of some of the most useful macrofossils in the Mississippian of western and northern Canada, but is not intended as a formal sequence of zones.

Lower Mississippian fossils are widespread in the eastern Cordillera and in subsurface from the United States border north to the southwestern District of Mackenzie, and are locally present in north–central British Columbia.

Rocks of Kinderhookian age contain mainly brachiopods, such as the distinctive *"Platyrachella" rutherfordi* (Warren) and the split ribbed *Spirifer esplanadensis* Brown and *Spirifer missouriensis* Swallow. The last mentioned brachiopod, and several brachiopod species associated with it, also occur in the Kinderhookian of the Mississippian Valley. The oldest lithostrotionid coral in Canada—*Lithostrotionella microstylum* (White)—appears approximately at the Kinderhook–Osage boundary. This species is a useful guide fossil, since it is widespread in the Mississippi Valley region and United States Cordillera at a stratigraphic position similar to that in western Canada.

Among the most common fossils in the lower Osagian rocks in the Rocky Mountains are the spiriferid brachiopods such as *Spirifer* cf. *rowleyi* Weller and *Spirifer minnewankensis* Shimer. The latter species, however, includes a rather heterogeneous group of brachiopods and has been identified from rocks as young as late Osagian. Both species may range into the upper Kinderhookian. The stratigraphically lowest phaceloid lithostrotionid coral, *Lithostrotion* (*S.*) *mutabile* (Kelly), is present but rare in the lower to middle Osagian. The early Osagian spiriferids are succeeded by *Spirifer banffensis* Warren, *S. rundlensis* Warren, and several other brachiopods, but this fauna is rare, and rocks of middle to late Osagian age are generally unfossiliferous in western Canada. The phaceloid coral *L.* (*S.*) *sinuosum* (Kelly), ranges from late Osagian to early Meramecian age. Very few fossils, other than this species are present at this level. The horn coral *Vesiculophyllum* ranges through rocks of latest Kinderhookian to early Meramecian age. It is one of the most abundant corals.

Early Meramecian macrofossils are not abundant in western Canada and little work has been done on them. It is therefore difficult to define the boundary between the Lower and Upper Mississippian. Rocks that are probably of early Meramecian age typically contain unnamed horn corals, the two colonial corals *Lithostrotion* (*S.*) *oculinum* and *L.* (*S.*) *sinuosum,* and rare horn corals of the genus *Ekvasophyllum. L.* (*S.*) *oculinum* is a widespread species that has been found in the Cordillera and in subsurface from Arizona to northern British Columbia. It appears to be restricted to lower Meramecian rocks. The middle and upper Meramecian are marked by an abundant and diversified coral fauna, dominated by species of *Lithostrotionella, Lithostrotion* (*Siphonodendron*), and *Ekvasophyllum.* Several productoid and other brachiopods are also present. Above the main coral-bearing interval is a distinctive fauna including the horn coral *Faberophyllum* with abundant *?Gigantoproductus* and other brachiopods. This fauna is widespread in the southern Rocky Mountains but appears to be absent north of the Banff area. The brachiopod *Spirifer* cf. *pellaensis* Weller begins high in the Meramecian and ranges up through the Chesteran. *Lithostrotion* (*S.*) *genevievensis* Easton provides a useful key for correlation, since it occurs in the Rocky Mountains, in the northern Yukon (British Mountains), and is found in the St. Genevieve Limestone of the Mississippi Valley region (uppermost Meramecian). It has a considerable stratigraphic range in Canada, and may range from the uppermost Meramecian into the Chesteran.

Chesteran rocks in western Canada contain a varied brachiopod fauna and relatively few corals. *Spirifer leidyi* Norwood and Pratten is characteristic and abundant within the uppermost Meramecian and throughout the Chesteran of the Rocky Mountains and occurs in Ches-

TABLE XI-5 *Distribution and range of Carboniferous and Permian invertebrate fossils*

teran rock of the Mississippi Valley region. Lower Chesteran rocks contain *Inflatia* cf. *inflata* McChesney, along with the genera *Girtyella, Dielasma, Rhipidomella, Orthotetes,* and others. The middle to upper Chesteran is characterized by several brachiopods including *Diaphragmus elegans* Norwood and Pratten, *Eumetria vera* (Hall) and *Punctospirifer transversa* (McChesney), which are found in the Chesteran of the United States. The highest rocks assigned to the Chesteran in western Canada contain very little other than spiriferid brachiopods, and their exact age within the Chesteran is not known. There may be a hiatus between these and the overlying Morrowan rocks, but the faunal evidence is not conclusive.

Upper Mississippian fossils are abundant in the eastern Cordillera as far north as the South Nahanni River valley, and also occur in the northern Yukon and north–central British Columbia (Mamet and Gabrielse, 1969), but Meramecian and Chesteran colonial and horn corals occur in the Milford Group in the Lardeau map-area west of the Rocky Mountain Trench, southern British Columbia. In parts of Alberta and northeastern British Columbia, Chesteran and uppermost Meramecian fossils are absent because of erosion. In the mountains from Peace River north to the southeastern Yukon, the westernmost Mississippian rocks of the Besa River Formation are mainly black shale which contains only rare ammonoids, nautiloids, and pelecypods. These ammonoids are locally present in northern Yukon and include *Goniatites* and *Cravenoceras* which indicate correlation with the Viséan of Europe. They have not been found in association with the abundant coral–brachiopod fauna to the east in British Columbia.

In the arctic islands, Chesteran plants are the only Mississippian fossils that have been reported.

Pennsylvanian

Little is known of Pennsylvanian fossils in western and northern Canada. The most reliable Pennsylvanian age determinations are based on ammonoids and fusulinid foraminifers, although several other phyla are present.

Lower Pennsylvanian (Morrowan) fossils occur in the Rocky Mountains from the Jasper area south to the United States border. They include mainly brachiopods, such as *Neospirifer praenuntius* Easton, *Spirifer occiduus* Sadlick, and *Antiquatonia*? cf. *hermosana* Girty. The last species is restricted to a narrow interval low in the sequence. Morrowan ammonoids with brachiopods and corals occur at Barrow Dome, Melville Island.

Middle Pennsylvanian rocks are fairly widespread in northern and western Canada. They contain the first well-developed fusulinids, rare ammonoids, and numerous brachiopods, corals, bryozoans, and other phyla. The fauna is largely undescribed, and the main work that has been done deals with the fusulinids and ammonoids. Middle Pennsylvanian fusulinids are most abundant in the arctic islands where several genera occur, including *Profusulinella* and *Fusulinella,* which are worldwide zone fossils for rocks of this age. Definite Middle Pennsylvanian fossils are known in the northern Yukon from only one locality in the Ogilvie Mountains, where *Pseudostaffella* occurs. Desmoinesian fusulinids including *Fusulina, Fusulinella,* and others occur locally in south–central British Columbia. In the Rocky Mountains of southern Alberta and British Columbia there are Middle Pennsylvanian brachiopods, scaphopods, pelecypods, bryozoans, and fusulinids. This fauna and the underlying Morrowan fauna appear to be closely related to those in the western and central United States.

Fossils of definite Late Pennsylvanian age have not been found in western Canada. Their absence is caused, in some areas at least, by an erosional break beneath the overlying Permian beds, and may indicate a very widespread hiatus in the faunal succession. In the arctic islands, Upper Pennsylvanian rocks are dated by several fusulinid genera, including *Triticites* and *Quasifusulina.*

Permian

Permian rocks of northern and western Canada have been dated mainly by fusulinid foraminifers, ammonoids and brachiopods, but corals, bryozoans, and other phyla are locally abundant.

Fossils of Wolfcampian age occur in much of the eastern Cordillera, in western and central British Columbia, and in the arctic. Only the fusulinids have been studied. The remainder of the fauna is undescribed. In northeastern British Columbia and the southwestern District of Mackenzie there is a Lower Permian fauna consisting mainly of brachiopods, which is probably Wolfcampian, but may be Leonardian. This fauna includes *Spiriferella, Pterospirifer, Choristites, Waagenoconcha,* and other brachiopods. A similar fauna occurs below the Wolfcampian fusulinid *Schwagerina* cf. *paralinearis* Thorsteinsson in the Ogilvie Mountains of northern Yukon.

Wolfcampian faunas are succeeded by a widespread boreal fauna that has been found in the arctic islands, northern Yukon, central and western British Columbia, and in the Rocky Mountains as far south as Crowsnest Pass, British Columbia. This assemblage consists largely of brachiopods with some corals and other groups. It has been dated in several places by associated fusulinids and ammonoids as mainly Leonardian to early Guadalupian. The upper age limit of the fauna is in doubt in the arctic, however, as several of the genera that it includes range upward into rocks containing no fusulinids. Elements of the boreal fauna occur as far south as Oregon, U.S.A., but it is found mainly in a circumpolar belt including the arctic areas of North America, Europe, and Asia. The northern fauna has little in common with that from the standard sequence of southwestern United States and its age assignment, in terms of this sequence, is in doubt.

The youngest Permian fauna of central British Columbia is distinct from other Permian assemblages in Canada. It contains fusulinids such as *Yabeina* and other neoschwagerinids, which appear to overlie the boreal fauna.

Yabeina is the highest in the sequence and is thought to be the youngest Permian form in North America. These neoschwagerinids have also been found in the Mediterranean area, southern and eastern Asia, and the northwestern United States. Their absence from the arctic and southwestern United States faunas has led to uncertainty regarding age relationships between the youngest Permian of these regions and that of central British Columbia.

Atlantic Provinces

Carboniferous rocks of the Atlantic Provinces are divided into six units, in ascending order, the Horton, Windsor, Canso, Riversdale, Cumberland, and Pictou Groups. Their fauna closely resembles that of the standard European sequence. European age terminology is, therefore, applied to these rocks.

Lower Carboniferous

The Horton Group contains abundant plants and a non-marine fauna of Tournaisian age characterized by arthropods (*Hollinella*?, *Euestheria, Eoleaia*) and fish. These rocks are unconformably overlain by Viséan strata of the Windsor Group which contain marine fossils. The Windsor has been subdivided by Bell (1929) into five subzones as follows:

Upper Windsor. Zone of *Martinia galataea*
- Subzone E. *Caninia dawsoni, Chonetes politus*
- Subzone D. *Ovatia? semicubiculus*
- Subzone C. *Dibunophyllum lambii* and *Nodosinella priscilla*

Lower Windsor. Zone of *Composita dawsoni*
- Subzone B. *Diodoceras avonensis*
- Subzone A. Sparse fauna

Upper Carboniferous

The Canso Group contains a Namurian flora and a distinctive local assemblage of non-marine and brackish-water arthropods (*Leaia, Euestheria, Belinurus, Dithyrocaris, Beyrichiopsis)* and pelecypods *(Anthracomya, Carbonicola*).

The overlying fauna of the Riversdale and Cumberland Groups ranges through Westphalian A and B. Distinctive elements of this fauna include the arthropods *Anthrapalaemon, Carbonita, Xylobius,* the gastropod *Pupa,* and the pelecypods *Anthracomya* and *Naiadites.*

The uppermost Carboniferous unit, the Pictou Group, contains abundant insects (Blattoidea), other arthropods (*Carbonita, Euproops*) and pelecypods (*Anthraconauta*). This fauna is characteristic of Westphalian C and D rocks both in Europe and Nova Scotia.

Permian

Permian vertebrates (Langston, 1963) and a limited flora occur on Prince Edward Island. All were deposited in a non-marine environment and are considered to be early Permian.

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PLATE XII. Mississippian Fossils of Western Canada

(All figures natural size unless otherwise stated)

Figure 1 *Cleiothyridina lata* Shimer. Kinderhookian, Alberta; 21735.

Figure 2 *Spirifer missouriensis* Swallow. Kinderhookian, Alberta; 21736.

Figure 3 *Rhynchotetra usheri* Brown. Kinderhookian, Alberta; 21737.

Figure 4 *Cleiothyridina* cf. *obmaxima* (McChesney). Kinderhookian, Alberta; 21738.

Figure 5 "*Platyrachella*" *rutherfordi* (Warren). Kinderhookian, Alberta; 21739.

Figure 6 *Marginatia jasperensis* (Warren). Kinderhookian, Alberta; 10081.

Figure 7 *Spirifer esplanadensis* Brown. Kinderhookian, Alberta; 21741.

Figure 8 *Lithostrotionella microstylum* (White). x2, late Kinderhookian to early Osagian, Alberta; 16840.

Figure 9 *Spirifer minnewankensis* Shimer. Kinderhookian to early Osagian, Alberta; 21742.

Figure 10 *Spirifer* cf. *rowleyi* Weller. Osagian, Alberta; 21743.

Figure 11 *Lithostrotionella micra* Kelly. x2, Osagian, Alberta; 9648.

Figure 12 *Spirifer rundlensis* Warren. Osagian, Alberta; 8921.

Figure 13 *Vesiculophyllum* sp. Osagian, Mackenzie; 10570.

Figure 14 *Lithostrotion* (*Siphonodendron*) *oculinum* Sando. x2, Meramecian, subsurface, British Columbia; 21744.

Figure 15 *L.* (*S.*) *sinuosum* (Kelly). x2, late Osagian–Meramecian, Alberta; 16843.

Figure 16 *Spirifer bifurcatus* Hall. Meramecian, Alberta; 21745.

Figure 17 *Ekvasophyllum inclinatum* Parks. x2, Meramecian, Alberta; 21746.

Figure 18 *Lithostrotionella pennsylvanica* (Shimer). x2, Meramecian, Alberta; 16829.

PLATE XIII. Mississippian and Pennsylvanian Fossils of Western Canada

(All figures natural size unless otherwise stated)

Figure 1 *Lithostrotion (Siphonodendron) whitneyi* of Meek. x2, Meramecian, Alberta; 16832.

Figure 2 *Thysanophyllum astraeiforme* (Warren). x2, Meramecian, Alberta; 8911.

Figure 3 *Goniatites crenistria* Phillips. x2½, Viséan, northern Yukon, from Sellers and Furnish, 1960, Pl. 141, fig. 4.

Figure 4 *Spirifer* cf. *pellaensis* Weller. Meramecian, Alberta; 21747.

Figure 5 *Faberophyllum* sp. x2, Meramecian, Alberta; 21748.

Figure 6 ?*Gigantoproductus* sp. Meramecian, Alberta; 21749.

Figure 7 *Lithostrotion* (*Siphonodendron*) cf. *pauciradiale* (McCoy). x2, Chesteran, Alberta; 21750.

Figure 8 *L.* (*S.*) *genevievensis* Easton. x2, Chesteran, Alberta; 16844.

Figure 9 *Spirifer leidyi* Norwood and Pratten. Chesteran, Alberta; 21751.

Figure 10 *Punctospirifer transversa* (McChesney). Chesteran, Alberta; 21752.

Figure 11 *Neospirifer* cf. *triplicatus* (Hall). Morrowan, Alberta; 21753.

Figure 12 *Spirifer occiduus* Sadlick. Morrowan, Alberta; 21754.

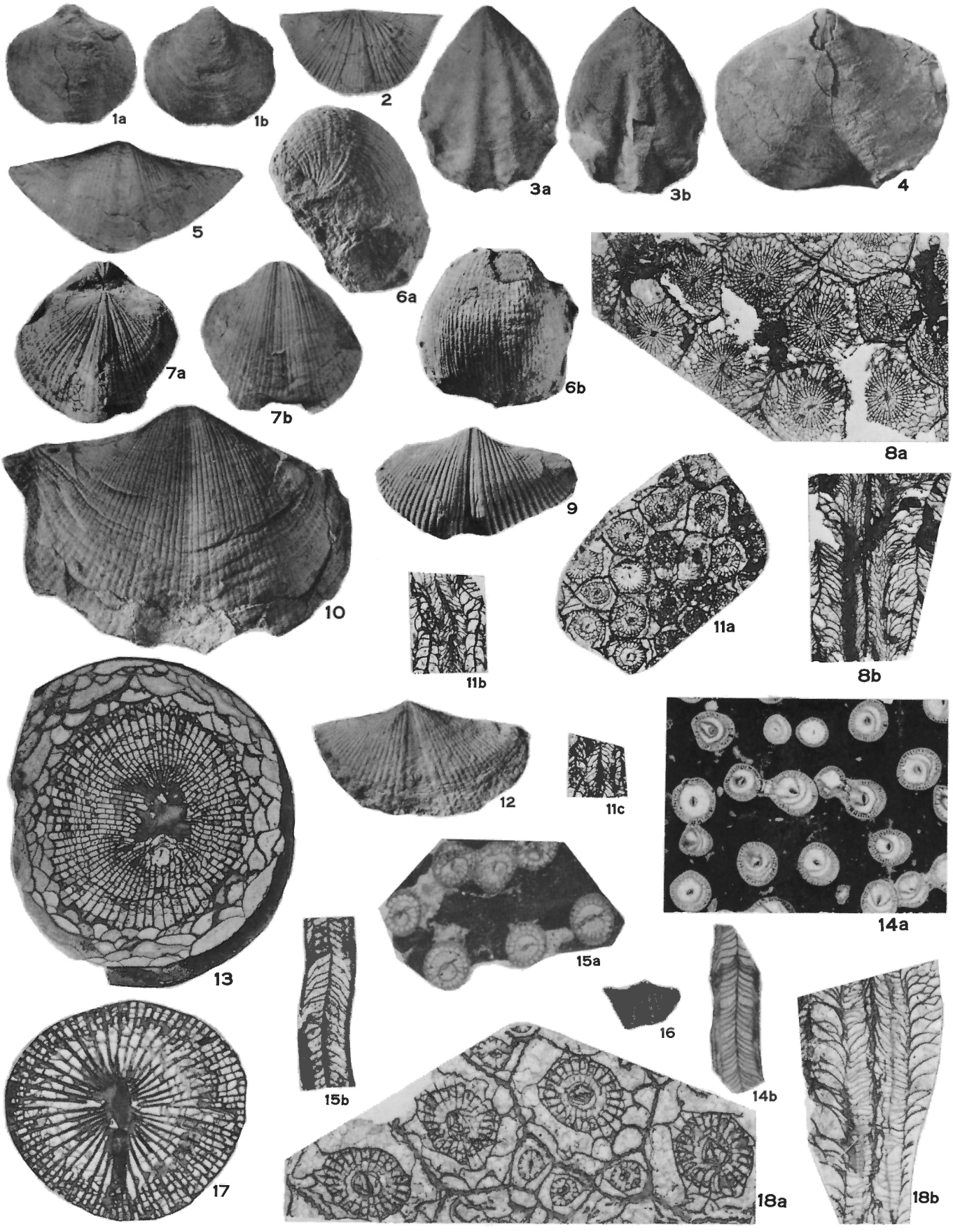
Figure 13 *Rhynchopora* cf. *magnicosta* Mather. Morrowan, Alberta; 21755.

Figure 14 ?*Reticulariina* sp. Morrowan, Alberta; 21756.

Figure 15 *Antiquatonia*? cf. *hermosana* (Girty). Morrowan, Alberta; 21757.

Figure 16 *Neospirifer praenuntius* Easton. Morrowan, Alberta; 21758.

Figure 17 *Plagioglypta* sp. Atokan, Alberta; 21759.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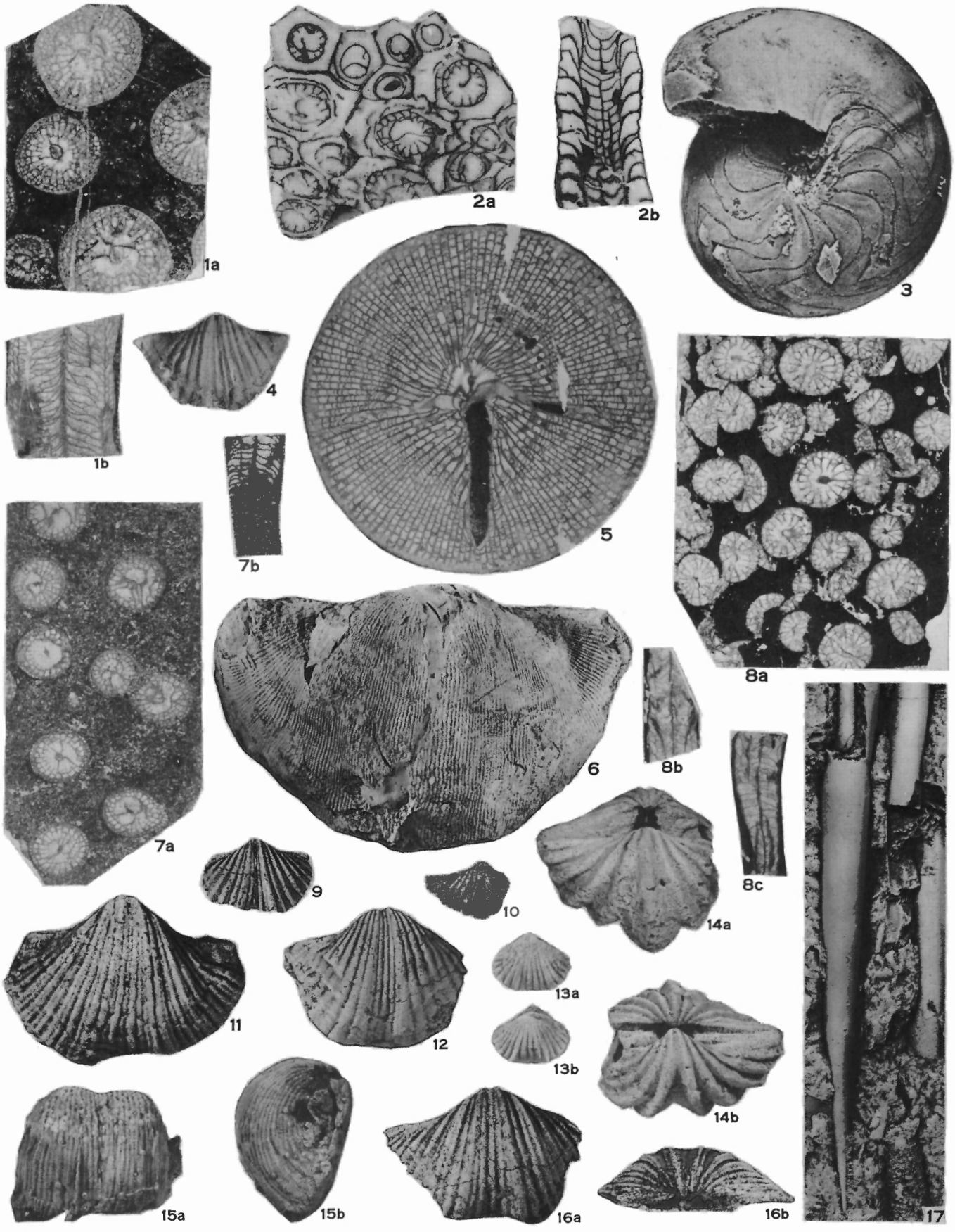
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PLATE XIV. Permian Fossils of Western and Northern Canada

(All figures natural size unless otherwise stated)

Figure 1 *Parafusulina belcheri* Thorsteinsson. x5, Leonardian, Grinnell Peninsula; 13977.

Figure 2 *Neospirifer* sp. Early Permian, northern Yukon; 21760.

Figure 3 *Antiquatonia* cf. *neoninflatus* Licharew. Leonardian, Grinnell Peninsula; 13527.

Figure 4 *Waagenoconcha* sp. Early Permian, northern Yukon; 21761.

Figure 5 *Horridonia* sp. Early Permian, northern Yukon; 21762.

Figure 6 *Pterospirifer* cf. *alatus* (Schlotheim). Leonardian, Grinnell Peninsula; 13752.

Figure 7 *Daubichites fortieri* (Harker). Leonardian, Grinnell Peninsula; 13773.

Figure 8 *Streptorhynchus kempei* Andersson. Leonardian or Guadalupian, Grinnell Peninsula; 13517.

Figure 9 *Spirifer striato-paradoxus* Toula. Leonardian or Guadalupian, Grinnell Peninsula; 13748.

Figure 10 "*Stenoscisma*" *pliçatum* (Kutorga). Leonardian or Guadalupian, Grinnell Peninsula; 13737.

Figure 11 *Spiriferella* sp. Wolfcampian or Leonardian, northern Yukon; 21763.

Figure 12 *Muirwoodia* sp. Wolfcampian or Leonardian, northern Yukon; 21764.

Figure 13 *Megousia* sp. Guadalupian, northern Yukon; 21765.

Figure 14 *Megousia* sp. Guadalupian, northern Yukon; 21766.

Figure 15 *Caninophyllum belcheri* (Harker). Leonardian, Grinnell Peninsula; 13503a.

Figure 16 *Kochiproductus freboldi* (Stepanow). Leonardian or Guadalupian, Grinnell Peninsula; 13534.

PLATE XV. Carboniferous Fossils of Eastern Canada

Figure 1 *Composita dawsoni* (Hall and Clarke). x1, Viséan, Hants co., N.S.; 7501a.

Figure 2 *Diodoceras avonensis* (Dawson). x1, Viséan, Hants co., N.S.; 4373.

Figure 3 *Conularia planicostata* Dawson. x1, Viséan, Irish Cove, N.S.; 21767.

Figure 4 *Aviculopecten lyelli* Dawson. x1, Viséan, Hants co., N.S.; 7553.

Figure 5 *Dibunophyllum lambii* Bell. x2, Viséan, Hants co., N.S.; 13589g.

Figure 6 *Cypridina acadica* Bell. x4, Viséan, Hants co., N.S.; 7742.

Figure 7 *Koninckophyllum* sp. x2, Viséan, Hants co., N.S., from Lewis, 1935, Pl. VII, fig. 2a, b.

Figure 8 *Spirifer adonis* Bell. x1, Viséan, Nova Scotia, holotype, Peabody Museum, Yale University.

Figure 9 *Ovatia lyelli* (de Verneuil). x1, Viséan, Hants co., N.S.; 7952.

Figure 10 *Martinia galataea* Bell. x1, Viséan, Hants co., N.S.; 7523.

Figure 11 *Schellwienella kennetcookensis* Bell. x1, Viséan, Hants co., N.S.; 7654.

Figure 12 *Diaphragmus avonensis* (Bell). x1, Viséan, Hants co., N.S.; 7953.

Figure 13 *Amplexizaphrentis enniskilleni* var. *minas* (Dawson). x2, Nova Scotia, from Lewis, 1935, Pl. V, fig. 1a.

Figure 14 *Spirifer nox* Bell. x1, Viséan, Hants co., N.S.; 7532.

Figure 15 *Euproops thompsoni* Raymond. x2½, Viséan, Cumberland co., N.S., from Copeland, 1957, Pl. 67, fig. 9.

Figure 16 *Beyrichiopsis lophota* Copeland. x30, Namurian, Cape Breton co., N.S.; 10387.

Figure 17 *Leaia baentschiana* (Beyrich). x2, Namurian, Richmond co., N.S.; 12793.

Figure 18 *Belinurus grandaevus* Jones and Woodward. x3, Namurian, Cumberland co., N.S.; 12804b.

Figure 19 *Leaia acutilirata* Copeland. x5½, Namurian, Pictou co., N.S.; 10396.

Figure 20 *Asmussia tenella* (Bronn). x6½, Westphalian, Pictou co., N.S.; 12853.

Figure 21 *Anthrapalaemon dubius* (Milne-Edwards). x1½, Westphalian, Cumberland co., N.S.; 12821.

Figure 22 *Carbonita agnes* (Jones). x30, Westphalian, Inverness co., N.S.; 12841.

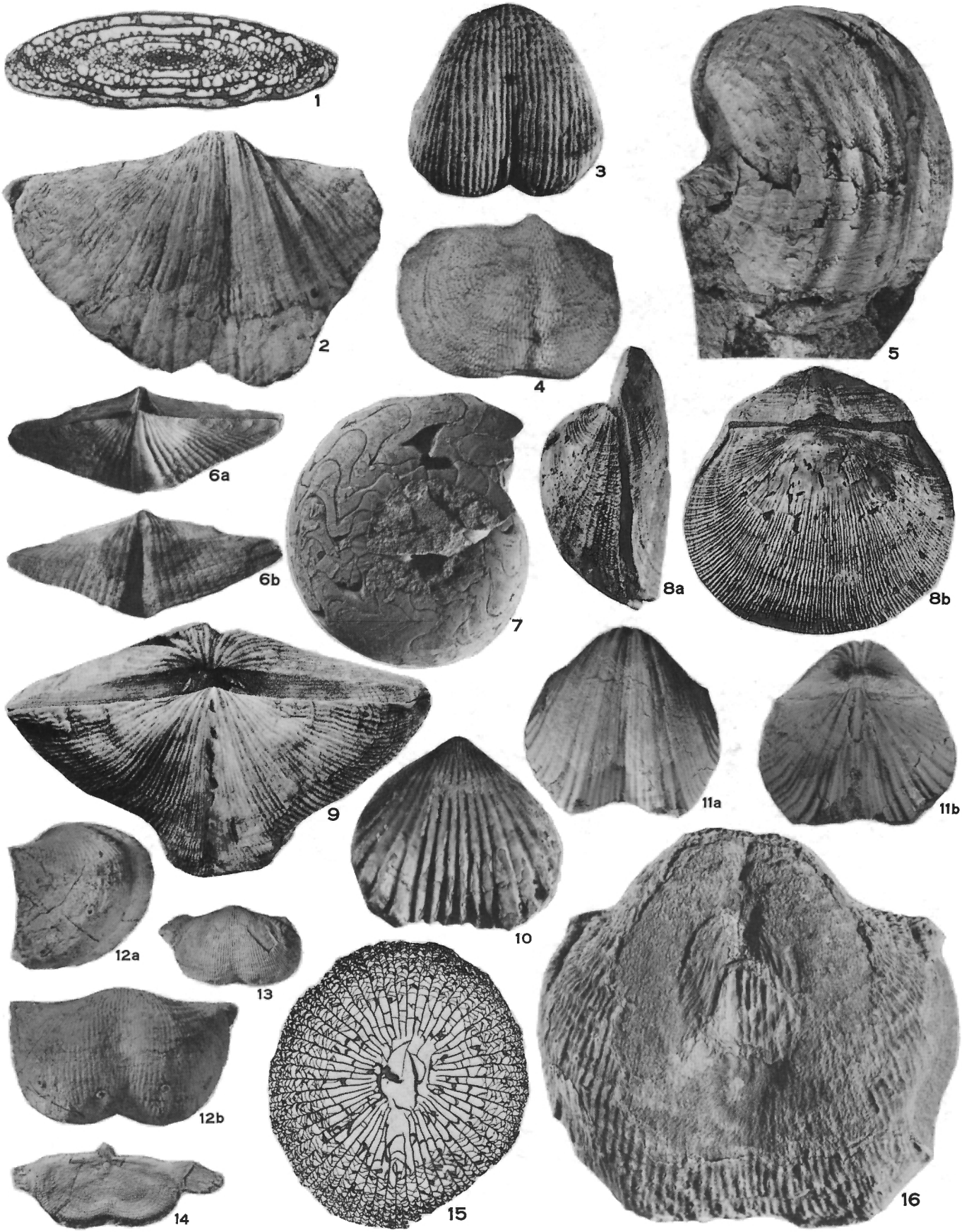
Figure 23 *Hilboldtina evelinae* (Jones). x30, Westphalian, Cumberland co., N.S.; 12832.

Figure 24 *Gutschickia bretonensis* Copeland. x30, Westphalian, Inverness co., N.S.; 12836b.

Figure 25 *Phylloblatta*? sp. x1½, Westphalian, Pictou co., N.S.; 12847e.

Figure 26 *Archimylacris morienensis* Copeland. x2½, Westphalian, Cape Breton co., N.S.; 10394.

Figure 27 *Euproops amiae* Woodward. x2, Westphalian, Cape Breton co., N.S.; 12808e.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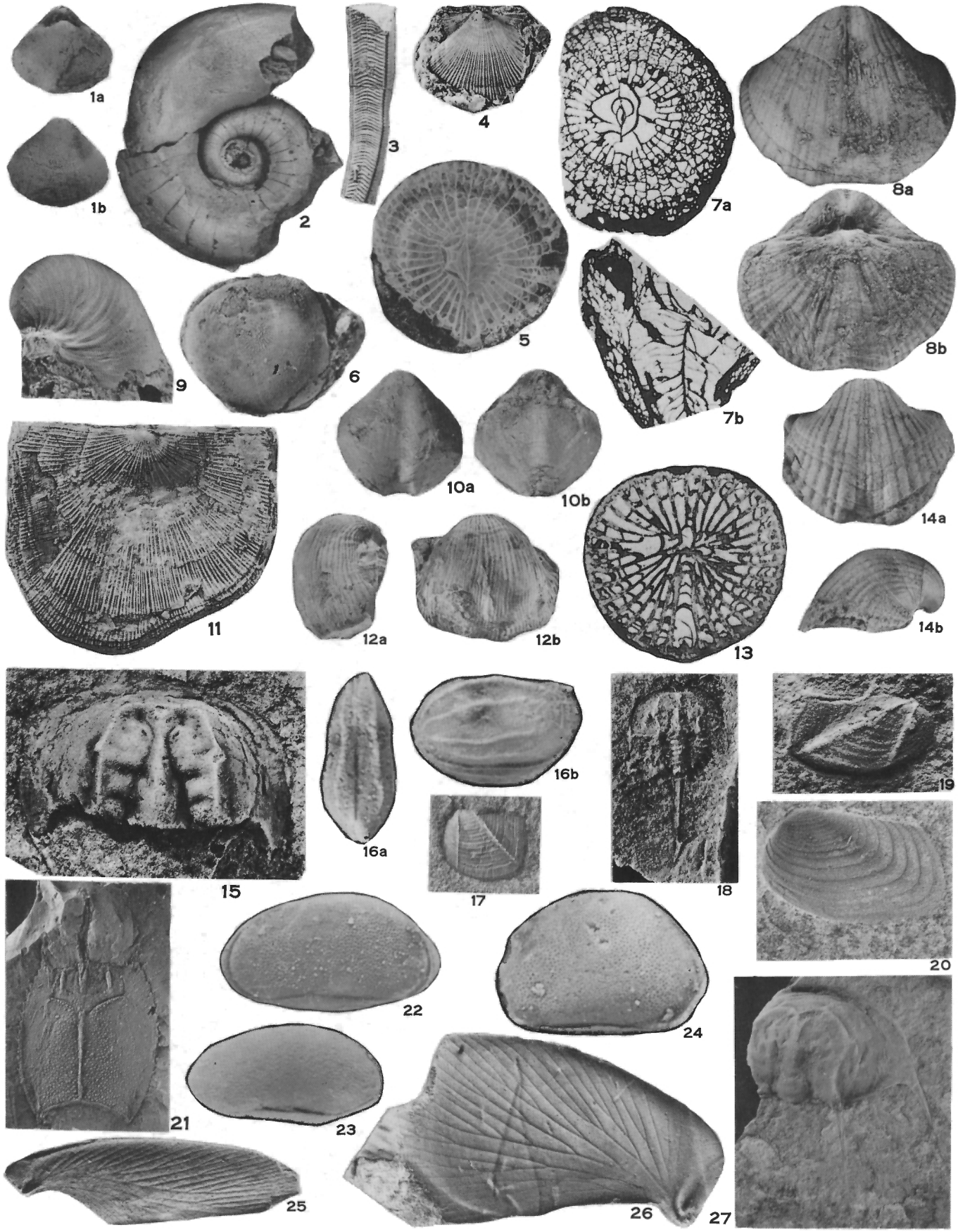
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MARINE TRIASSIC FAUNAS

by E. T. Tozer

Triassic rocks containing marine faunas occur in the arctic islands and in western Canada. In the west exposures are confined to the Cordillera but the rocks also extend beneath the plains of northern Alberta and northeastern British Columbia. The following groups of marine organisms are known from the Triassic rocks of Canada: sponges, scleractinian corals; hydrozoans; crinoids (columnals only); echinoids; ophiuroids; bryozoans, terebratulid, rhynchonellid and spiriferid brachiopods; a great variety of bivalve molluscs; gastropods; ammonoids, nautiloids; belemnoids; ganoid fish and ichthyosaurs. The ammonoids and bivalves have proved most useful for dating rocks and have received most study. Corals, a bryozoan, brachiopods, nautiloids, and fish have received some attention, but many of the Triassic fossils have not yet been studied in any detail.

In discussing the distribution of Triassic faunas in Canada, it is convenient to distinguish four main regions: the arctic islands, and the eastern, western, and northern parts of the Cordillera. The arctic islands region refers to the Triassic deposits of Sverdrup Basin. The eastern Cordillera region includes the formations of the Foothills and Eastern Ranges of the Rocky Mountains in northeastern British Columbia and western Alberta. The western Cordillera region comprises the parts of British Columbia and southern Yukon that lie west and south of Rocky Mountain and Tintina Trenches. The northern Cordillera region, which, strictly speaking, represents the northern part of the eastern Cordillera, includes the parts of Yukon and the Northwest Territories that lie northeast of Tintina Trench.

Boundaries of the Triassic System

Virtually the whole of Triassic time is represented by marine deposits in Canada, although a complete faunal sequence is not known in any one region. The oldest faunas, known only from the arctic islands, are correlated with the *Otoceras* beds of the Himalayas, which are generally taken to define the base of the Triassic System. The *Otoceras* beds of Canada rest disconformably upon Permian rocks. As previously noted, these Permian rocks do not represent the youngest part of that system. In Canada, as in most and probably all parts of the world, there was a break in marine sedimentation between the Permian and the Triassic.

In most parts of Canada latest Triassic time is unrepresented by marine rocks. Upper Norian marine rocks occur in all regions but marine Rhaetian beds have been recognized only at Tyaughton Creek, British Columbia, in the western Cordillera. Hettangian, lowermost Jurassic, rocks are also present at this locality, and there is apparently no detectable stratigraphic break between the Triassic and Jurassic rocks. In the arctic islands there are non-marine beds, the upper part of Heiberg Formation, lying between marine upper Norian and Sinemurian beds. These non-marine strata may include beds of Rhaetian age.

The boundaries of the Triassic are thus marked by evidence of widespread marine regression. At the base of the Triassic the interval of regression lies mainly, or entirely, within the Permian. At the top, however, the regressive interval commonly spans both latest Triassic and earliest Jurassic time, but in one place at least this gap is bridged by marine sediments.

Faunal Facies

For present purposes two types of faunas may be distinguished among the assemblages known in Canada.

First there are the faunas in which the dominant fossils are cephalopods, mainly ammonoids, and thin-shelled bivalves, such as *Claraia, Posidonia, Daonella, Halobia,* and *Monotis.* The exact environment in which these animals lived is unknown, but it is certain that they do not represent a normal shallow-water benthonic assemblage. They probably did not live on the sea floor and are accordingly referred to as the pelagic faunas.

The second marine faunal facies is clearly benthonic and includes sponges, corals, echinoderms, bivalves, and brachiopods. The bivalve faunas include trigoniids and other dimyarian groups, ostreids, limids, pernids, pectinids, etc., commonly in rich variety. These bivalve faunas are quite unlike the monotonous suite of Pectinacea *(Claraia,* etc.) that characterizes the pelagic facies. Ammonoids and pelagic bivalves are occasionally associated with the benthonic forms, but they are generally rare. At some levels ammonoids are relatively abundant in both the benthonic and pelagic assemblages. *Nathorstites* (upper Ladinian) is one example; this genus is commonly found with brachiopods and ostreid bivalves, but it nevertheless achieves its greatest abundance in the pelagic facies. In the upper Norian *Rhabdoceras* occurs in both benthonic and pelagic assemblages.

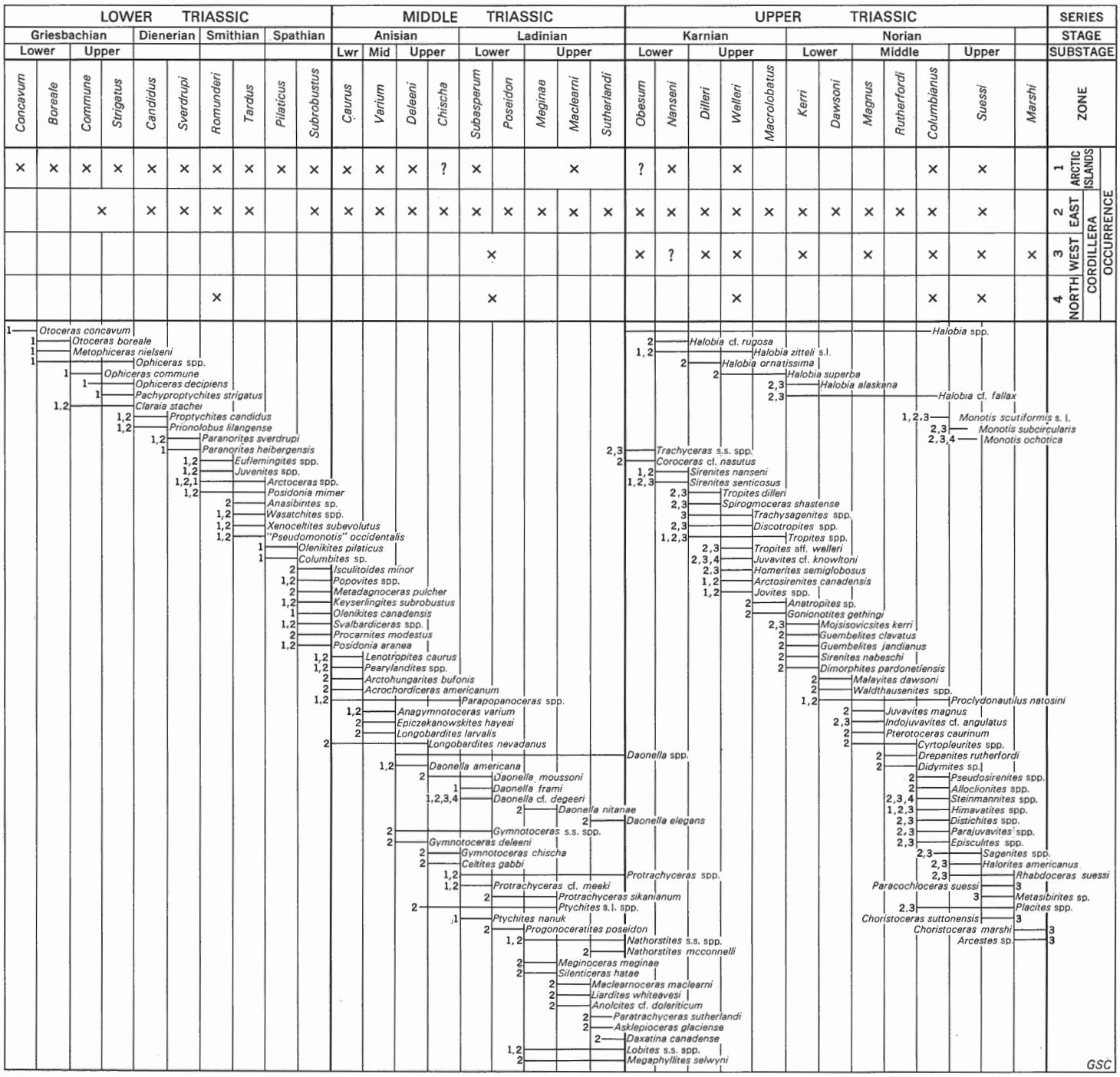
Faunal Provinces

The pelagic faunas of the Triassic include many cosmopolitan genera and even species. Consequently correlation of rocks that contain pelagic faunas is usually a fairly easy matter. The pelagic faunas of low latitudes, as in British Columbia, appear to show greater taxonomic diversity than contemporary assemblages in the arctic islands. For example, only three ammonoid genera are known in the upper Ladinian of arctic islands whereas in northeastern British Columbia about a dozen genera occur in each of the three zones of the upper Ladinian. These differences are not obviously related to conditions of sedimentation and more likely indicate a taxonomic diversity gradient

related to variations in the temperature of the Triassic seas. The apparent decrease in diversity towards the north suggests that the Triassic seas that covered the site of the arctic islands were cooler than those of the Cordillera. Comparable differences between the faunas of the eastern Cordillera and the arctic islands occur in the Dienerian, Spathian, Anisian, Karnian, and Norian. Despite these apparent gradients there are, at all levels, a few cosmopolites that satisfy the needs of correlation. The pelagic faunas provide no evidence that isolated zoogeographic provinces existed in Triassic time.

The benthonic faunas of the western Cordillera region are in some respects unlike those of the other regions and seem to form a distinct zoogeographic province. These

TABLE XI-6 *Range and distribution of Triassic pelagic faunas. (Numbers refer to occurrence in the appropriate region.)*



faunas are all Upper Triassic. They include the only Triassic coral faunas known in Canada. Also present are megalodont bivalves, and a rich, but as yet poorly known, fauna of brachiopods and gastropods. The coral faunas occur at several levels in the Karnian and Norian and attain a latitude of 61°N in southern Yukon. There can be no doubt that these faunas testify to warm water conditions. At several levels, notably in the Dilleri and Welleri Zones of the Karnian, and in the Columbianus and Suessi Zones of the Norian, pelagic faunas, identical with those of the eastern Cordillera region are also present in the western Cordillera. Whatever the factor or feature that restrained Triassic corals from reaching the eastern and northern parts of the Cordillera, and also the arctic islands, it did not constitute a comparable barrier to the pelagic faunas.

As already mentioned, the evidence from the pelagic faunas suggests that the Triassic deposits of the eastern Cordillera region were deposited in warmer waters than those of the arctic islands. This presents no problems; the arctic islands are at a higher latitude than the eastern Cordillera today, and according to the paleomagnetic data this was also true in Triassic time. The Triassic rocks of the western Cordillera were evidently deposited in even warmer water than those of the eastern Cordillera. This is dramatically shown by the presence in southern Yukon of thick bioclastic formations containing coral faunas that range in age from Karnian to upper Norian. No comparable rocks occur in the contemporary deposits of the eastern Cordillera which are today at a lower latitude. The close similarities between the pelagic faunas of the eastern and western parts of the Cordillera, and the extraordinarily rich representation of cosmopolitan pelagic faunas in the eastern Cordillera show, without any doubt, that the eastern Cordillera was not isolated at any time between the Griesbachian and the upper Norian from the routes of migration open to the pelagic faunas. There is, therefore, no faunal evidence to support the placing of a land barrier between the sites of Triassic deposition now preserved in the eastern and western parts of the Cordillera. The unrivalled sequence of Norian ammonoid faunas preserved in the Pardonet Formation of Peace River show close relationships to the faunas of the Alps, Timor, and the Himalayas. There is no evidence that these faunas penetrated to the Peace River area via the arctic; nor did they come from the south or east. The Peace River area must have had free access to the Circumpacific and Tethyan seaways of Upper Triassic time.

Possibly the faunal peculiarities of the western Cordillera may be explained by post-Triassic tectonic events. There is evidence from various parts of the Cordillera that post-Triassic, right lateral transcurrent movement, of the order of 250 miles, has affected segments of the Cordillera. The Tintina Trench, according to Roddick, may define one such segment of the Cordillera. Structural interpretations of this nature introduce the possibility that the warm-water Triassic deposits of the western Cordillera were originally deposited south of the latitude of the contemporary rocks in the eastern Cordillera. Their present juxtaposition at comparable latitudes would thus be the result of movement of the western rocks to the north, in post-Triassic time. This, rather than an ecological barrier on the site of the Rocky Mountain and Tintina Trenches, seems a more satisfactory explanation for the observed anomalies of faunal distribution.

The Sequence of Triassic Faunas

The Triassic rocks of Canada provide a fairly continuous sequence of strata containing pelagic faunas. In these rocks the ranges of the faunas can be empirically determined, and a succession of zones forming a time-scale can be established. With the benthonic faunas this is not so. There are no continuous sequences with benthonic assemblages that cover more than two stages at most. The ranges of the various benthonic faunas are thus not well known. The record provided by the pelagic faunas permits something that approaches continuous observation, throughout Triassic time. In contrast, the record for the benthonic groups represents little more than an occasional glimpse of the life of the time.

In the pelagic succession thirty-one ammonoid zones have been recognized in Canada and these zones are grouped into eight stages (Table XI-6). The four Lower Triassic stages are defined in arctic Canada; those of the Middle and Upper Triassic were defined in Alpine Europe. The range in time and geographic distribution of the principal pelagic guide fossils are also shown. The faunal sequence in the arctic islands and the eastern Cordillera region is far more complete than in the western and northern parts of the Cordillera. The incompleteness of the sequence in the northern part partly reflects lack of knowledge. In the western Cordillera the absence of Lower Triassic faunas and the rarity of Middle Triassic is probably real for, as shown in Chapter VIII, widespread volcanic and orogenic activity occurred which may account for the rarity of faunas for this interval.

The record of benthonic faunas is very fragmentary. A few, however, deserve mention because they are locally useful for stratigraphic purposes. Dienerian rocks of the arctic islands, in the shallow-water facies, locally contain *Myalina* coquina. *Arcticopora christiei* from Ellesmere Island, which is one of the very few trepostome bryozoans known younger than Paleozoic, is probably Dienerian. The Anisian is not well represented by benthonic faunas, but *Spiriferina stracheyi* has been described from the eastern Cordillera.

In Ladinian time, benthonic faunas became fairly abundant in eastern Cordillera region and arctic islands. Spiriferid brachiopods and bivalves are most abundant in rocks of calcareous sandstone and siltstone facies, such as the "Grey Beds" of the eastern Cordillera, and the Schei Point Formation of arctic islands. The Ladinian and Karnian benthonic faunas are not obviously dissimilar and usually include one or two species of spiriferinid and

terebratuloid brachiopods, and up to a dozen species of bivalves, including representatives of *Meleagrinella, Oxytoma, Gervillia, Hoernesia, Entolium, Pecten, Ostrea, Lopha, Gryphaea, Plagiostoma, Mysidioptera, Mytilus, Modiola, Myophoria, Cardinia,* and *Pleuromya.*

A Karnian, probably upper Karnian, assemblage characterized by *Mysidioptera poyana* is fairly widely distributed in the eastern Cordillera region and may also occur in the southern Yukon. Another notable local fauna, probably also upper Karnian, characterized by *Gryphaea* cf. *arcuatiformis, Plicatula,* and other bivalves, is widespread in the upper Schei Point beds that rim the Sverdrup Basin.

The lower Norian in the eastern Cordillera is known mainly from pelagic faunas, but species of *Pleuromya* and *Gryphaea chakii* are locally present in lower Norian beds. The *Meleagrinella antiqua* fauna which occurs in the lower part of the Heiberg Formation of arctic islands is probably lower or middle Norian, but somewhat similar assemblages are also known from the underlying beds in the Blaa Mountain Formation that are definitely Karnian. *Oxytoma kiparisovae* is a member of this fauna and specimens of *Halobia* are also known. Typical dimyarian bivalves do not occur and it is possible that this assemblage represents some sort of pelagic fauna; at any rate, it is not a typical benthonic assemblage. Large megalodont bivalves (*Megalodon canadensis*) occur in coralline limestone immediately below *Monotis subcircularis* at Tyaughton Creek in the western Cordillera. These fossils are probably middle or upper Norian.

The richest Triassic benthonic fauna known in Canada is that of the Norian of the western Cordillera, where it occurs in the Lewes River Group of Yukon, the Tyaughton Group of southern British Columbia, the Sutton and Bonanza Groups of Vancouver Island, and in other unnamed formations. This fauna includes corals, sponges, echinoids, brachiopods, bivalves and gastropods, and cephalopods. Widely distributed species are *Spondylospira lewesensis, Pecten tyaughtonae, "Pecten" yukonense,* at least nine species of Trigoniidae (*"Myophoria" suttonensis, "Myophoria" cairnesi, "Trigonia" textilis,* etc.) *Cassianella lingulata, Plicatula perimbricata.* Also present are representatives of *Plagiostoma, Rhaetavicula, "Astarte", Modiolus,* etc. More than thirty species of bivalves are certainly present. Ammonoids, including *Rhabdoceras* and *Paracochloceras* occur at some localities associated with many of the benthonic fossils, and their presence shows that most of the occurrences of this fauna are upper Norian. It should be noted, however, that at least some species (e.g., "*Myophoria*" *suttonensis*) range up into the Marshi Zone, of Rhaetian age.

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1967: Middle and Upper Triassic spiriferinid brachiopods from the Canadian Arctic Archipelago; *Geol. Surv. Can.,* Bull. 155. (Includes most references to papers of Triassic brachiopods from the eastern Cordillera.)

McLearn, F. H.
1960: Ammonoid faunas of the Upper Triassic Pardonet Formation, Peace River Foothills, British Columbia; *Geol. Surv. Can.,* Mem. 311.

Tozer, E. T.
1967: A standard for Triassic time; *Geol. Surv. Can.,* Bull. 156. (Includes references to all previously published reports that deal with pelagic faunas, and many references to papers on benthonic faunas.)

PLATE XVI. Lower Triassic Ammonoids and Bivalves

Figures 1a, b *Otoceras boreale* Spath. Lower Griesbachian, Blind Fiord Formation, Ellesmere Island;14015.

Figures 2a, b *Ophiceras commune* Spath. Upper Griesbachian, Blind Fiord Formation, Axel Heiberg Island; 14030.

Figures 3a, b *Pachyproptychites strigatus* (Tozer). Upper Griesbachian, Blind Fiord Formation, Axel Heiberg Island; 14035.

Figure 4 *Claraia clarai* (Emmrich). Griesbachian, Sulphur Mountain Formation, Alberta; 14196.

Figure 5 *Claraia stachei* Bittner. Griesbachian, Grayling Formation, northeastern British Columbia; 14228.

Figures 6a, b *Prionolobus lilangense* (Krafft). Dienerian, Toad Formation, northeastern British Columbia; 21768.

Figures 7a, b *Paranorites sverdrupi* Tozer. Dienerian, Blind Fiord Formation, Ellesmere Island; 14277.

Figures 8a, b *Arctoceras blomstrandi* (Lindstrom). Smithian, Blind Fiord Formation, Ellesmere Island; 14067.

Figure 9 *Posidonia mimer* Oeberg. Smithian, Blind Fiord Formation, Ellesmere Island; 14201.

Figures 10a, b *Prosphingites spathi* Frebold. Smithian, Blind Fiord Formation, Ellesmere Island; 14085.

Figures 11a, b *Juvenites needhami* Tozer. Smithian, Toad Formation, northeastern British Columbia; 14290.

Figures 12a, b *Anakashmirites borealis* Tozer. Smithian, Blind Fiord Formation, Ellesmere Island; 14073.

Figures 13a, b *Euflemingites romunderi* Tozer. Smithian, Blind Fiord Formation, Ellesmere Island; 14050.

Figures 14a, b *Wasatchites tardus* (McLearn). Smithian, Toad Formation, northeastern British Columbia; 9470.

Figures 15a, b *Xenoceltites subevolutus* Spath. Smithian, Toad Formation, northeastern British Columbia; 14303.

Figure 16 *"Pseudomonotis" occidentalis kindlei* McLearn. Smithian, Toad Formation, northeastern British Columbia; 9478, 9598.

Figure 17 *"Pseudomonotis" occidentalis* Whiteaves. Smithian, Toad Formation, northeastern British Columbia; 4728.

Figures 18 *Posidonia aranea* Tozer. Spathian, Toad Formation, northeastern British Columbia; 14230.

Figures 19a, b *Olenikites canadensis* Tozer. Spathian, Blaa Mountain Formation, Ellesmere Island; 14094.

(Plates XVI, XVII, and XVIII—all figures are natural size unless otherwise stated)

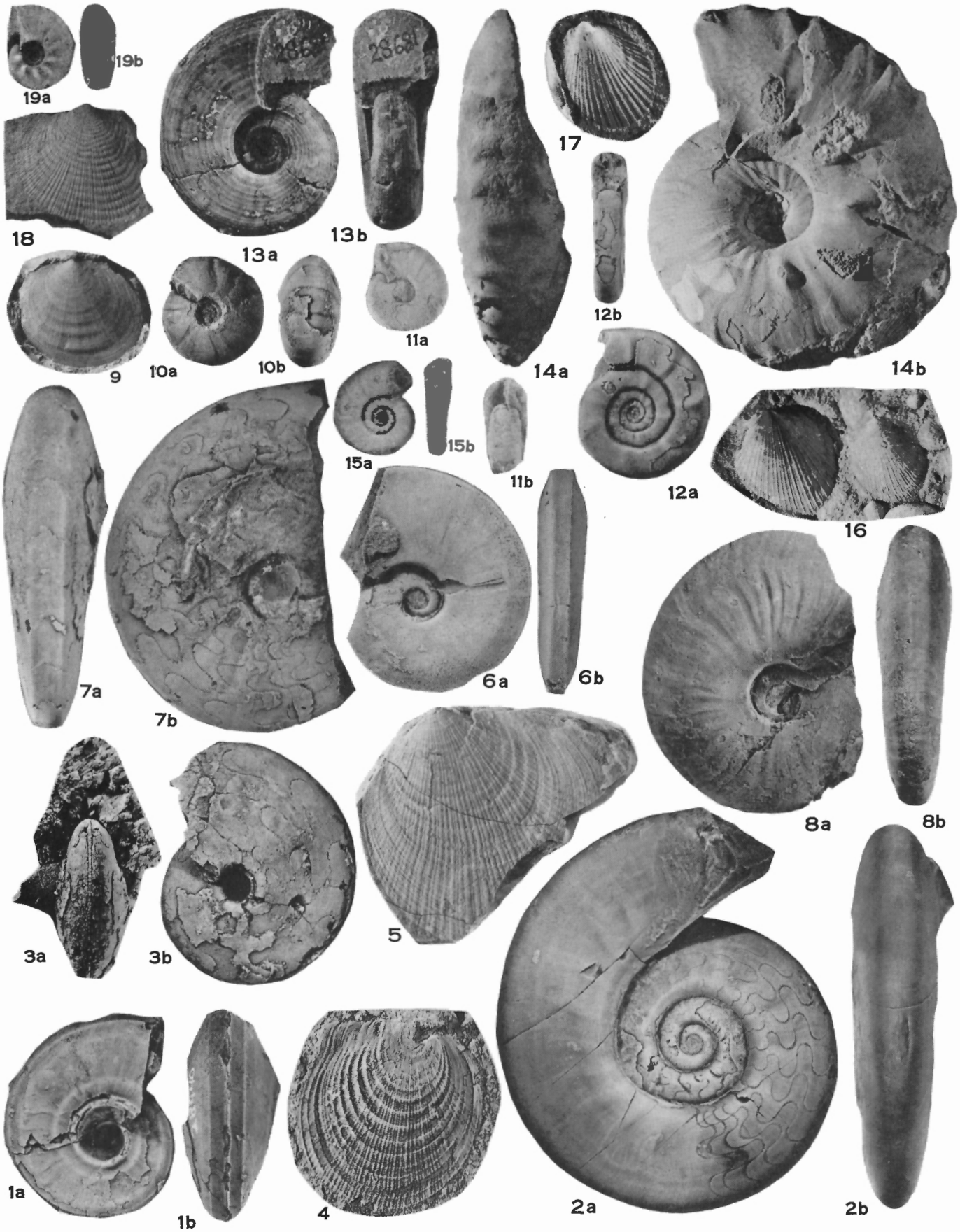
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PLATE XVII. Middle Triassic Ammonoids and Bivalves

Figures 1a, b *Lenotropites caurus* (McLearn), Lower Anisian, Toad Formation, northeastern British Columbia; 9585.

Figures 2a-c *Ptychites* cf. *trochlaeformis* (Lindström). Upper Anisian, Schei Point Formation, Exmouth Island; 14186, 14187.

Figures 3a, b *Anagymnotoceras helle* (McLearn). Middle Anisian, Toad Formation, northeastern British Columbia; 9592.

Figures 4a, b *Parapopanoceras selwyni* McLearn. Middle Anisian, Toad Formation, northeastern British Columbia; 9580, 9579.

Figures 5a, b *Longobardites nevadanus* Hyatt and Smith. Anisian, Toad Formation, northeastern British Columbia; 6450.

Figures 6a, b *Gymnotoceras deleeni* McLearn. Upper Anisian, Toad Formation, northeastern British Columbia; 21707.

Figure 7 *Daonella frami* Kittl. Lower Ladinian, Blaa Mountain Formation, Ellesmere Island; 14205.

Figures 8a, b 9a, b *Progonoceratites poseidon* Tozer. Lower Ladinian, Liard Formation, northeastern British Columbia; 18887, 18888.

Figures 10a, b *Protrachyceras sikanianum* McLearn. Upper Ladinian, Liard Formation, northeastern British Columbia; 9044.

Figures 11a, b *Meginoceras meginae* McLearn. Upper Ladinian, Liard Formation, northeastern British Columbia; 8811, 9042.

Figures 12a, b *Lobites pacianus* McLearn. Upper Ladinian, Liard Formation, northeastern British Columbia; 9525, 8789.

Figure 13 *Daonella nitanae* McLearn. Upper Ladinian, Liard Formation, northeastern British Columbia; 8773.

Figures 14a, b *Spiriferina abichi* Oppel. Ladinian, Liard Formation, northeastern British Columbia; 4733.

Figures 15a-c *"Terebratula" liardensis* Whiteaves. Ladinian, Liard Formation, northeastern British Columbia; 4734.

Figures 16a, b *Maclearnoceras maclearni* Tozer. Upper Ladinian, Liard Formation, northeastern British Columbia; 14296, 14298.

Figure 17 *Hoernesia woyoniana* McLearn. Upper Ladinian, "Grey Beds", northeastern British Columbia; 8768.

Figure 18 *Modiolus ahsisi* McLearn. Upper Ladinian, "Grey Beds", northeastern British Columbia; 8767.

Figure 19 *Daonella elegans* McLearn. Upper Ladinian, Liard Formation, northeastern British Columbia; 9537.

Figures 20a, b *Nathorstites mcconnelli* (Whiteaves). Upper Ladinian, Liard Formation, northeastern British Columbia; 21769.

Figures 21a, b *Daxatina canadensis* (Whiteaves). Upper Ladinian, Liard Formation, northeastern British Columbia; 4718.

Figures 22a, b *Paratrachyceras sutherlandi* McLearn. Upper Ladinian, Liard Formation, northeastern British Columbia; 18903.

PLATE XVIII. Upper Triassic Ammonoids, Bivalves, and Corals

Figures 1a, b *Sirenites nanseni* Tozer. Lower Karnian, Blaa Mountain Formation, Ellesmere Island; 14158.

Figure 2 *Mysidioptera poyana* (McLearn). Karnian, "Grey Beds", northeastern British Columbia; 8772.

Figure 3 *Halobia zitteli* Lindstrom. Karnian, Schei Point Formation, Ellesmere Island; 14208.

Figures 4a, b *Discotropites* sp. Upper Karnian, Pardonet Formation, northeastern British Columbia; 14239.

Figures 5a, b *Tropites johnsoni* Smith. Upper Karnian, northwestern British Columbia; 14243.

Figures 6a, b *Homerites semiglobosus* (Hauer). Upper Karnian, Pardonet Formation, northeastern British Columbia; 14235.

Figures 7a, b *Juvavites* cf. *knowltoni* Smith. Upper Karnian, Pardonet Formation, northeastern British Columbia; 14244.

Figures 8a, b *Mojsisovicsites kerri* (McLearn). Lower Norian, Pardonet Formation, northeastern British Columbia; 9048, 18911.

Figures 9a, b *Malayites dawsoni* McLearn. Lower Norian, Pardonet Formation, northeastern British Columbia; 8836, 12604.

Figures 10a, b *Himavatites columbianus* McLearn. Middle Norian, Pardonet Formation, northeastern British Columbia; 9265.

Figure 11 *Meleagrinella antiqua* Tozer. Probably Lower Norian, Heiberg Formation, Axel Heiberg Island; 14212.

Figure 12 *Gryphaea chakii* McLearn. Norian, Pardonet Formation, northeastern British Columbia; 8770.

Figure 13 *Elysastraea profunda* (Reuss). x1.5, Upper Norian, Sutton Formation, Vancouver Island; 7807.

Figure 14 *Thecosmilia dawsoni* (Clapp and Shimer). x1.5, Upper Norian, Sutton Formation, Vancouver Island; 7806.

Figures 15a-c *Spondylospira lewesensis* (Lees). Upper Norian, Lewes River Group, Yukon; 14259, 14260, 14261.

Figures 16a, b *"Myophoria" suttonensis* Clapp and Shimer. Upper Norian, Sutton Formation, Vancouver Island; 14256.

Figure 17 *Paracochloceras suessi* Mojsisovics. x2, Upper Norian, Bonanza Group, Vancouver Island; 17016.

Figures 18a, b *Rhabdoceras suessi* Hauer. Upper Norian, Lewes River Group, Yukon; 14265, 14264.

Figures 19a, b *Cassianella lingulata* Gabb. Upper Norian, Bonanza Group, Vancouver Island; 14249.

Figure 20 *Monotis subcircularis* Gabb. Upper Norian, southwestern Yukon; 14265.

Figures 21a-c *Choristoceras marshi,* Hauer. Rhaetian, Tyaughton Group, southern British Columbia; 18912, 18913.

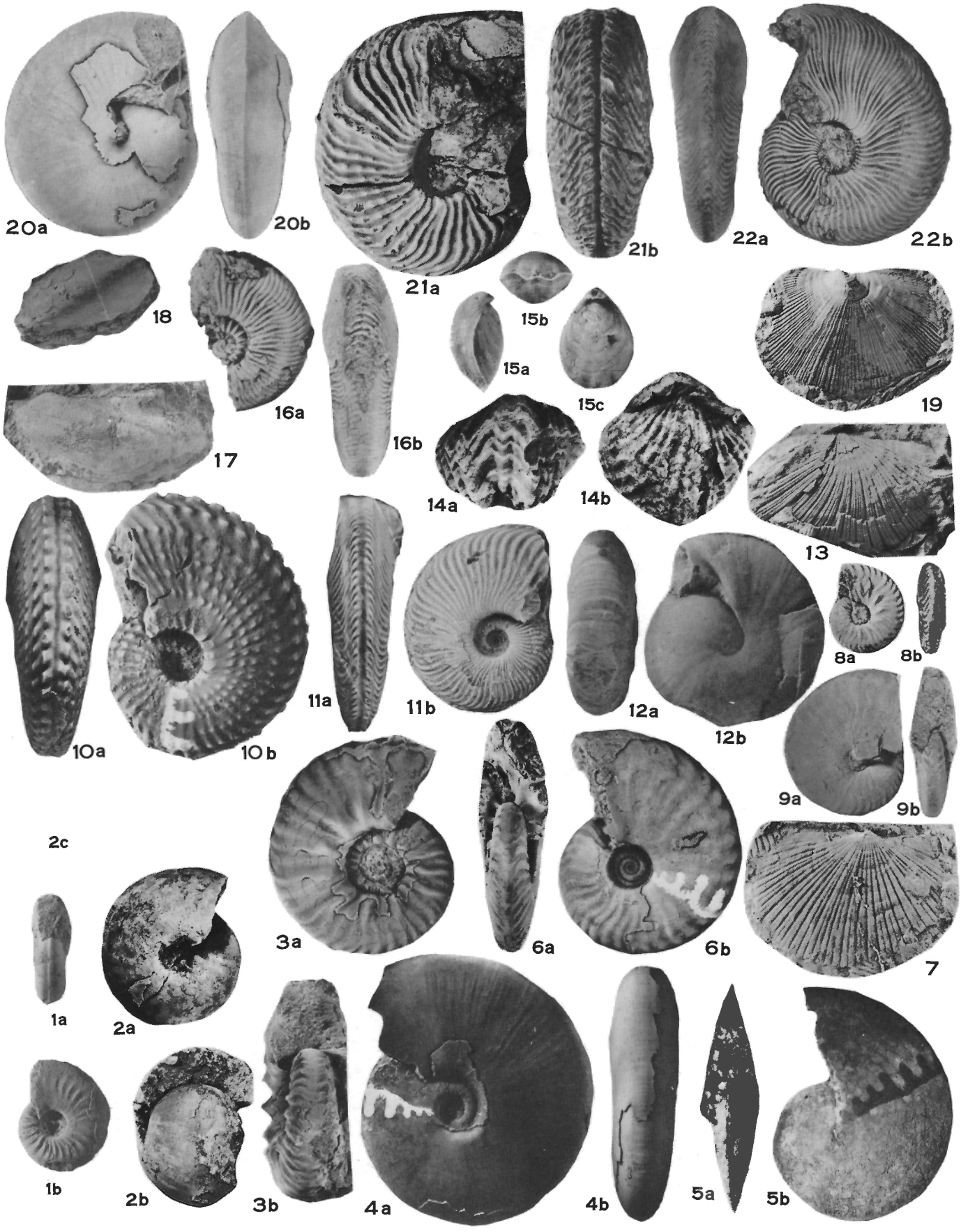
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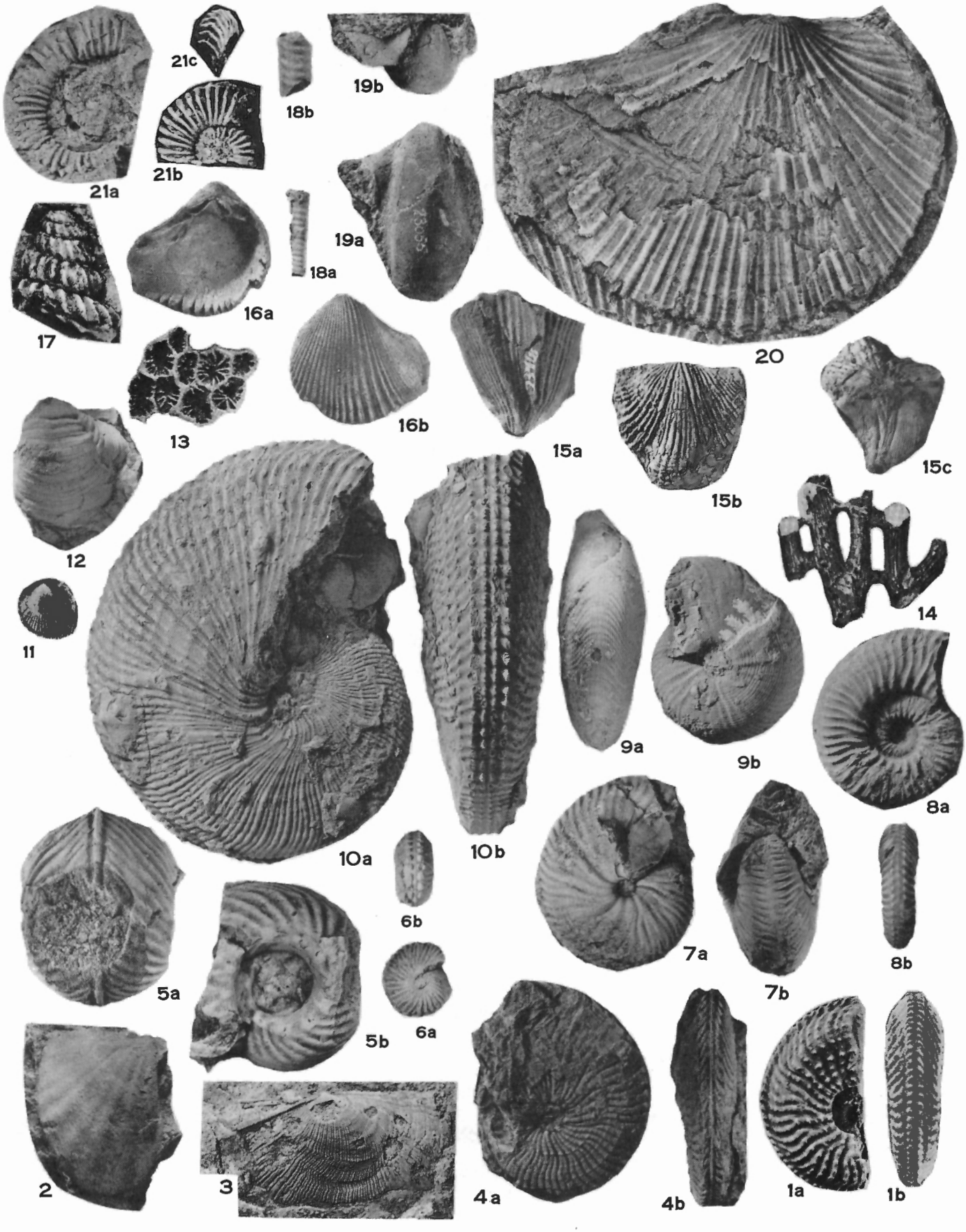
PLATE XVIII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MARINE JURASSIC FAUNAS

by Hans Frebold

Distribution and General Features

Faunas of Jurassic age are present in arctic and western Canada, which regions represent two distinctive zoogeographical provinces (Table XI-7). The arctic area includes the northern part of the Canadian arctic islands, northwestern parts of the Northwest Territories, and northern and central Yukon. The western area includes the subsurface of the southern part of the Prairie Provinces, Manitoba, Saskatchewan, and eastern Alberta, the Rocky Mountains and Foothills, various parts of central and western British Columbia, including Vancouver Island and the Queen Charlotte Islands.

The subdivision of the Jurassic into zones and stages is based on ammonites, and in the Upper Jurassic also on representatives of the pelecypod genus *Buchia*. The ammonites also establish the zoogeographical provinces. Of other invertebrates belemnites, pelecypods, gastropods, corals, a few brachiopods, echinoderms, arthropods, and microfaunas occur. There are a few marine vertebrate remains. Of the invertebrate macrofossils, pelecypods and gastropods have also provided valuable guide fossils. Microfaunas have particular importance in the subsurface where macrofossils are rarely available.

A great number of faunal gaps, often coincidental with absence of sediments, are characteristic of the Jurassic in Canada. This is evident when comparing the Jurassic faunas of western Canada with those of northwestern Europe or other parts of the world. A considerable number of the ammonites characteristic of certain zones are missing, sometimes amounting to the greater part of a stage.

Zoogeography

Lower Jurassic ammonite faunas in Canada do not show any zoogeographical differences. Their character is not basically different from that of those of other parts of the world, and most show close affinities, even at the specific level, to the Lower Jurassic ammonites of northwestern Europe. As far as the Hettangian ammonites of British Columbia are concerned, similarities to the northeastern Alps exist mainly on generic level. Thus, the well-known cosmopolitan character of the Lower Jurassic ammonite faunas is also clearly evident in both western and arctic Canada. Lower Jurassic pelecypods seem to be evenly distributed in these areas, and even show a considerable independence of facies. Corals are, however, restricted to southwestern British Columbia, where they are found in rocks closely associated with volcanic and greywacke sequences.

The same zoogeographical conditions exist in the lower Bajocian, earliest Middle Jurassic. In the middle Bajocian, the Bathonian, lower and middle Callovian — that is throughout most of the Middle Jurassic, as the presence of upper Callovian is not recognized in Canada —clear zoogeographical differences in the ammonite faunas are apparent, particularly between arctic and western Canada. The arctic province is characterized, in ascending order, by the genera *Arkelloceras* (probably middle Bajocian), *Cranocephalites* (upper Bajocian and lower Bathonian), *Arctocephalites* (mainly middle Bathonian), *Arcticoceras* (upper Bathonian), and various arctic species of the genus *Cadoceras* (upper Bathonian and lower Callovian). With few exceptions these faunas are more or less completely represented in East Greenland, Spitsbergen, Frans Josef Land, Novaya Zemlya, and northern Siberia but are still unknown in western Canada. In the western province, the middle Bajocian is characterized uniformly by such genera as *Stephanoceras, Stemmatoceras, Teloceras, Chondroceras,* and others, which are all absent in the arctic province. No typical arctic Bathonian ammonites were found in the western province. The lower Callovian ammonite genera of western Canada as *Paracephalites, Warrenoceras, Paracadoceras, Lilloettia, Imlayoceras,* and *Kepplerites* are absent in arctic Canada, but also within the western province some differences in the lower Callovian ammonite fauna seem to exist. The genera *Paracephalites, Warrenoceras,* and *Imlayoceras* of the Fernie Group of the Rocky Mountains and Foothills have hitherto not been found in any area of British Columbia whereas *Lilloettia* and *Paracadoceras* found in British Columbia do not appear in the Fernie Group. On the other hand, close faunal and stratigraphic relationships exist between the Callovian ammonite faunas of the Fernie Group and those of the United States western Interior. A fairly rich *Cadoceras* fauna different from that of other parts of western Canada and arctic Canada is present on Vancouver Island.

Beginning with the lower Oxfordian of the Upper Jurassic, the differences between the arctic and western faunal provinces of Canada are no longer recognizable, at least on the basis of present knowledge. The ammonite genera *Cardioceras* and *Amoeboceras* of the lower and upper Oxfordian and lower Kimmeridgian respectively are uniformly distributed in both arctic and western Canada and the same applies to the upper Oxfordian–lower Kimmeridgian pelecypod guide species *Buchia concentrica.* Also in younger Upper Jurassic beds the various *Buchia* species appear widely distributed in both northern and western Canada. However, as ammonites of these beds are extremely rare, it cannot yet be stated with confidence that no different faunal provinces existed in Canada during the late Upper Jurassic.

With the exception of the corals that seem to be restricted to regions of British Columbia, no other groups

TABLE XI-7 | *Distribution and succession of Jurassic guide fossils*

EPOCH	STAGES	ARCTIC ISLANDS	RICHARDSON AND BRITISH MOUNTAINS	CENTRAL YUKON TERRITORY	SOUTHERN YUKON TERRITORY	QUEEN CHARLOTTE ISLANDS	VANCOUVER ISLAND	WESTERN AND INTERIOR PARTS OF BRITISH COLUMBIA	ROCKY MOUNTAINS	SOUTHERN PLAINS
UPPER JURASSIC	Tithonian: Purbeckian s. st. (Aquilonian, upper Volgian)	*Buchia fischeriana*	*Buchia fischeriana*	Not known	*Buchia fischeriana*	Absent		*B. terebratuloides*	*Buchia* cf. *fischeriana*	Present
		Buchia piochii	*Buchia piochii*					*B. fischeriana*	*Titanites occidentalis*, *Buchia piochii*	
	Tithonian: Portlandian or lower Volgian	*Buchia mosquensis* *Buchia* spp. (?Stratigraphic position doubtful; *Dorsoplanites*)	*Buchia mosquensis* *Buchia* spp.				*Buchia* cf. *blanfordiana*	*Buchia* cf. *blanfordiana*	*Buchia mosquensis*	
	Tithonian: Upper Kimmeridgian						*Buchia mosquensis*	*Buchia mosquensis*		
	Middle Kimmeridgian									
	Lower Kimmeridgian Upper Oxfordian	*Amoeboceras* sp., *Buchia concentrica*	*Amoeboceras* sp., *Buchia concentrica*				*Buchia concentrica*	*Amoeboceras* spp., *Buchia concentrica*	*Amoeboceras* spp., *Buchia concentrica*	
	Lower Oxfordian	*Cardioceras* aff. *mirum*	*Cardioceras* spp.	*Cardioceras* spp.	Absent		*Cardioceras* spp.	*Cardioceras* spp.	*Cardioceras* spp., *Goliathiceras* cf. *crassum*	
MIDDLE JURASSIC	Upper Callovian	Not Known	Not Known	Not Known			Not Known	Not Known	Not Known	Not Known
	Middle Callovian	Not Known	Not Known	Not Known		Not Known	*Cadoceras* spp.	*Lilloettia lilloettensis* *Stenocadoceras* spp. *Pseudocadoceras* spp.	Not Known	Not Known
	Lower Callovian	*Cadoceras septentrionale*	*Cadoceras septentrionale*			*Kepplerites* spp.	Not Known	*Kepplerites* spp. *Cadoceras* sp., *Paracadoceras* sp. etc.	*Imlayoceras miettense*	*K. mclearni*
		Cadoceras bodylevskyi							*Kepplerites mclearni*	
		Kepplerites not known	*Kepplerites* not known						*Kepplerites* aff. *K. tychonis*	
		Cadoceras spp.	*Cadoceras* spp.	*Cadoceras?* sp.					*Warrenoceras henryi*	
			Cadoceras spp.						*Paracephalites glabrescens P. hashimotoi*	
	Bathonian	*Arcticoceras ishmae*	*Arcticoceras kochi*			Not Known		Not Known	Not Known	
		Arctocephalites spp.	*Cadoceras crassum?*							
			Arctocephalites elegans	*Arctocephalites?* sp.						
		Cranoc. vulgaris	Not Known							
	Upper Bajocian	Not Known	*Cranocephalites borealis*							
	Middle Bajocian	*Arkelloceras* spp. *Inoceramus lucifer*	*Arkelloceras elegans*, *Inoceramus lucifer*			*Stephanoceras* spp. *Chondroceras* spp.		*Stephanoceras* spp., *Chondroceras* spp.	*Stephanoceras* spp., *Stemmatoceras* spp., *Teloceras* spp., *Chondroceras* spp.	Present
	Lower Bajocian	*Leioceras opalinum*, *Pseudolioceras m'clintocki* *O. jacksoni*	*Erycites* cf. *howelli* and *Pseudolioceras* sp.		Possibly present in part	Not Known		*Tmetoceras regleyi*	Not Known	
LOWER JURASSIC	Toarcian	*Pseudolioceras compactile*, *Catacoeloceras spinatum*	*Pseudolioceras* sp.		*Grammoceras* aff. *G. fallaciosum*	*Harpoceras* spp. *Dactylioceras* spp.	*Harpoceras* spp.	*Grammoceras* *Phymatoceras* sp.	*Grammoceras boreale* *Phymatoceras* sp.	Absent
		Dactylioceras commune *Harpoceras* cf. *exaratum*	*Dactylioceras commune* *Harpoceras* sp.		?			*Dactylioceras* spp., *Harpoceras* cf. *exaratum*	*Dactylioceras* sp., *Harpoceras* cf. *exaratum*	
	Pliensbachian	Not Known	Not Known	Not Known	Not Known	Not Known		?	*Pleuroceras ?*	
			Amaltheus stokesi		*Amaltheus stokesi*, *Arieticeras* spp., *Leptaleoceras* sp.			*Amaltheus stokesi*, *Arieticeras* spp., *L. pseudoradians*	*Amaltheus* spp. *Amaltheus stokesi*	
			Not Known		*Prodactylioceras* sp.	Possibly present		*Becheiceras* cf. *bechei*		
					Probably absent	*Fanninoceras* spp. *Tropidoceras*	*Fanninoceras* *Acanthopleuroceras*	*Platypleuroceras* sp.	*Phricodoceras* sp.	
	Sinemurian	*Echioceras* s.l.	*Echioceras* s.l.			*Echioceras*	*Echioceras*	*Echioceras* sp.	*Eoderoceras* sp.	
		Gleviceras sp.	*Arctoasteroceras jeletzkyi*, *Oxynoticeras oxynotum*, *Gleviceras* sp.			*Asteroceras ?*	*Asteroceras* sp.	*Asteroceras* sp.		
					Arnioceras sp. *Paracoroniceras* cf. *gmündense*	*Arniotites* spp.	*Arniotites* spp.	*Arniotites* spp. *Coroniceras*	*Arietites* spp. *Arnioceras* spp.	
		Arietites s.l.	*Arietites* s.l.		Not Known					
	Hettangian	Not Known	Not Known		*Psiloceras* cf. *erugatus*	Not Known	Not Known	*Psil. (Curv.) columbiae* *Psiloceras canadense* *Psil.* aff. *planorbis*	Not Known	

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seem to be dependent on facies. This may be due partly to the fact that such groups have not yet been studied to the same extent as the ammonites.

Succession of Faunas

The succession of ammonites and buchias is shown in Table XI-7; the latter is based on studies mainly by J. A. Jeletzky. The succession of ammonites in the Lower Jurassic and the lower Bajocian established in various areas of Canada is in agreement with standard sequences in northwestern Europe, and the Hettangian is similar to a certain degree to that of the northeastern Alps. Wherever in Canada a more continuing section could be studied it follows the European pattern and that of other parts of the world, and accordingly the same succession of ammonite zones as in Europe can be applied to the Lower Jurassic of Canada. However, the local or regional incompleteness of the Canadian Lower Jurassic zonal sequence has to be emphasized. The absence of the Hettangian faunas so richly represented in parts of British Columbia in other areas of western and northern Canada is caused by non-deposition of beds of this age. Lower and upper Pliensbachian have now been found in western Canada and are considered to be primarily absent in other areas.

The succession of ammonite faunas and ammonite zones of the Canadian arctic in the upper Bajocian, Bathonian, and Callovian (Table XI-7) is almost the same as the succession in East Greenland established by Callomon (1959). In the middle Bajocian of western Canada, ammonite faunas follow again the northwest European pattern whereas the succession of ammonite faunas in the uppermost Bathonian and the lower Callovian of the Fernie Group in the Rocky Mountains and Foothills is almost completely the same as in the western Interior of the United States. The absence of upper Callovian ammonites in western Canada and the western Interior indicates a considerable gap in the sequence. In the Oxfordian and lower Kimmeridgian, the succession of ammonite fauna is the same as known from other parts of the world, but in the younger Upper Jurassic stages, ammonites are very rare and, if present, poorly preserved. Consequently, the subdivision of the beds concerned is based mainly on various *Buchia* species that show a zonal arrangement similar to that of other regions.

Selected References

Comprehensive lists of references, including publications by S. S. Buckman, C. H. Crickmay, H. Frebold, F. H. McLearn, P. S. Warren, and J. F. Whiteaves are given in:

Frebold, Hans
1964: Illustrations of Canadian fossils. Jurassic of western and arctic Canada; *Geol. Surv. Can.*, Paper 63–4.

Frebold, H., and Tipper, H. W.
1970: Status of the Jurassic in the Canadian Cordillera of British Columbia, Alberta, and southern Yukon; *Can. J. Earth Sci.*, vol. 7, pp. 1–22.

Other publications are:

Callomon, J. H.
1959: The ammonite zones of the Middle Jurassic beds of East Greenland; *Geol. Mag.*, vol. 96, No. 6.

Copeland, M. J.
1960: *Erymastacus bordenensis* a new Mesozoic decapod from the Canadian arctic; *Geol. Surv. Can.*, Bull. 60.

Jeletzky, J. A.
1964: Illustrations of Canadian fossils. Lower Cretaceous marine index fossils of the sedimentary Basins of western and arctic Canada; *Geol. Surv. Can.*, Paper 64-11.
1965: Late Upper Jurassic and early Lower Cretaceous fossil zones of the Canadian Western Cordillera, British Columbia; *Geol. Surv. Can.*, Bull. 103.

Plate XIX. Lower Jurassic Ammonites

Figure 1 *Eolytoceras tasekoi* Frebold. Hettangian; Taseko Lakes, British Columbia; 20059.

Figure 2 *Psiloceras* ex aff. *planorbis* (Sowerby). Hettangian; Taseko Lakes, British Columbia; 20053.

Figures 3a, b *Psiloceras canadense* Frebold. Hettangian; Taseko Lakes, British Columbia; 62504.

Figure 4 *Discamphiceras*? *tipperi* Frebold. Hettangian; Taseko Lakes, British Columbia; 19926.

Figure 5 *Psiloceras* (*Curviceras*) *columbiae* Frebold. Hettangian; Taseko Lakes, British Columbia; 19921.

Figure 6 *Psiloceras* (*Curviceras*) *columbiae* Frebold. Hettangian; Taseko Lakes, British Columbia.

Figure 7 *Charmasseiceras marmoreum* (Oppel). Hettangian; Taseko Lakes, British Columbia; 20056.

Figures 8a, b *Charmasseiceras marmoreum* (Oppel). Hettangian; Taseko Lakes, British Columbia; 20052.

Figure 9 *Paracaloceras multicostatum* Frebold. Hettangian; Taseko Lakes, British Columbia; 19939.

Figure 10 *Paracoroniceras* cf. *gmündense* (Oppel). Laberge Group, Sinemurian; Atlin, British Columbia; 15969.

Figure 11 *Oxynoticeras oxynotum* (Quenstedt). Upper Sinemurian; Aklavik Range, N.W.T.; 14631.

Figure 12 *Arctoasteroceras jeletzkyi* Frebold. Upper Sinemurian; Aklavik Range, N.W.T.; 14623.

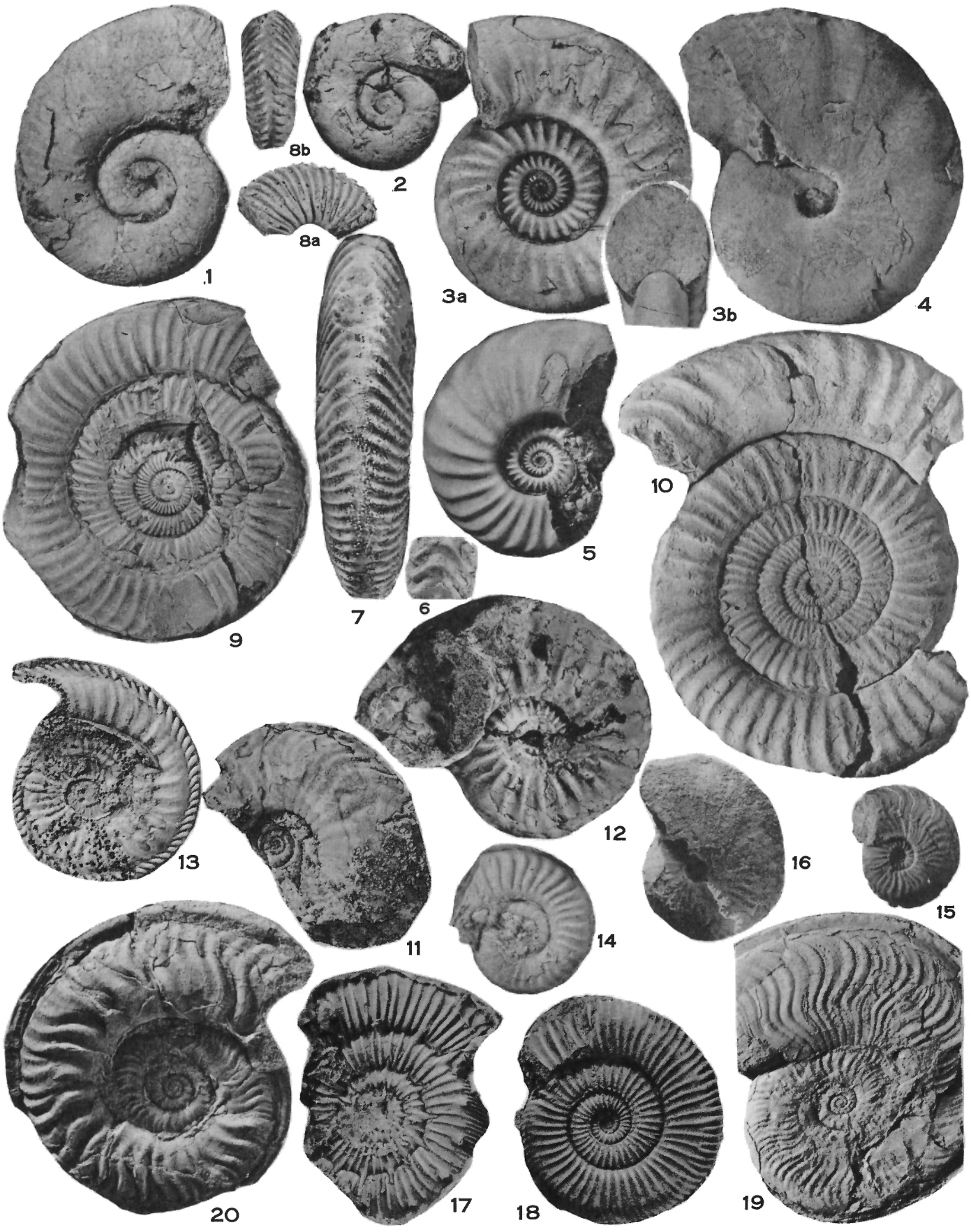
Figure 13 *Amaltheus stokesi* (Sowerby). Upper Pliensbachian; southern Yukon; 15974.

Figure 14 *Arieticeras* cf. *A. algovianum* (Oppel). Upper Pliensbachian; Telegraph Creek, British Columbia; 15989.

Figure 15 *Fanninoceras kunae* var. *latum* McLearn. Maude Formation, Pliensbachian: Queen Charlotte Islands; 9058.

Figure 16 *Fanninoceras fannini* McLearn. Maude Formation, Pliensbachian; Queen Charlotte Islands; 6493.

(Plates XIX–XXII—all figures natural size)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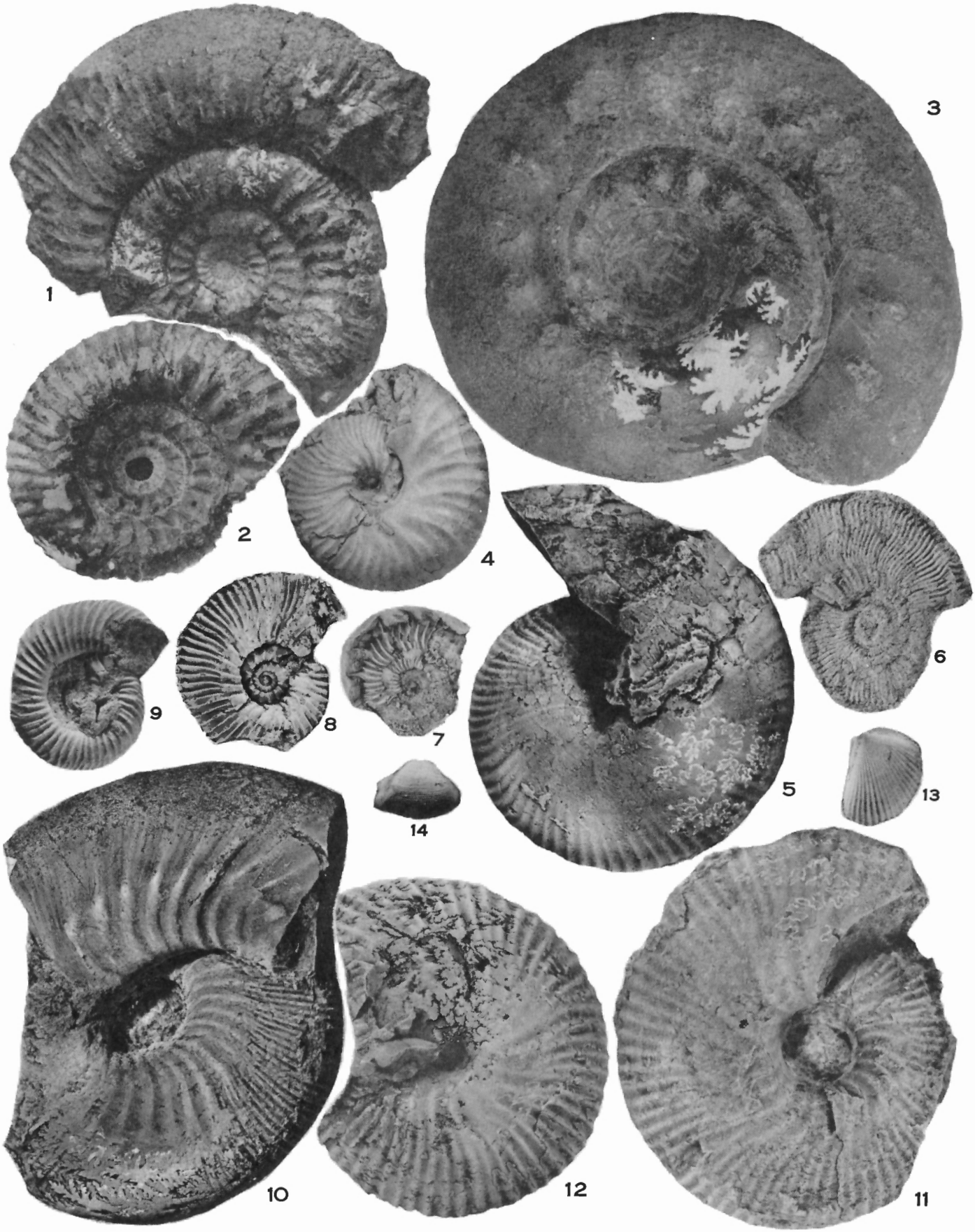
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Figure 17 *Catacoeloceras spinatum* (Frebold). Wilkie Point Formation, Toarcian; Prince Patrick Island; 13361.

Figure 18 *Dactylioceras commune* (Sowerby). Wilkie Point Formation, Toarcian; Prince Patrick Island; 13355.

Figure 19 *Harpoceras* cf. *H. exaratum* (Young and Bird). Toarcian; Tulsequah, British Columbia; 16003.

Figure 20 *Grammoceras boreale* (Whiteaves). Laberge Group, Toarcian; southern Yukon; 9703.

PLATE XX. Middle Jurassic Ammonites and Pelecypods

Figure 1 *Stephanoceras yakounense* McLearn. Yakoun Formation, middle Bajocian; Queen Charlotte Islands; 9057.

Figure 2 *Normannites canadensis* McLearn. Yakoun Formation, middle Bajocian; Queen Charlotte Islands; 9019.

Figure 3 *Zemistephanus funteri* McLearn. Yakoun Formation, middle Bajocian; Queen Charlotte Islands; 9007.

Figure 4 *Chondroceras allani* (McLearn). Middle Bajocian; Tulsequah, British Columbia; 16024.

Figure 5 *Warrenoceras rierdonense* (Imlay). Fernie Group, lower Callovian; Alberta; 14693.

Figure 6 *Kosmoceras* (*Gulielmiceras*) *knechteli* Imlay. Lower Callovian; Esterhazy Shaft, Saskatchewan; 14696.

Figure 7 *Torricellites? spinosum* Frebold. Fernie Group, lower Callovian; Alberta; 14710.

Figure 8 *Kepplerites newcombii* (Whiteaves). Yakoun Formation, lower Callovian; Queen Charlotte Islands; 5990.

Figure 9 *Cobbanites* sp. indet. aff. *engleri* Frebold. Fernie Group, lower Callovian; Alberta; 12908.

Figure 10 *Kepplerites mcevoyi* (McLearn). Fernie Group, lower Callovian; Alberta; 5018.

Figure 11 *Paracephalites glabrescens* Buckman. Fernie Group, upper Bathonian or lower Callovian; Alberta; 14705.

Figure 12 *Imlayoceras miettense* Frebold. Fernie Group, lower Callovian; Alberta; 14707.

Figure 13 *Lima albertensis* McLearn. Fernie Group, upper Bathonian or lower Callovian; Alberta; 6075.

Figure 14 *Corbula munda* McLearn. Enlarged x3, Fernie Group, upper Bathonian or lower Callovian; Alberta; 6092.

PLATE XXI. Middle Jurassic Ammonites and Pelecypods

Figure 1 *Pseudolioceras m'clintocki* (Haughton). Wilkie Point Formation, lower Bajocian; Mackenzie King Island; 14658.

Figure 2 *Leioceras opalinum* (Reinecke). Wilkie Point Formation, lower Bajocian; Prince Patrick Island; 13379.

Figure 3 *Oxytoma jacksoni* (Pompeckj). Wilkie Point Formation; lower Bajocian; Prince Patrick Island; 13390.

Figures 4a-c *Arkelloceras tozeri* Frebold. Wilkie Point Formation, probably middle Bajocian; Prince Patrick Island; 13404.

Figure 5 *Cranocephalites borealis* (Spath). Bug Creek Formation, upper Bajocian; Aklavik Range, N.W.T.; 15103.

Figure 6 *Cranocephalites vulgaris* Spath. Wilkie Point Formation, lower Bathonian; Prince Patrick Island; 13398.

Figure 7 *Arctocephalites elegans* Spath. Bathonian; Richardson Mountains; 15108.

Figure 8 *Arcticoceras kochi* Spath. Bathonian; Porcupine River, Yukon; 15117.

Figure 9 *Arcticoceras ishmae* (Keyserling). Wilkie Point Formation, Bathonian; Prince Patrick Island; 15120.

Figure 10 *Cadoceras septentrionale* Frebold. Savik Formation, Callovian; Axel Heiberg Island; 17657.

Figure 11 *Cadoceras* cf. *falsum* Voronets. Savik Formation, Callovian; Axel Heiberg Island; 17659.

Figure 12 *Cadoceras canadense* Frebold. Callovian; Aklavik Range, N.W.T.; 17664.

PLATE XXII. Upper and Middle Jurassic Ammonites and Pelecypods

Figure 1 *Titanites occidentalis* Frebold. Lower Kootenay Sandstone, Portlandian; Fernie, British Columbia.

Figure 2 *Buchia* cf. *blanfordiana* (Stoliczka). Uppermost Portlandian?; Vancouver Island; 16584.

Figures 3a, b *Buchia fischeri* (d'Orbigny). Aquilonian; Aklavik Range, N.W.T.; 17991.

Figure 4 *Buchia piochii* (Gabb) var. *mniovnikensis* (Pavlow). Mould Bay Formation, upper Portlandian; Prince Patrick Island; 17120.

Figure 5 *Buchia mosquensis* (von Buch) sensu lato. Middle Kimmeridgian to lower Portlandian; Dave Lord Ridge, Yukon; 17990.

Figure 6 *Buchia mosquensis* (von Buch) sensu lato. Middle Kimmeridgian to lower Portlandian; Richardson Mountains, Yukon; 17989.

Figures 7a, b *Buchia concentrica* (Sowerby) var. *erringtoni* (Gabb). Tyaughton Lake, British Columbia; 17012.

Figure 8 *Pleuromya postculminata* McLearn. Fernie Group, upper Bathonian or lower Callovian; Alberta; 6089.

Figure 9 *Oxytoma blairmorensis* McLearn. Fernie Group, lower Callovian; Alberta; 6050.

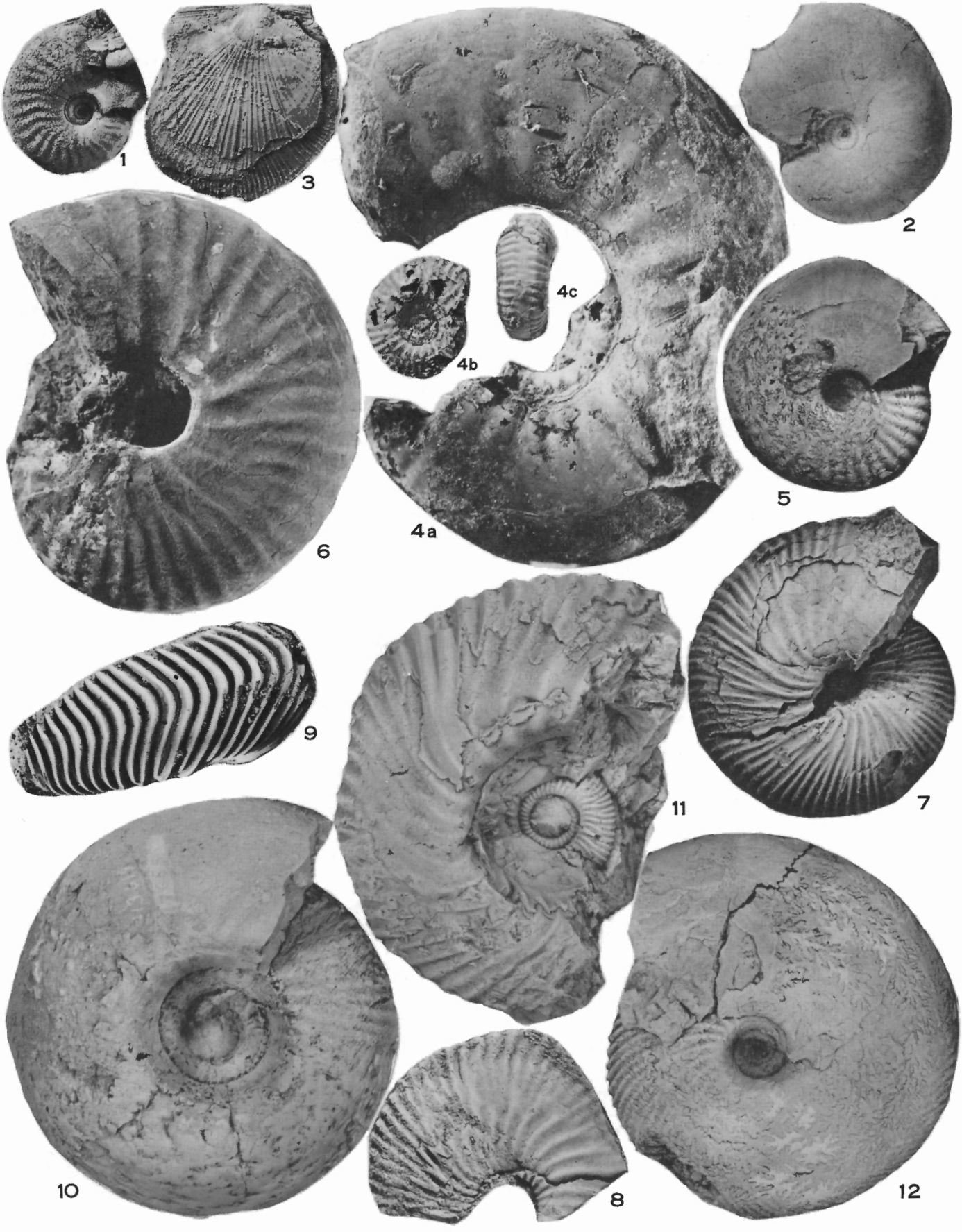
Figure 10 *Sonninia gracilis* (Whiteaves). Fernie Group, middle Bajocian; Alberta; 4809.

Figure 11 *Inoceramus lucifer* Eichwald. Wilkie Point Formation, probably middle Bajocian; Prince Patrick Island; 13417.

Figure 12 *Cardioceras canadense* Whiteaves. Fernie Group, Oxfordian; British Columbia; 7437.

Figure 13 *Cardioceras* (*Scarburgiceras*) *alphacordatum* Spath. Fernie Group, Oxfordian; Alberta; 13892.

Figure 14 *Amoeboceras* sp. indet. Mould Bay Formation; upper Oxfordian or lower Kimmeridgian; Mackenzie King Island; 15131.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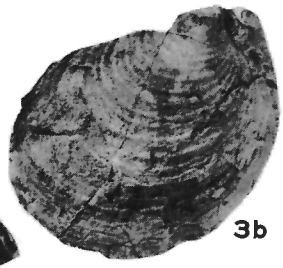

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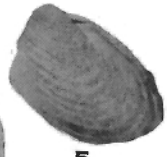

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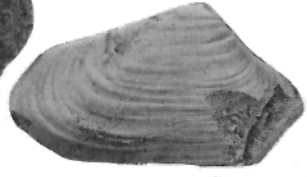

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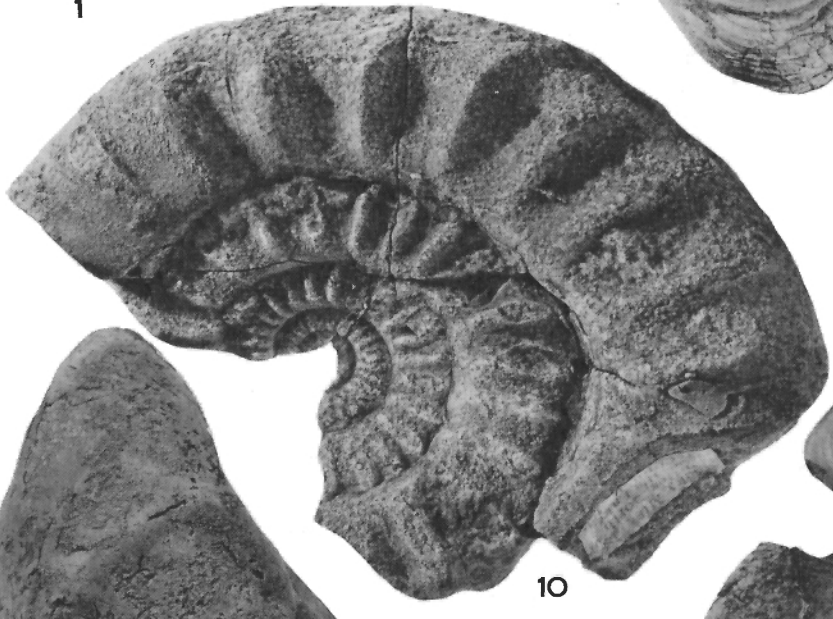

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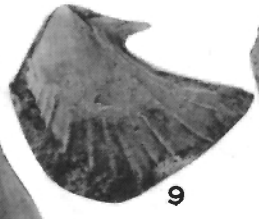

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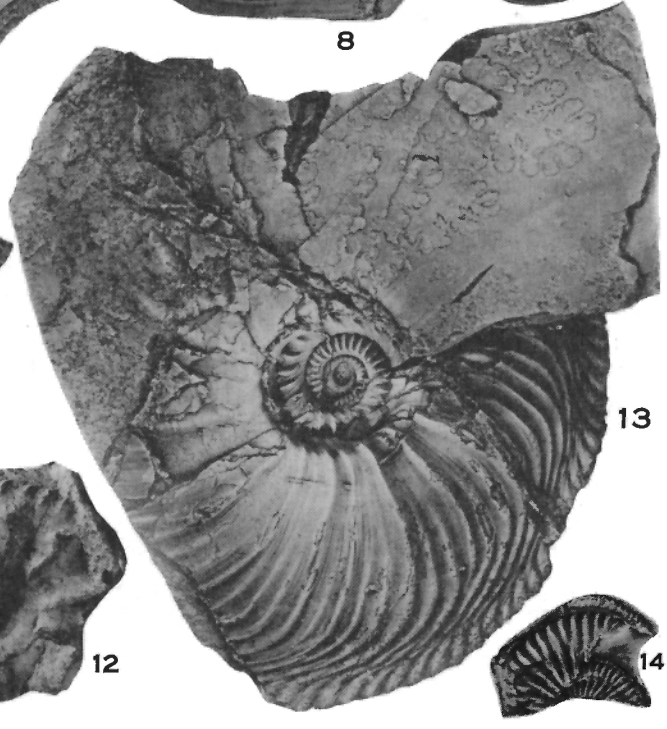

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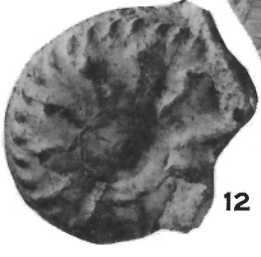

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CRETACEOUS MACROFAUNAS

by J. A. Jeletzky

Considerable parts of the Cretaceous System are present throughout much of western and arctic Canada, the faunal successions of which are shown in Table XI-8. None of these faunal successions, however, is complete, and the missing standard zones and stages vary greatly from one region to another.

Only a few groups of macrofossils are practically useful for subdivision of the Canadian Cretaceous rocks. As in some other systems, ammonites are by far the most useful index fossils but are rather unevenly distributed. They may be extremely rare or absent in the Neocomian (Berriasian to Aptian) part of the system, and there less reliable index fossils such as species of *Inoceramus* and *Buchia* (also known as *Aucella*) have to be used. These two pelecypod genera occur in abundance and variety in most marine facies. The remaining pelecypods, including the Trigoniidae, appear to be far too long-ranging or facies-bound to be useful. Although valuable for intercontinental correlation, crinoids and belemnites are too rare and restricted in occurrence for regional or interregional zonal indices. The sponges, echinoderms, corals, brachiopods, gastropods, and crustaceans are very rare and little understood.

In non-marine rocks, reptiles, especially dinosaurs, have been found to be the best index fossils available, but they are rather unevenly distributed and generally rare. More common, but often less reliable, are non-marine mollusca, plants, and ostracoda.

Boundaries and Major Divisions

The lower boundary of the system is drawn at the base of the Berriasian stage which includes the *Berriasella grandis* zone. The upper boundary is placed at the top of the Maestrichtian stage where ammonites, belemnites, inocerami, dinosaurs, and other Mesozoic index fossils apparently disappear completely (Jeletzky, 1962; Jeletzky and Clemens, 1965). The two-fold subdivision of the Cretaceous System is followed. The boundary between the Lower and Upper Cretaceous is drawn between the Albian and Cenomanian stages as defined by Muller and Schenck (1943), Cobban and Reeside (1952, p. 104), and Jeletzky (1968). In the western Interior region this boundary is placed tentatively between *Neogastroplites mclearni* and *N. septimus* zones.

A close similarity of all boreal marine faunas of Canada with those of northern Eurasia permits general use of the international standard stages based on West European fossil zones and type localities. Furthermore, it is easy to use these stages in the non-marine sequences because of widespread intertonguing with marine rocks. A separate stage scale has recently been proposed by Russell (1964) for the non-marine Cretaceous of the western Interior region. These stages are used as regional zones.

The international standard stages can also be used in the North Pacific faunal realm. However, the North Pacific faunas of Canada have to be correlated with the mixed faunas of South America, India, and Madagascar before they can be correlated with the international standard stages and zones (Popenoe, *et al.*, 1960) and regional zones of other parts of Canada.

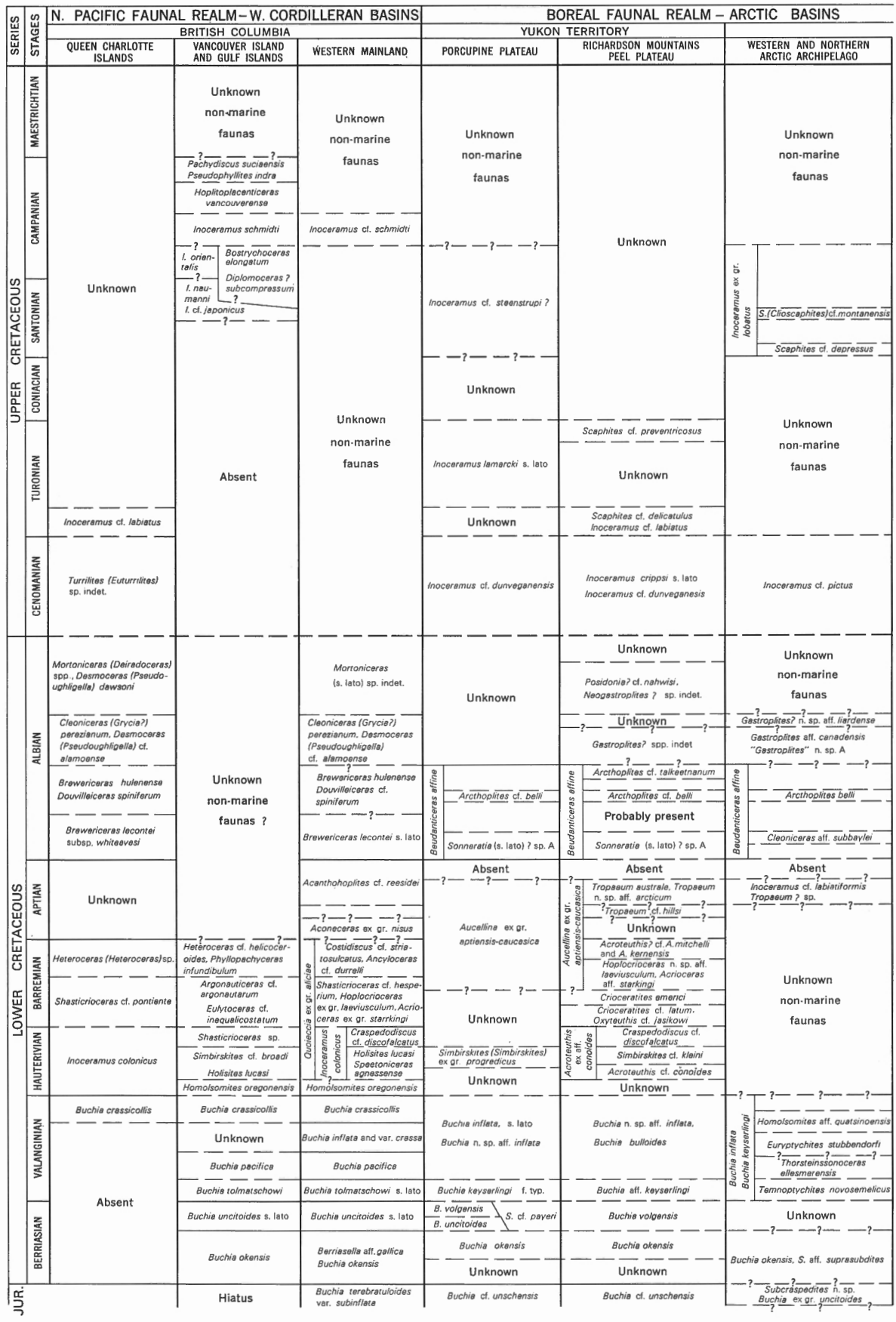
Marine Faunal Realms

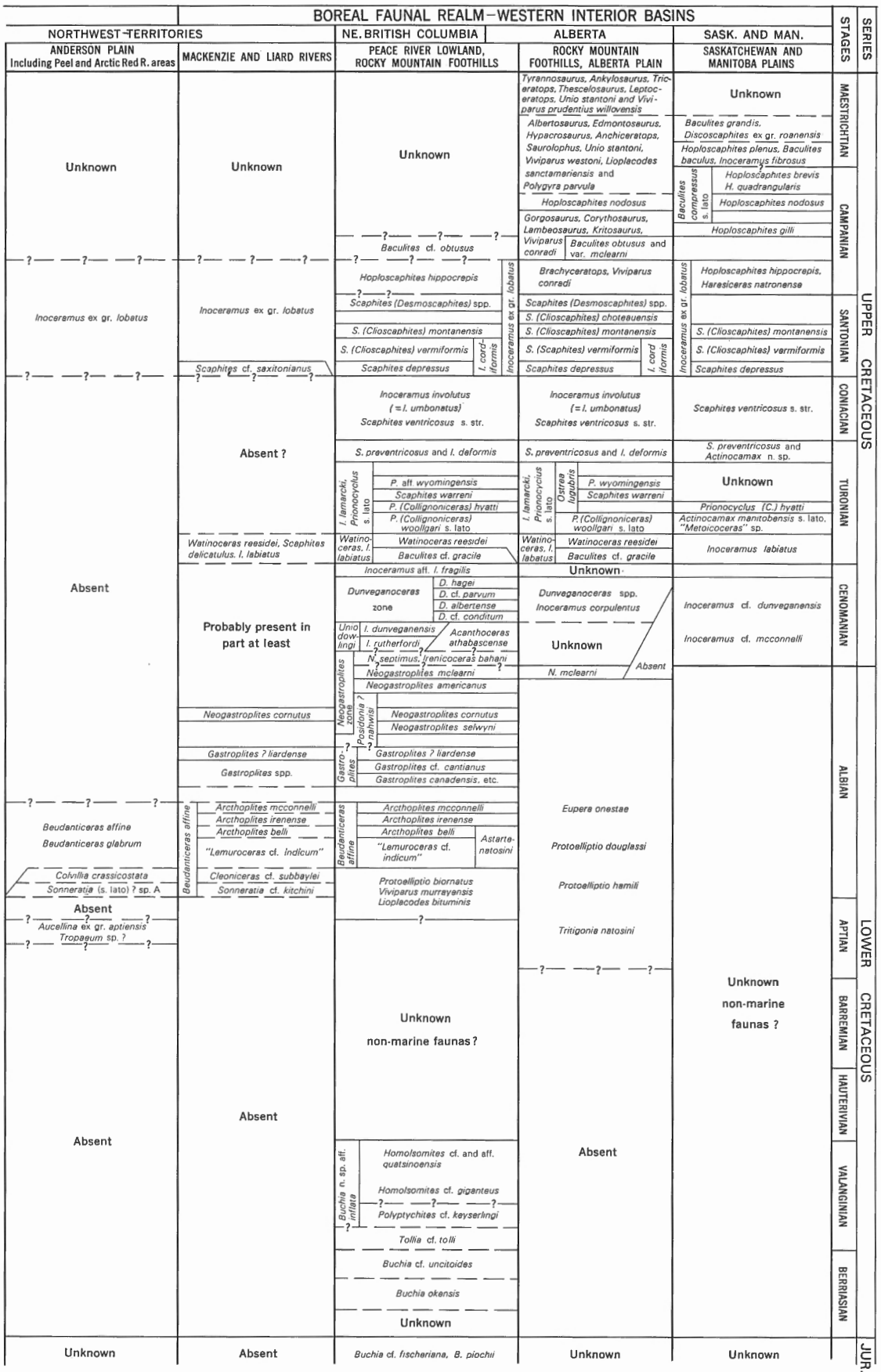
On the whole, the Cretaceous was a time of a progressively increasing differentiation of marine paleozoogeographical provinces or faunal realms. This differentiation was caused by a gradual but major redistribution of land and sea areas in western and arctic Canada during the Lower Cretaceous, followed in the late Upper Cretaceous by a gradual retreat of the seas beyond the present shoreline, completed apparently before the end of the Maestrichtian.

Two principal faunal realms existed throughout Cretaceous time. The North Pacific realm was dominated by marine faunas closely related to those of the Indo-Pacific region, Tethyan, and Andean Geosynclines (Jeletzky, 1965). Its faunas are, however, peculiar in that a number of endemic taxa are present, e.g., *Homolsomites, Shasticrioceras, Shastoceras, Buchia tolmatschowi, Buchia pacifica,* as well as others diagnostic of the Pacific slope of Siberia, Japan, Alaska, and California–Oregon region, e.g., *Inoceramus* ex gr. *naumanni–orientalis–schmidti, Inoceramus colonicus,* and yet others diagnostic of the boreal realm, e.g., *Tollia, Subcraspedites, Simbirskites, Hoplocrioceras, Tropaeum,* and a number of *Buchia* species. The North Pacific realm was restricted to the western Cordilleran region and to the Peace River region during Berriasian to Barremian time. Thereafter it was restricted to the western Cordillera region alone.

The Boreal faunal realm characterized by marine faunas almost identical to those of northern Eurasia was restricted to the Arctic Archipelago and parts of the mainland in northern Yukon during Berriasian to Aptian time. In Albian and Upper Cretaceous time it also included most of the Interior Plains and eastern Cordillera region. On the basis of the generally minor differences between the Albian and Upper Cretaceous marine faunas of the western Interior region and those of arctic Canada, two groups of marine basins may be distinguished in this realm.

TABLE XI-8 | *Succession and distribution of Upper and Lower Cretaceous faunas*




BOREAL FAUNAL REALM—WESTERN INTERIOR BASINS
NORTHWEST TERRITORIES
ANDERSON PLAIN Including Peel and Arctic Red R. areas
MACKENZIE AND LIARD RIVERS
NE. BRITISH COLUMBIA
PEACE RIVER LOWLAND, ROCKY MOUNTAIN FOOTHILLS
ALBERTA
ROCKY MOUNTAIN FOOTHILLS, ALBERTA PLAIN
SASK. AND MAN.
SASKATCHEWAN AND MANITOBA PLAINS
STAGES
SERIES
UPPER CRETACEOUS
LOWER CRETACEOUS
MAESTRICHTIAN
CAMPANIAN
SANTONIAN
CONIACIAN
TURONIAN
CENOMANIAN
ALBIAN
APTIAN
BARREMIAN
HAUTERIVIAN
VALANGINIAN
BERRIASIAN
JUR.
Unknown
Tyrannosaurus, Ankylosaurus, Triceratops, Thescelosaurus, Leptoceratops, Unio stantoni and Viviparus prudentius willovensis
Albertosaurus, Edmontosaurus, Hypacrosaurus, Anchiceratops, Saurolophus, Unio stantoni, Viviparus westoni, Lioplacodes sanctameriensis and Polygyra parvula
Hoploscaphites nodosus
Gorgosaurus, Corythosaurus, Lambeosaurus, Kritosaurus, Viviparus conradi
Baculites obtusus and var. mclearni
Baculites cf. obtusus
Baculites grandis, Discoscaphites ex gr. roanensis
Hoploscaphites plenus, Baculites baculus, Inoceramus fibrosus
Baculites compressus s. lato
Hoploscaphites brevis
H. quadrangularis
Hoploscaphites nodosus
Hoploscaphites gilli
Inoceramus ex gr. lobatus
Hoploscaphites hippocrepis
Haresiceras natronense
Brachyceratops, Viviparus conradi
Scaphites (Desmoscaphites) spp.
S. (Clioscaphites) choteauensis
S. (Clioscaphites) montanensis
S. (Scaphites) vermiformis
I. cordiformis
Scaphites depressus
Scaphites cf. saxitonianus
Inoceramus involutus (= I. umbonatus) Scaphites ventricosus s. str.
Scaphites ventricosus s. str.
Absent ?
S. preventricosus and I. deformis
S. preventricosus and Actinocamax n. sp.
I. lamarcki, Prionocyclus s. lato
P. aff. wyomingensis
Scaphites warreni
P. (Collignoniceras) hyatti
P. (Collignoniceras) woollgari s. lato
Ostrea lugubris
Prionocyclus (C.) hyatti
Actinocamax manitobensis s. lato, "Metoicoceras" sp.
Watinoceras reesidei, Scaphites delicatulus, I. labiatus
Watinoceras, I. labiatus
Watinoceras reesidei
Baculites cf. gracile
Inoceramus labiatus
Inoceramus aff. I. fragilis
Absent
Probably present in part at least
Dunveganoceras zone
D. hagei
D. cf. parvum
D. albertense
D. cf. conditum
Dunveganoceras spp.
Inoceramus corpulentus
Inoceramus cf. dunveganensis
Inoceramus cf. mcconnelli
Unio dowlingi
I. dunveganensis
I. rutherfordi
Acanthoceras athabascense
N. septimus, Irenicoceras bahani
Neogastroplites mclearni
N. mclearni
Neogastroplites americanus
Neogastroplites zone
Posidonia ? nahwisi
Neogastroplites cornutus
Neogastroplites selwyni
Gastroplites ? liardense
Gastroplites spp.
Gastroplites
Gastroplites cf. cantianus
Gastroplites canadensis, etc.
Beudanticeras affine
Beudanticeras glabrum
Arcthoplites mcconnelli
Arcthoplites irenense
Arcthoplites belli
"Lemuroceras cf. indicum"
Astartenatosini
Eupera onestae
Protoelliptio douglassi
Protoelliptio hamili
Colvillia crassicostata
Sonneratia (s. lato) ? sp. A
Cleoniceras cf. subbaylei
Sonneratia cf. kitchini
Protoelliptio biornatus
Viviparus murrayensis
Lioplacodes bituminis
Absent
Aucellina ex gr. aptiensis
Tropaeum sp. ?
Tritigonia natosini
Unknown non-marine faunas ?
Absent
Homolsomites cf. and aff. quatsinoensis
Homolsomites cf. giganteus
Polyptychites cf. keyserlingi
Buchia n. sp. aff. inflata
Tollia cf. tolli
Buchia cf. uncitoides
Buchia okensis
Unknown
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Absent
Buchia cf. fischeriana, B. piochii
Unknown
Unknown

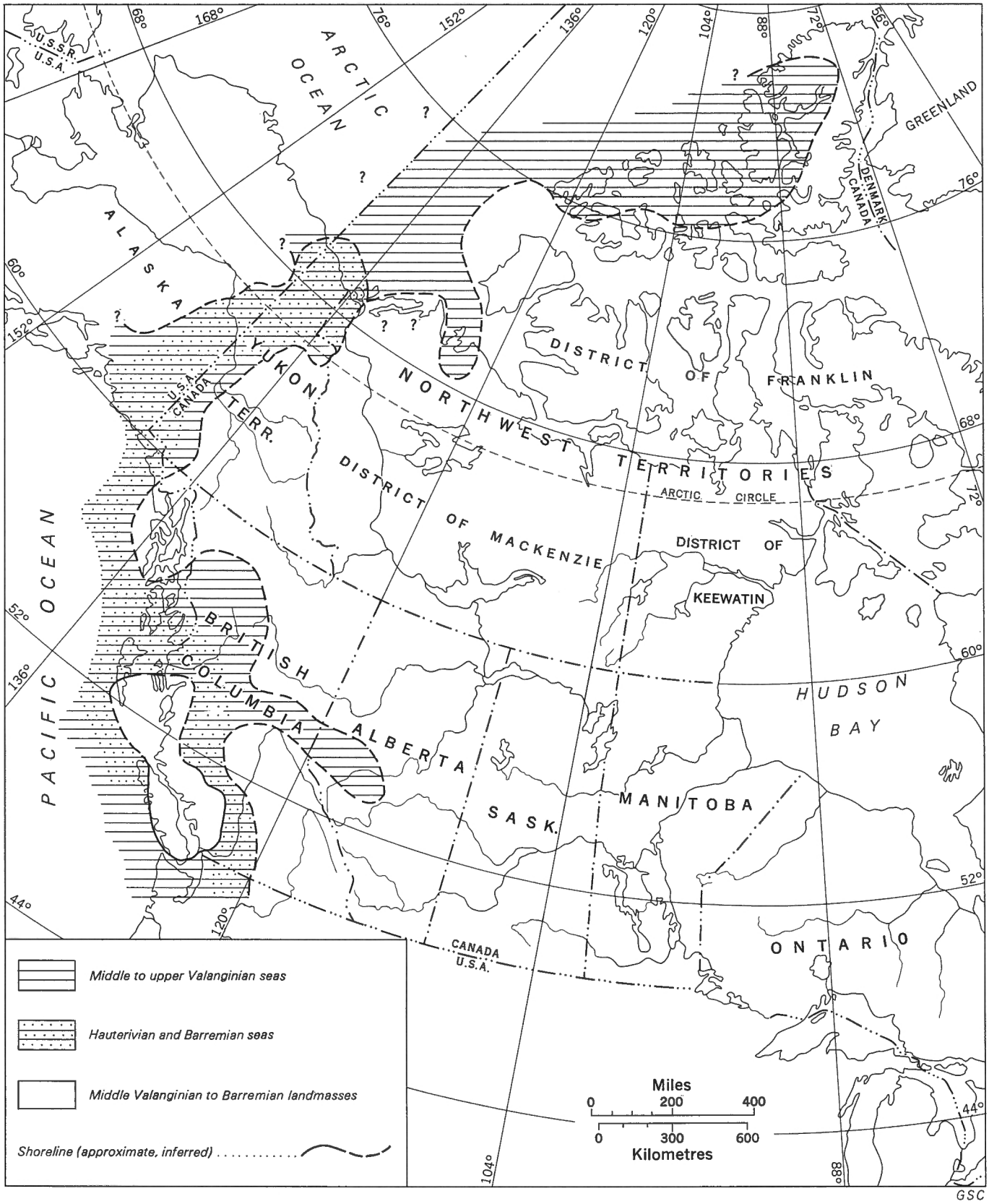

FIGURE XI-2. Inferred maximum extent of Lower Cretaceous (Berriasian to Barremian) seas in western and arctic Canada.

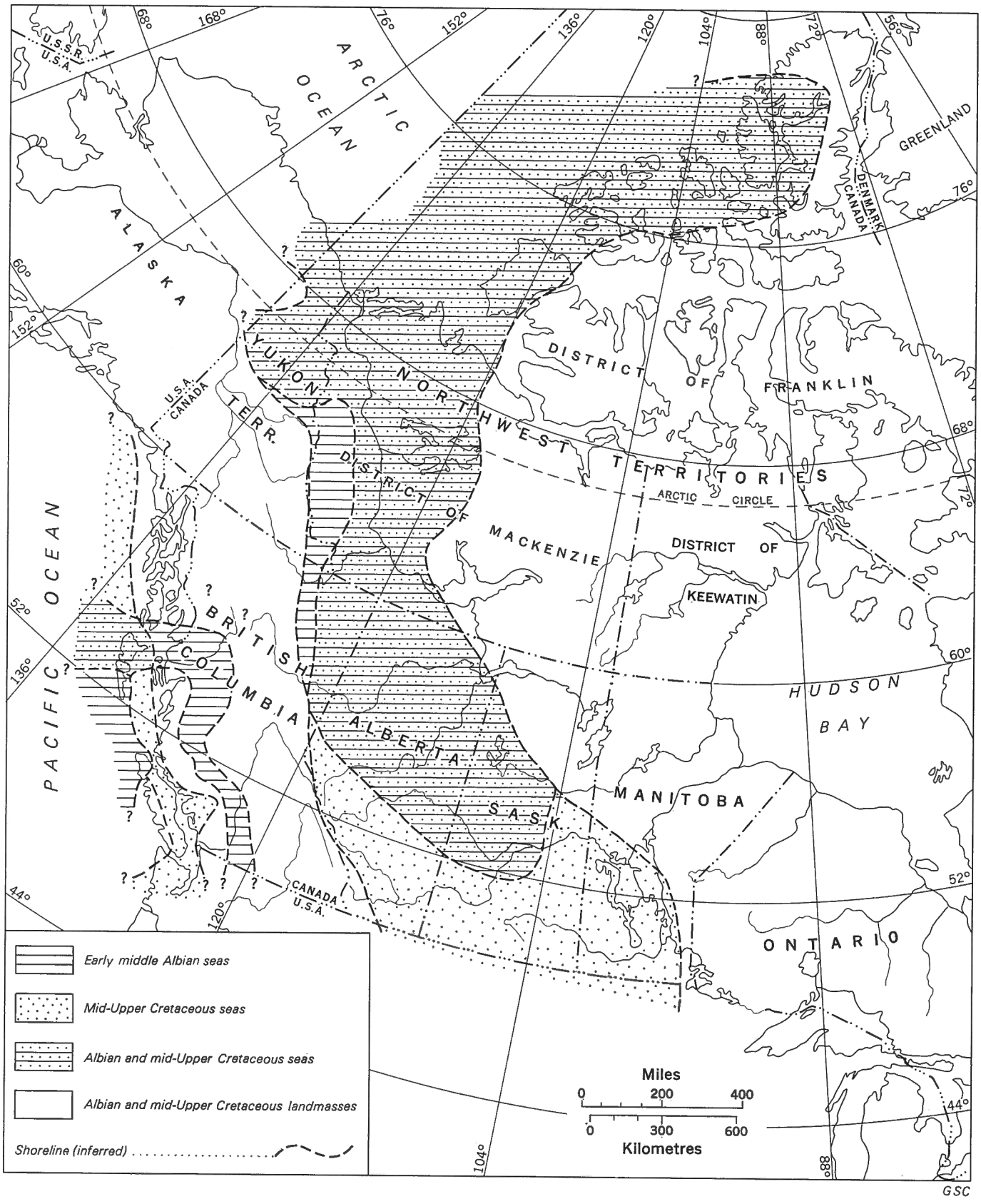

FIGURE XI-3. Inferred maximum extent of Early middle Albian and mid-Upper Cretaceous seas in western and arctic Canada.

A characteristic feature of most boreal Cretaceous faunas in Canada is their lack of diversity; they are, as a rule, dominated by only a few, or even a single, molluscan species occurring in great abundance. This extreme faunal depauperation was probably caused by a relative coldness of water and abundance of silt and clay particles compared to more southerly regions of North America and to the North Pacific realm.

The marine faunas of the North Pacific realm contain a great variety of ammonite, pelecypod, and gastropod species and genera. This is especially true of the mid- to late Upper Cretaceous faunas of the Nanaimo Group (Usher, 1952; Whiteaves, 1876-1903) on Vancouver Island.

Even the earliest Cretaceous, Berriasian, marine faunas of the western Cordillera and Peace River regions differ markedly from their arctic counterparts in the presence of Tethyan ammonites such as *Berriasella, Spiticeras, Neocomites,* and marked endemism of *Buchia* species (Jeletzky, 1964, 1965). However, these endemisms are strongly tempered by the presence of a number of species and genera common to the boreal and North Pacific realms which permit reasonably accurate zonal correlations of their Berriasian faunas. The same conditions persisted through Valanginian, Hauterivian, and Barremian times.

The relatively easy exchange of the Berriasian to Barremian marine faunas between the western Cordillera, Peace River, and arctic regions of Canada was facilitated by the existence of at least two apparently more or less narrow seaways as shown in Figure XI-2. Dawson Strait apparently extended from the Kluane Lake–St. Elias area into the Dawson City area, and thence into the northern Yukon and northern Richardson Mountains. Vanderhoof Strait extended from the headwaters of Skeena River into the Carbon Creek area connecting Bowser and Peace River Basins (Fig. XI-2). The inferred maximum extent and general configuration of the Berriasian to Valanginian seas of western and arctic Canada is exemplified by those of the mid- to late Valanginian time; Berriasian and Lower Valanginian seas were similarly distributed but somewhat less widespread. A strong regression resulted in an apparently complete withdrawal of the Hauterivian seas from the Peace River region and their restriction to the depositional troughs in western Cordillera, northern Yukon, and northern Richardson Mountains. The Barremian seas were apparently as extensive as the Hauterivian seas, except that they also penetrated at times into the Peace River region via Vanderhoof Strait (Jeletzky and Tipper, 1968).

In the western Cordillera, Aptian seas were even more restricted than those of Hauterivian–Barremian time. So far as known, they were limited by strong early Aptian uplifts of Vanderhoof and Dawson Straits. However, in northern Yukon and Richardson Mountains Aptian seas were almost as widespread as the Barremian. They probably extended along the Arctic coast to connect directly with the Aptian seas of the European arctic. This radically new configuration of Aptian land and seas (Warren and Stelck, 1961) was similar to that of early middle Albian time (Fig. XI-3), except that the arctic Aptian seas apparently did not extend south beyond Dawson City area, southern Richardson Mountains and a narrow belt of the Arctic coast between the Mackenzie Delta and Darnley Bay area. The complete isolation of the Aptian western Cordilleran marine basins from those of the arctic and western Interior regions persisted throughout Albian and Upper Cretaceous time and is reflected by almost total dissimilarity of the faunas. Except for the extremely rare occurrence of *Arcthoplites* ex gr. *belli* in the *Brewericeras hulenense* fauna of Queen Charlotte Islands (Jones, Murphy, and Packard, 1965; McLearn, *in press)* and the occurrence of *Inoceramus* cf. *labiatus* in the Honna and Skidegate Formations of the same area, the North Pacific and Boreal realms of Canada lack any common species following the Barremian whereas the western Cordilleran basins have a great many species in common with Japan, India, Pacific coast of Siberia, California, etc.

The western Interior boreal basins were connected with the Gulf of Mexico Tethyan seas as early as the middle Albian (Warren and Stelck, 1961). This connection persisted thereafter at least until late Santonian–early Campanian time and possibly into Maestrichtian time. The rather restricted Pacific Upper Cretaceous transgression probably began in Santonian time, except in Queen Charlotte Islands but the western Interior Upper Cretaceous transgression apparently reached its maximum already in lower Turonian (*Inoceramus labiatus*) time.

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PLATE XXIII. Early Lower Cretaceous (Berriasian and Valanginian) Fossils, North Pacific and Boreal Realms

Figures 1a-1b *Acroteuthis subquadratus* (Roemer). Upper Valanginian; Northwest Territories; 17253.

Figures 2a-2c *Buchia inflata* (Toula) var. *crassa* (Pavlow). Mid-Valanginian; British Columbia; 17333.

Figures 3a-3d *Buchia tolmatschowi* (Sokolov) f. typ. Lower Valanginian; British Columbia; 16623.

Figures 4a-4c *Buchia crassicollis* (Keyserling) f. typ. Upper Valanginian; British Columbia; 16659.

Figures 5a-5b *Buchia volgensis* (Lahusen) s. str. Upper Berriasian; Northwest Territories; 16601.

Figures 6a-6c *Neocomites (Neocomites?)* aff. *indomontanus* Uhlig. Middle Valanginian; British Columbia; 17223.

Figures 7a-7b *Spiticeras (Spiticeras)* sp. indet. Juven., upper Berriasian; British Columbia; 16608.

Figures 8a-8b *Tollia (Tollia) paucicostata* (Donovan) var. Lower Valanginian; British Columbia; 16626.

Figures 9a-9c *Buchia uncitoides* (Pavlow) var. *acutistriata* (Crickmay). Upper Berriasian; British Columbia; 16603.

Figures 10a-10b *Tollia (Subcraspedites)* aff. *analogus* (Bogoslovsky). Upper Berriasian; Yukon; 17138.

Figures 11a-11c *Buchia keyserlingi* (Lahusen) f. typ. Lower Valanginian; Northwest Territories; 17163.

Figures 12a-12b *Tollia (Tollia) tolli* (Pavlow) var. *latelobata* Pavlow. Lower Valanginian; Northwest Territories; 17167.

Figures 13a-13c *Buchia okensis* (Pavlow). Lower Berriasian; Northwest Territories; 17122.

Figures 14a-14b *Berriasella* aff. *gallica* Mazenot. Lower Berriasian; British Columbia; 21835.

PLATE XXIV. Mid-Lower Cretaceous (Hauterivian and Barremian) Fossils, North Pacific Realm

Figures 1a-1c *Speetoniceras agnessense* Imlay. x1, lower Hauterivian; British Columbia; 21821.

Figures 2a-2d *Inoceramus colonicus* Anderson. x1, Hauterivian and lower Barremian; British Columbia; 21822.

Figures 3a-3b *Costidiscus* cf. *striatisulcatus* (d'Orbigny). x1, upper Barremian; British Columbia; 21823.

Figures 4a-4b *Crioceratites (Hemihoplites)* n. sp. ex aff. *C. (H.) soulieri* (Matheron). x1, Barremian; British Columbia; 21824.

Figures 5a-5b *Heteroceras (Heteroceras)* cf. *heliceroides* (Karsten). x1, upper Barremian; British Columbia; 21825.

Figures 6a-6c *Homolsomites oregonensis* (Anderson). x1, lower Hauterivian; British Columbia; 21826.

Figure 7 *Eulytoceras* n. sp. ex aff. *E. inaequalicostatus* (d'Orbigny). x1, Barremian; British Columbia; 21827.

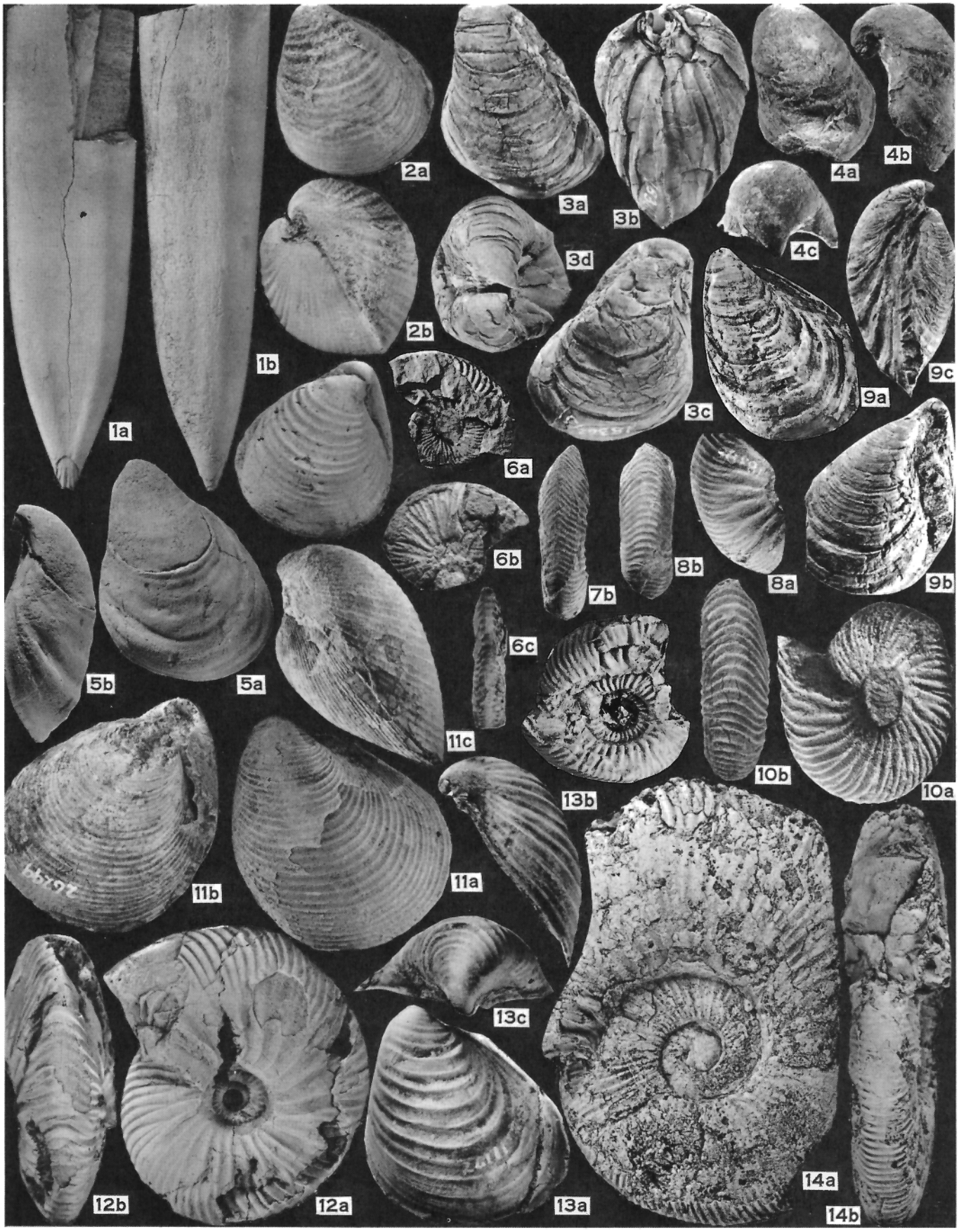
Figures 8a-8b *Phyllopachyceras infundibulum* (d'Orbigny). x1, Barremian; British Columbia; 21828.

Figures 9a-9c *Shasticrioceras* aff. *hesperium* Anderson. x1, Barremian; British Columbia; 21829.

Figures 10a-10c *Heteroceras (Heteroceras)* cf. *heliceroides* (Karsten). x1, upper Barremian; British Columbia; 21830.

Figures 11a-11b *Simbirskites (Hollisites) lucasi* Imlay. xl, lower Hauterivian; British Columbia; 21831.

(Plates XXIII–XXVIII — all figures are natural size unless otherwise stated)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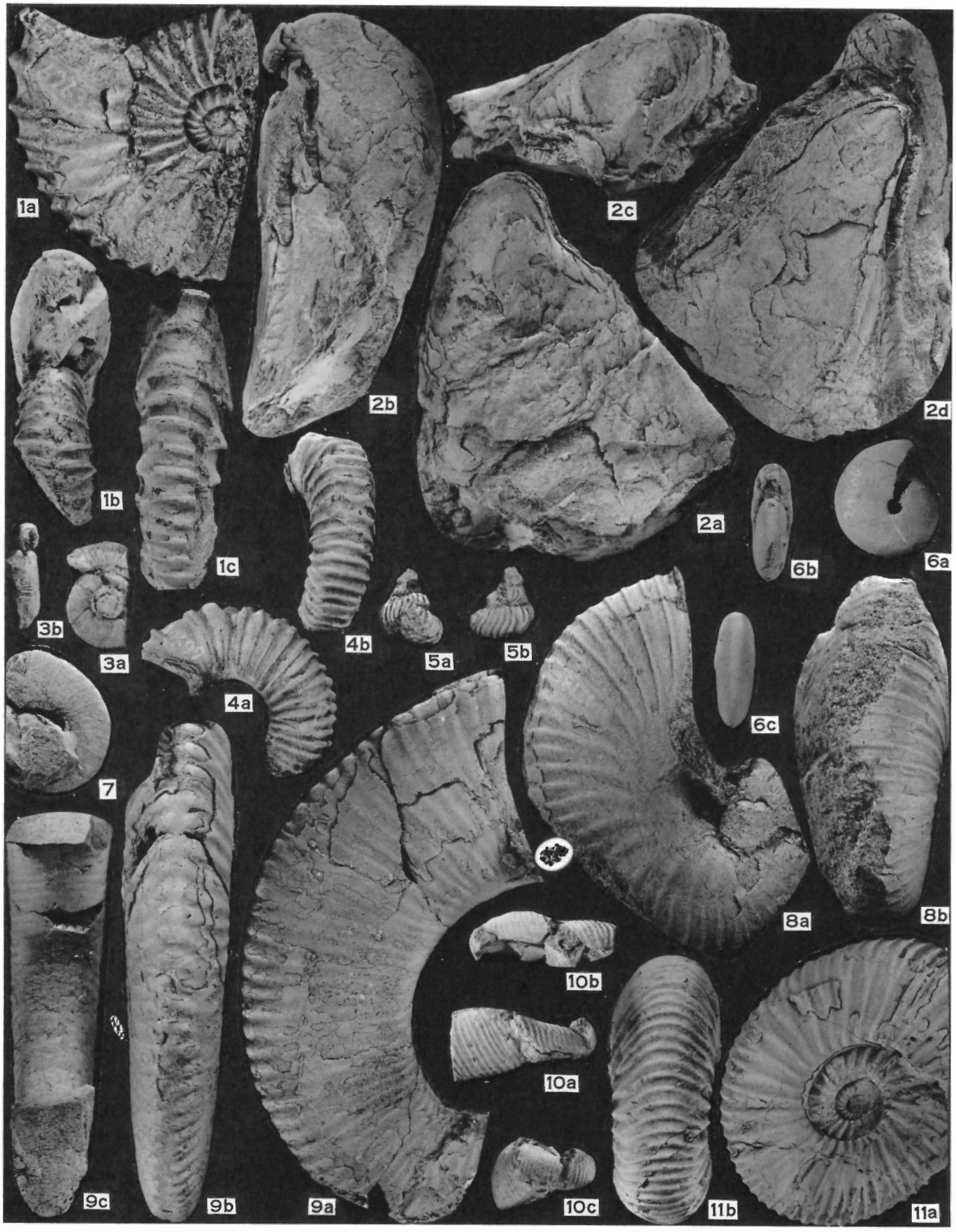
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PLATE XXV. Mid- to Late Lower Cretaceous (Barremian to Albian) Fossils, North Pacific to Boreal Realms

Figures 1a-1b *Douvilleiceras spiniferum* (Whiteaves). Early middle Albian; British Columbia; 21240.

Figures 2a-2b *Cleoniceras* (*Grycia?*) *perezianum* (Whiteaves). Middle Albian; British Columbia; 21233.

Figures 3a-3c *Gabbioceras?* ex gr. *wintunium* Anderson. Lower Aptian or upper Barremian; British Columbia; 21818.

Figures 4a-4c *Douvilleiceras spiniferum* (Whiteaves). Early middle Albian; British Columbia; 5993.

Figures 5a-5b *Ancyloceras (Dissimilites)* n. sp. ex aff. *A.* (*D.*) *dissimile* (d'Orbigny). Barremian; British Columbia; 21819.

Figure 6 *Argonauticeras* aff. *argonautarum* Anderson. Lower Aptian or upper Barremian; British Columbia; 21820.

Figures 7a-7b *Brewericeras hulenense* (Anderson). Early middle Albian; British Columbia; 4984.

Figures 8a-8b *Aucellina gryphaeoides* (J. deC. Sowerby). Albian; British Columbia; 17396.

Figures 9a-9b *Arcthoplites belli* (McLearn). Early middle Albian; British Columbia; 17407.

Figures 10a-10b *Gastroplites canadensis* (Whiteaves). Late middle Albian; British Columbia; 7430.

Figures 11a-11b *Aucellina aptiensis* (d'Orbigny) Pompeckj, 1901, var. *nassibianzi* Sokolov, 1908. Aptian; Mackenzie District, N.W.T.; 17319.

Figures 12a-12b *Tropaeum* n. sp. aff. *arcticum* (Stolley). Early upper Aptian; Mackenzie District, N.W.T. (?); 17322.

Figures 13a-13b *Beudanticeras affine* (Whiteaves). Lower and early middle Albian; Alberta; 17400.

Figures 14a-14b *Neogastroplites maclearni* Reeside and Cobban. Late upper Albian; Alberta; .13658.

Figure 15 *Posidonia? nahwisi* (McLearn) var. *goodrichensis* (McLearn). Early upper Albian; British Columbia; 8943.

PLATE XXVI. Upper Cretaceous (Turonian to Lower Campanian) Fossils, Boreal Realm

Figures 1a-1c *Inoceramus lamarcki* Parkinson. Upper Turonian; Alberta; 21836.

Figures 2a-2b *Scaphites depressus* Reeside. Basal Santonian; Alberta; 21837.

Figure 3 *Prionocyclus* (*Prionocyclus*) *wyomingensis* Meek var. *robusta* Haas. Late upper Turonian; Alberta; 21838.

Figures 4a-4b *Watinoceras* cf. *coloradoense* (Henderson). Lower Turonian; British Columbia; 21839.

Figures 5a-5b *Haresiceras natronense* Reeside. Early lower Campanian; Manitoba; 21840.

Figure 6 *Inoceramus labiatus* (Schlotheim). Lower Turonian; Alberta; 21841.

Figures 7a-7b *Scaphites carlilensis* Morrow. Mid-upper Turonian; Manitoba; 21842.

Figures 8a-8b *Prionocyclus* (*Prionocyclus*) *wyomingensis* Meek var. *elegans* Haas. Late upper Turonian; Alberta; 21843.

Figures 9a-9b *Prionocyclus* (*Collignoniceras*) *woolgari* (Mantell). Early upper Turonian; British Columbia; 21844.

Figure 10 *Desmoscaphites* aff. *bassleri* Reeside. Uppermost Santonian; Alberta; 21845.

PLATE XXVII. Upper Cretaceous (Cenomanian to Maestrichtian) Fossils, Boreal Realm

Figures 1a-1b *Ponteixites robustus* Warren. Uppermost Campanian or ? basal Maestrichtian; Saskatchewan; 21846.

Figures 2a-2b *Inoceramus coulthardi* McLearn. Basal Santonian; British Columbia; 6104.

Figure 3 *Inoceramus patootensis* de Loriol. Upper Santonian to early lower Campanian; Alberta; 21847.

Figures 4a-4b *Baculites compressus eliasi* Cobban. Uppermost Campanian; Saskatchewan; 21848.

Figure 5 *Inoceramus corpulentus* McLearn. Upper Cenomanian; British Columbia; 6109.

Figure 6 *Scaphites hippocrepis* deKay. Early lower Campanian; Saskatchewan; 21849.

Figures 7a-7b *Scaphites nodosus* Owen. Mid-upper Campanian; Saskatchewan; 5369.

Figure 8 *Inoceramus steenstrupi* de Loriol. x½, early lower Campanian and ?upper Santonian; Northwest Territories; 21851.

Figures 9a-9b *Scaphites brevis* Meek. Uppermost Campanian or ?basal Maestrichtian; Saskatchewan; 21852.

Figures 10a-10b *Scaphites* (*Discoscaphites*) ex gr. *roanensis* Stephenson. Lower? Maestrichtian; Saskatchewan; 21853.

PLATE XXVIII. Mid- to Late Upper Cretaceous (Santonian to Campanian) Fossils, North Pacific Realm

Figures 1a-1b *Inoceramus schmidti* Michael s. str. Early upper Campanian?; British Columbia; 5832.

Figures 2a-2b *Inoceramus naumanni* Yokoyama s. str. Upper Santonian?; British Columbia; 21832.

Figures 3a-3b *Epigoniceras epigonum* (Kossmat). Upper Santonian to early lower Campanian; British Columbia; 10015.

Figures 4a-4b *Neophylloceras ramosum* (Meek). Latest upper Campanian?; British Columbia; 5811.

Figures 5a-5c *Inoceramus elegans* Sokolov. Early upper Campanian?; British Columbia; 21833.

Figure 6 *Nostoceras hornbyense* (Whiteaves). Latest upper Campanian?; British Columbia; 10069.

Figure 7 *Bostrychoceras elongatum* (Whiteaves). x½, upper Santonian to early lower Campanian; British Columbia; 10062.

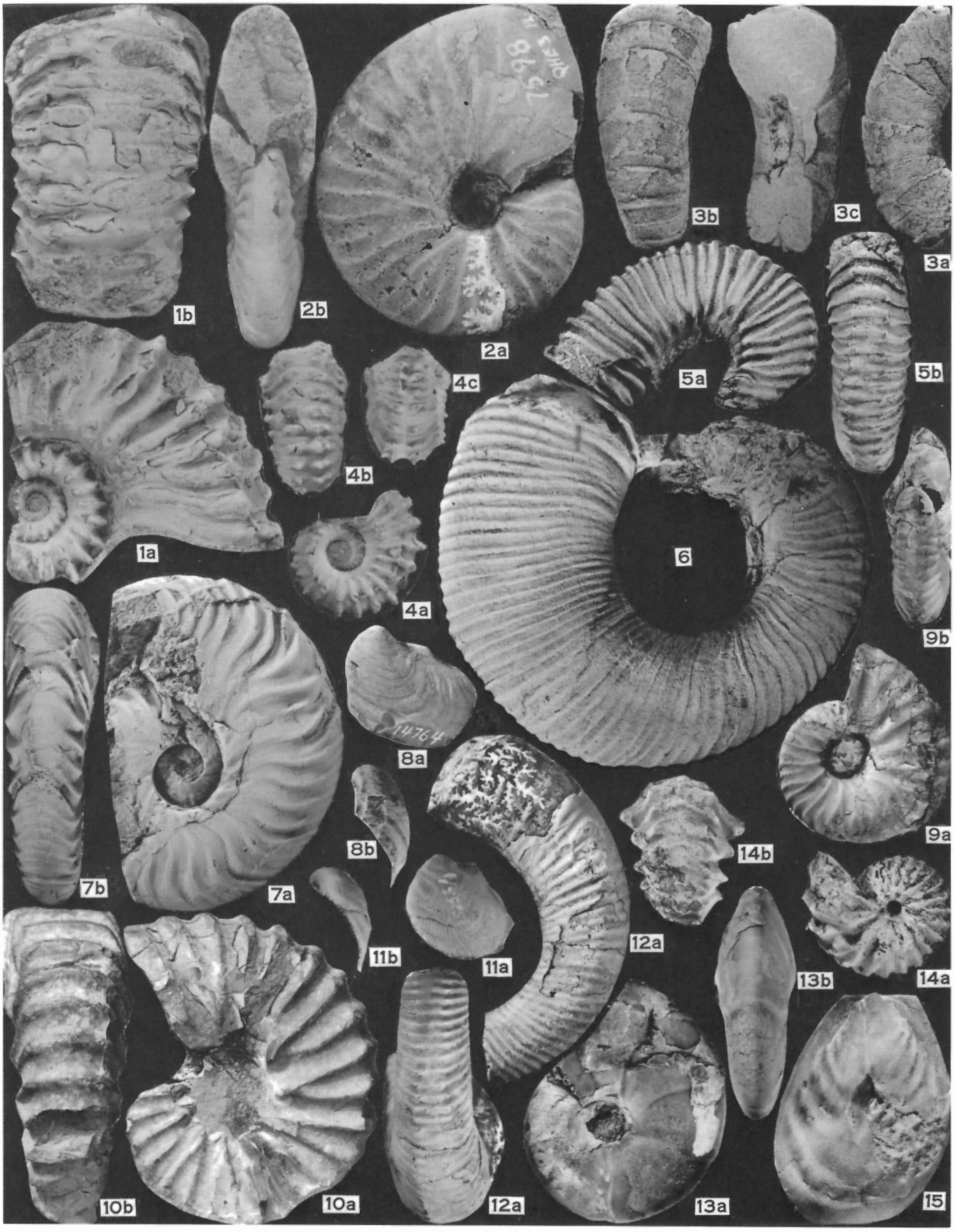
Figure 8 *"Hamites" obstrictus* Jimbo. Latest upper Campanian?; British Columbia; 5958.

Figures 9a-9c *Inoceramus* cf. *sachalinensis* Sokolov. Early upper Campanian?; British Columbia; 21834.

Figures 10a-10b *Schluteria selwyniana* (Whiteaves). Upper Santonian to early lower Campanian; Washington, U.S.A.; 5803b.

Figures 11a-11b *Ptychoceras vancouverense* Whiteaves. Upper Santonian to early lower Campanian; British Columbia; 5798.

Figures 12a-12b *Pachydiscus suciaensis* (Meek). Latest upper Campanian?; British Columbia; 10035.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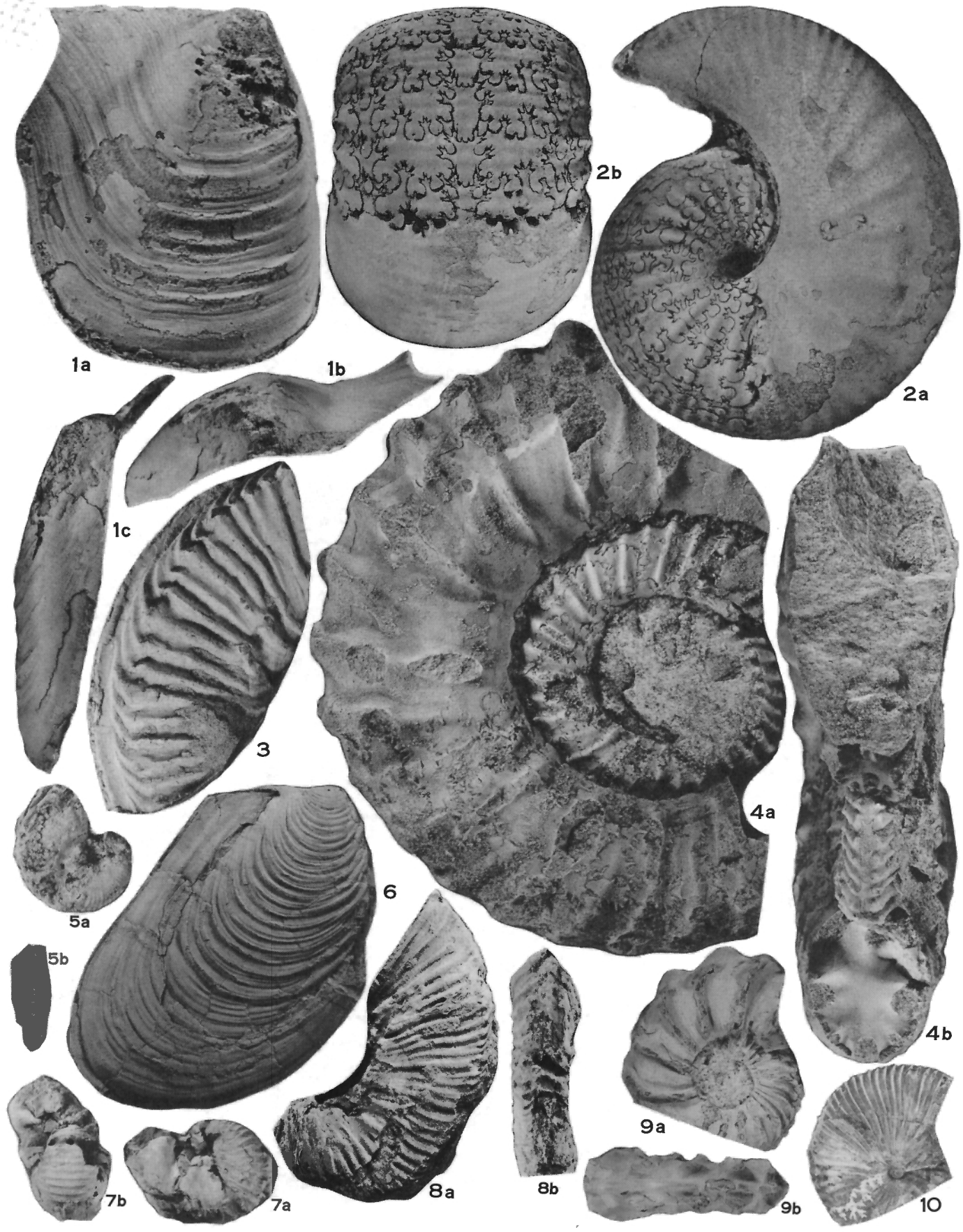
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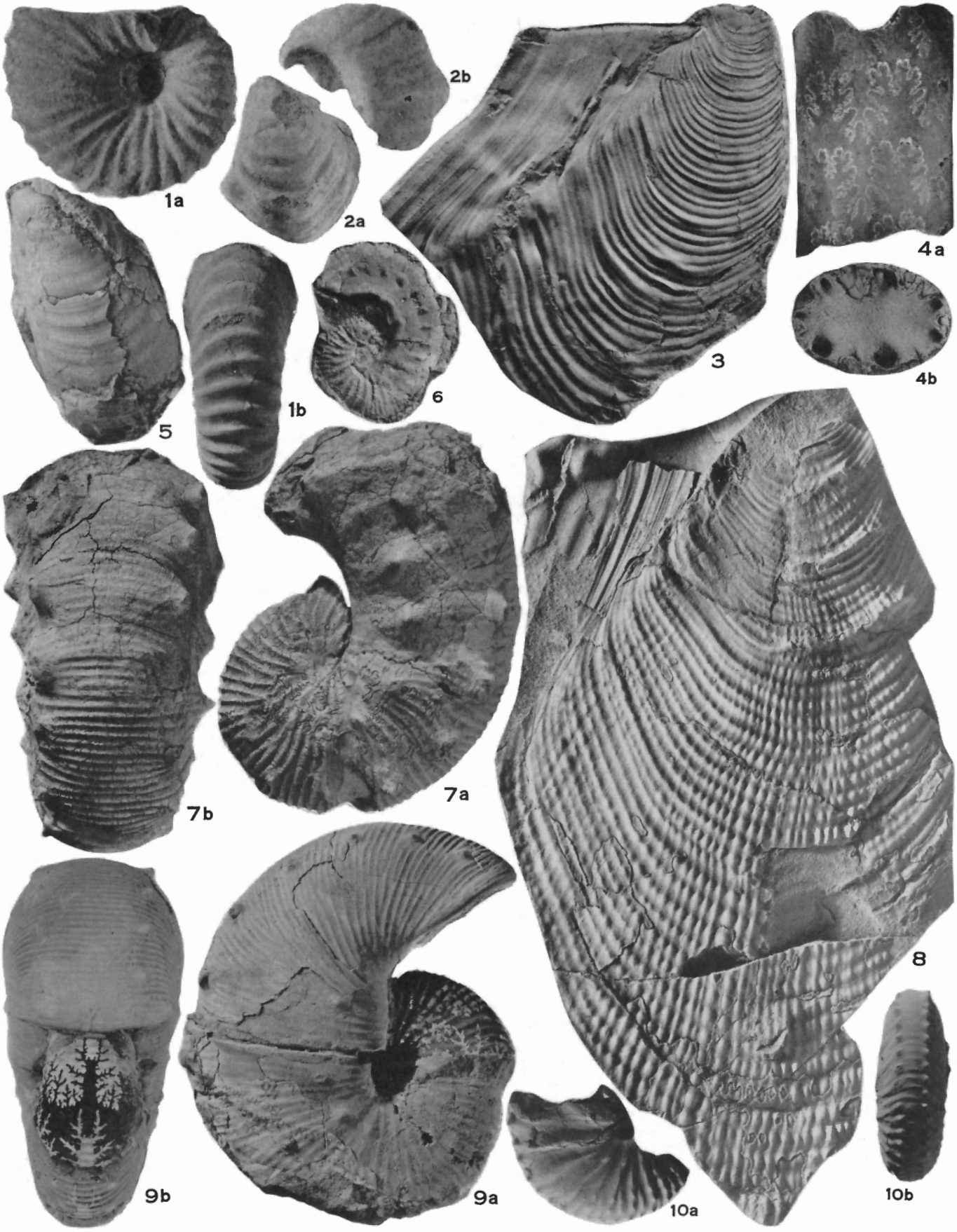
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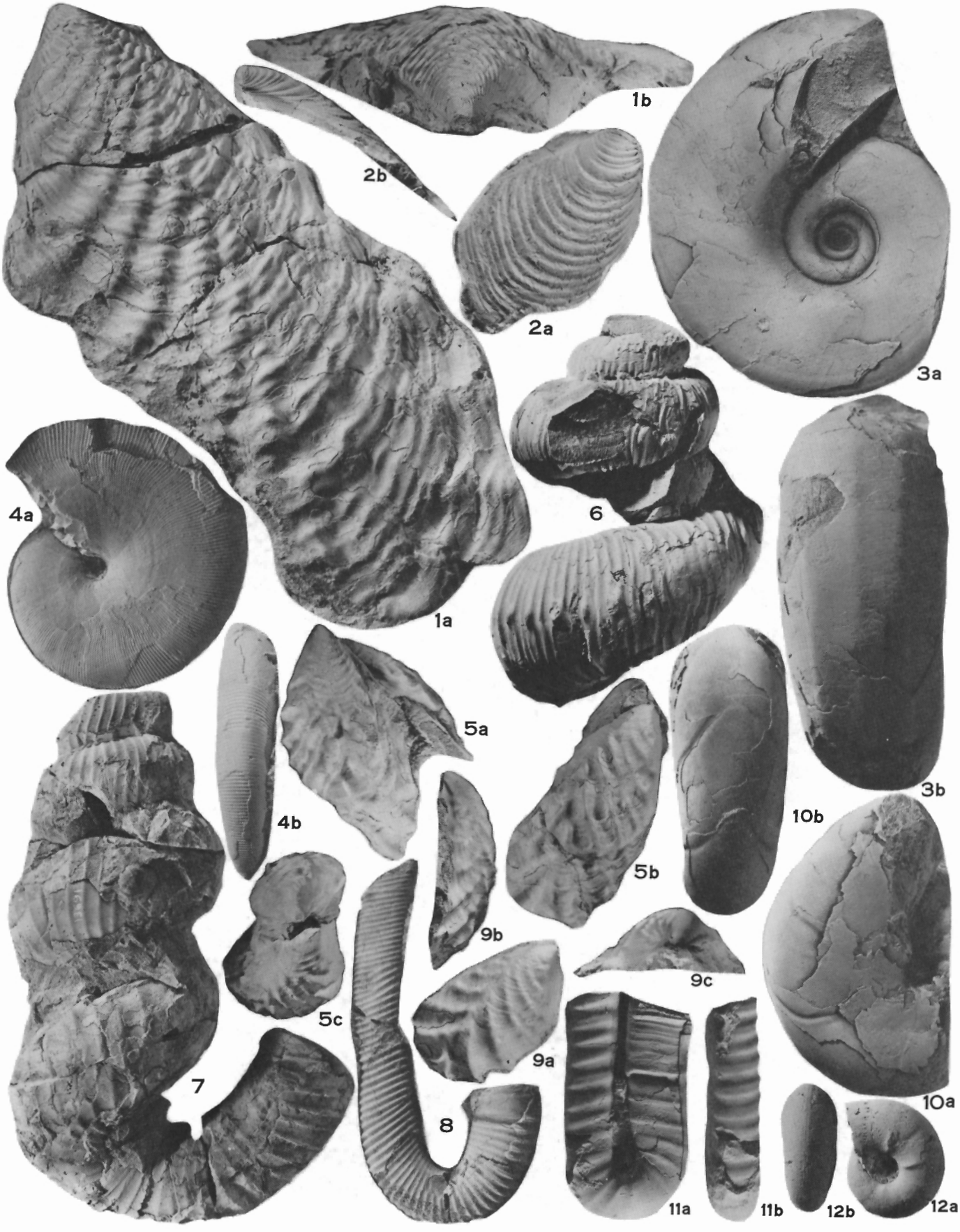
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PALEOBOTANY

by D. C. McGregor

Until about fifteen years ago the attention of paleobotanists in Canada was directed almost exclusively towards megascopic remains. More recently, however, palynologists have disclosed a great variety of microfossils, including spores, pollen, acritarchs, dinoflagellates, diatoms, and other algal, fungal and cuticular fragments in Canadian rocks, and paleobotanical research in this country is now directed towards both megafossils and microfossils. A brief résumé of the development and present status of paleobotanical studies in Canada is given here.

One of the more recent emphases in Canadian paleobotany involves the organic microplankton (acritarchs and dinoflagellates). Acritarchs (PL. XXXI, FIGS. 40, 44), apparently all marine but of unknown biological affinities, have been discovered in rocks representing all the geological systems from Cambrian to Tertiary. Dinoflagellates (PL. XXXI, FIG. 45), both marine and non-marine, have a shorter geological range in Canada, from Permian to Recent. Although the study of Canadian acritarchs and dinoflagellates is still in its infancy, these fossils have considerable promise for local stratigraphic zonation and long-distance correlation.

The remains of terrestrial plants, in the form of petrified, carbonized or mummified megascopic fragments, and highly varied assortment of spores and pollen are known in rocks ranging in age from Devonian to Pleistocene. They are most abundant in Devonian and Carboniferous rocks of the Gaspé Peninsula and the Maritime Provinces, Devonian rocks of the arctic, and Middle Jurassic through Tertiary rocks of western Canada and the Sverdrup Basin. Table XI-9 summarizes the known occurrences of taxonomically identifiable trilete spores, pollen, and terrestrial megaplants in Canada.

The Lower Devonian *Psilophyton–Drepanophycus–Prototaxites* flora of Gaspé (PL. XXIX, FIG. 1) was the first to be described, by Sir William Dawson in several papers between 1856 and 1882; it is now known throughout the world as a classic example of one of the earliest known land floras. In addition to megascopic remains, trilete spores are abundant in the Gaspé Sandstone (PL. XXXI, FIGS. 1–4). Both the megafossils and the microfossils in these beds have definite interregional and intercontinental affinities. Similar floras occur in the Sextant Formation of the Hudson Bay Lowland, the Old Red Sandstone of Scotland, and Lower Devonian rocks of Germany, Belgium, Spain, and eastern Siberia.

Upper Devonian plant remains have been discovered in several parts of Canada. Well-preserved specimens of the gymnosperm *Archaeopteris* (PL. XXIX, FIG. 2), associated with Frasnian fish, occur in the Escuminac Formation of northern New Brunswick. *Archaeopteris* has been found only in the Upper Devonian, and is of worldwide distribution in both northern and southern hemispheres. Occurring with it but not produced by it are large spores with distinctive bifurcate appendages, similar to the specimen illustrated on PLATE XXXI, FIGURE 7. These spores also have a worldwide distribution, but a somewhat longer stratigraphic range than *Archaeopteris*—from Emsian to latest Famennian in Canada. Rare single specimens of this type of spore have been reported from Lower Carboniferous rocks outside of Canada, but they are abundant only in the Devonian.

Archaeopteris and the "anchor-spined" spores occur together in the Griper Bay Formation of the Arctic Archipelago (PL. XXXI, FIG. 7) and in the Imperial Formation of the Mackenzie Basin. Other spores occur in great abundance and variety in these beds, as well as in other marine and non-marine Givetian to Famennian rocks of Canada (PL. XXXI, FIGS. 5, 6, 8). Several successive spore assemblage zones are recognized in Middle and Upper Devonian rocks of southeastern, western, and arctic Canada, many species of which have worldwide distribution, for example *Archaeoperisaccus* spp. and *Hymenozonotriletes lepidophytus* Kedo (PL. XXXI, FIGS. 5, 6).

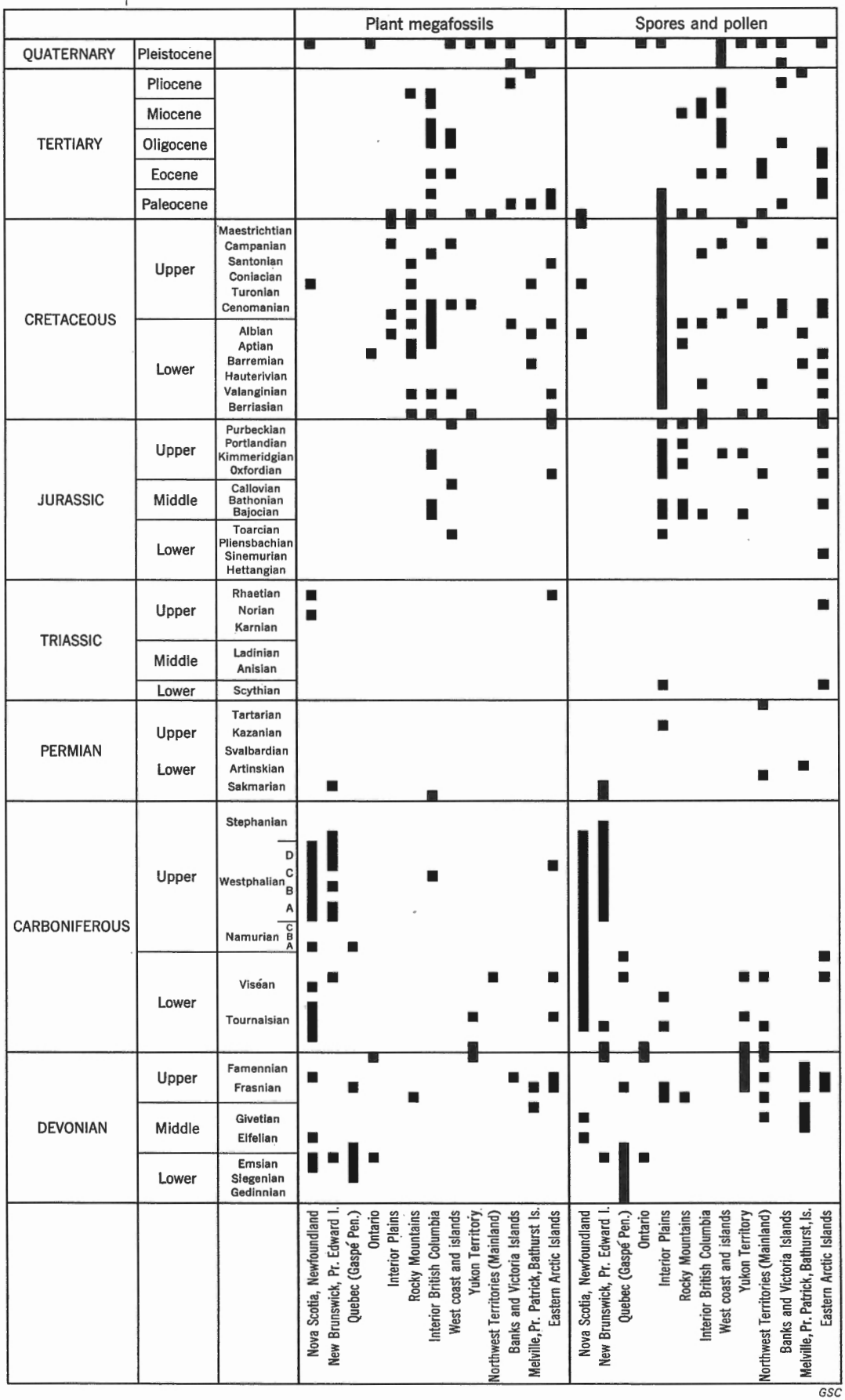
The Carboniferous floras of New Brunswick and Nova Scotia are well known as a result of the work of W. A. Bell. Plant megafossil zones have been established by Bell in the Horton Group (Tournaisian), Windsor Group (Viséan), Canso Group (late Viséan and early Namurian), Riversdale Group (late Namurian–early Westphalian), Cumberland Group (Westphalian B), and Pictou Group (*Lonchopteris* and *Linopteris obliqua* zones, Westphalian C, and the *Ptychocarpus unitus* zone, Westphalian D). Representative plant fossils from the Carboniferous of the Maritimes are illustrated on PLATE XXIX, FIGURES 3–10.

Spore assemblage zones have also been recognized in the Carboniferous rocks of the Maritimes. It is now possible to correlate using spores, not only from shales bearing megaplant fossils but also from other intervals including the coals themselves, where megafossils are not identifiable.

Inter-regional studies of Lower Carboniferous spores in Canada have resulted in the recognition of two distinct and apparently synchronous Viséan assemblages, one in the Windsor Group of the Maritimes and the other in the Golata and Mattson Formations of western Canada, and the Emma Fiord Formation of the arctic islands. The latter assemblage includes the spores shown on PLATE XXXI, FIGURES 9 and 10 and has a remarkable similarity in species to those of the Viséan of the western U.S.S.R. and Spitsbergen.

As information is accumulated on stratigraphic range, distribution, and facies sensitivity of Carboniferous spores and pollen, certain individual taxa are assuming biostrati-

TABLE XI-9 *Known occurrences (1966) of taxonomically identifiable plant fossils in Canada*



graphic significance. For example, *Lophozonotriletes rarituberculatus* (PL. XXXI, FIG. 14) is characteristic of Lower Carboniferous (Tournaisian) assemblages in various parts of the world. In the Upper Carboniferous, species of *Vestispora* and *Torispora* (PL. XXXI, FIGS. 11 and 16) and several other genera are important biostratigraphic markers because of their restricted time range and wide geographic distribution.

There are relatively few Permian or Triassic fossil plants in Canada. The Permian megafossils that occur on Prince Edward Island are too poorly preserved to be of value for correlation or dating purposes. Spores and pollen, on the other hand, do occur in several areas. For example, Lower Permian spores and pollen (PL. XXXI, FIGS. 13, 15, and 27) including *Nuskoisporites, Vittatina* and disaccate Striatiti, have been recovered from Prince Edward Island, from the upper beds of the Mattson Formation of Liard Plateau, and from the Sabine Bay Formation of Melville Island. Upper Permian spores and pollen, including *Vittatina* (PL. XXXI, FIG. 20), occur in the subsurface of the Interior Plains and the Yukon Territory.

Late Triassic megafossil plants, apparently similar to those that occur in Rhaetic to Liassic strata of Eastern Greenland, have been discovered in the upper part of the Heiberg Formation in the Queen Elizabeth Islands. The upper Heiberg also contains well-preserved spores and pollen (PL. XXXI, FIGS. 18, 19, 21, and 24), some of which are identical to species in Rhaetic to Liassic rocks of Europe. Lower and Upper Triassic plant remains occur in certain other areas of Canada as well (Table XI-9), but their botanical affinities and stratigraphic implications are not yet known.

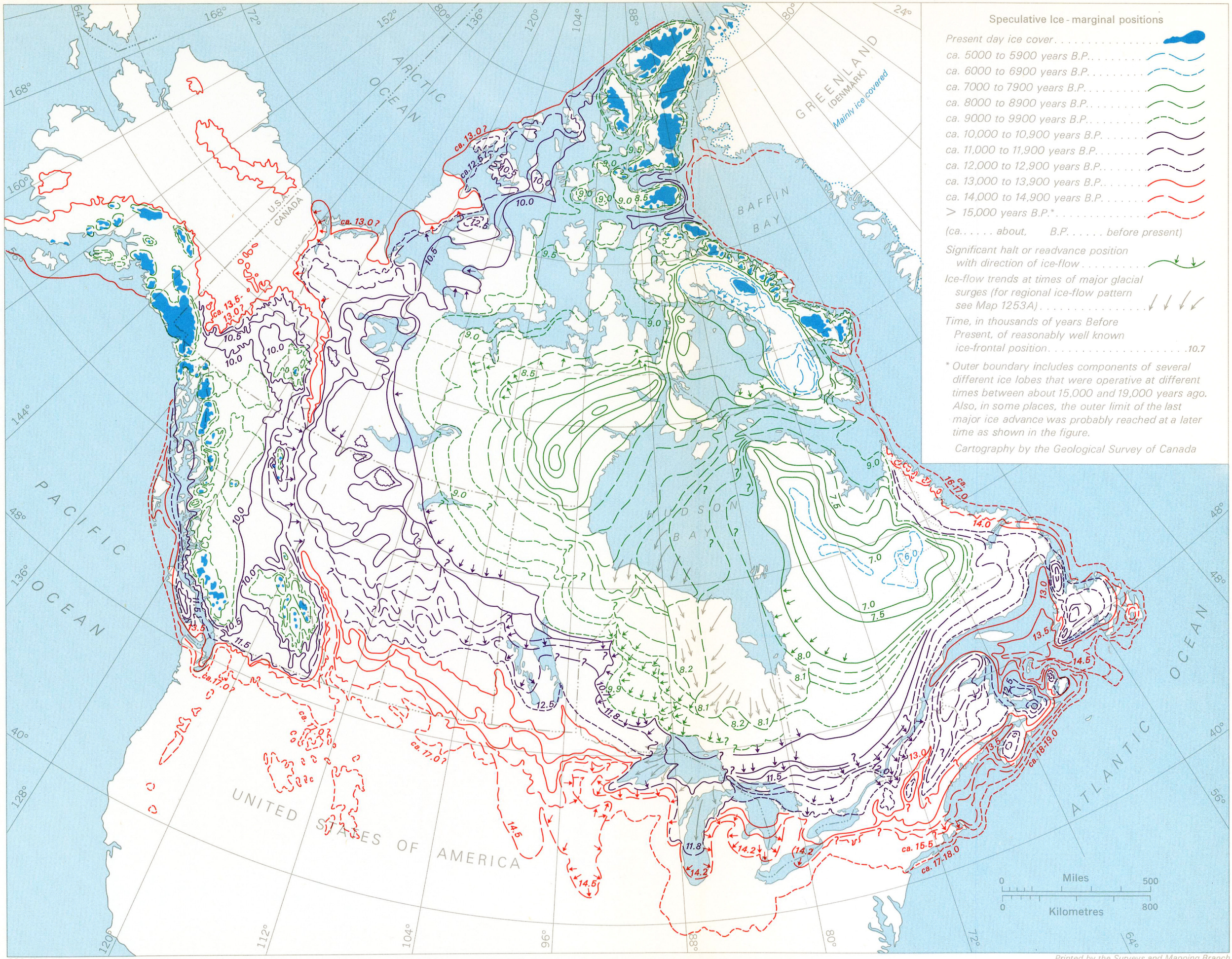
The principal areas of occurrence of Jurassic terrestrial plants are parts of British Columbia, the southern Interior Plains, and the Queen Elizabeth Islands. One of the most significant areas, both botanically and stratigraphically, is the west coast of Vancouver Island and the Queen Charlotte Islands. The fossils on Vancouver Island, comprising petrifactions, carbonized compressions of megaplants (PL. XXX, FIG. 1), and corroded spores occur in beds dated as Toarcian by marine faunas. This flora existed around the fringes of the Pacific from Mexico to southeast Asia. A slightly younger, probably Callovian, flora of undetermined affinities has also been found on Vancouver Island and the Queen Charlotte Islands. Lower, Middle, and Upper Jurassic plant megafossils and microfossils, usually highly carbonized, occur in various parts of the western mainland. Well-preserved spores and pollen have been obtained from Sinemurian and Callovian to Kimmeridgian strata of the Queen Elizabeth Islands (PL. XXXI, FIGS. 17, 22, 23, and 25), and megaplants have been reported but not collected.

Cretaceous sedimentary deposits are widespread in Canada, and have received probably more attention by paleobotanists than those of any other age. They are of special significance both geologically and botanically, for they are an important source of coal, oil, and gas, and contain floras in which angiospermous plants first make their appearance.

Both Lower and Upper Cretaceous megafossil plants of western Canada (PL. XXX, FIGS. 2–9, 12, 15) have been studied extensively by W. A. Bell, and in recent years palynological investigations encompassing most of the Lower and Upper Cretaceous of western Canada have been undertaken. Much of the palynological data is still unpublished, but sufficient information is now available to establish both local and long-distance age correlations based on miospore assemblages. In some instances, individual miospore taxa may be significant zone fossils, provided various factors affecting their occurrence such as lithological variation, diagenetic compaction, and paleoclimatic effects, are properly understood. For example, species of *Plicatella, Appendicisporites, Aquilapollenites* (PL. XXXI, FIG. 39), *Wodehouseia* (PL. XXXI, FIG. 33), and *Proteacidites* (PL. XXXI, FIG. 34) appear to have similar stratigraphic ranges in the Cretaceous of North America, Europe, U.S.S.R., and Australia. In the Arctic Archipelago, the Cretaceous Isachsen, Christopher, Hassel, and Kanguk Formations contain plant remains, the spores and pollen of which are currently under study.

Thick sequences of plant-bearing Tertiary sediments occur in western Canada and the arctic islands. The earliest studies in western Canada were of fossil leaves, fruits, and seeds by J. W. Dawson, D. P. Penhallow, E. W. Berry, and more recently by W. A. Bell. They revealed the presence of Paleocene, Middle Eocene, and Miocene–Pliocene floras in several small intermontane sedimentary basins of British Columbia and Yukon, and in widespread non-marine formations of the Interior Plains. Systematic studies of spores and pollen from the plains are in progress. In British Columbia, Rouse and Mathews have related the megafloral and microfloral content of several deposits to the radiogenic ages of associated ash beds.

Accounts of Tertiary megafossil plants of the Canadian arctic were written by Heer and Nathorst several decades ago, but need extensive revision. However, even though many specimens may be incorrectly identified, the early to middle Tertiary floras of the arctic strongly suggest a temperate climate, the paleogeographical and paleoecological implications of which have not been thoroughly investigated. In the Pleistocene epoch, interglacial and interstadial intervals have been delimited in Canada by paleobotanical studies. The development of muskeg and its relationship to forest growth on peat have also been studied using paleobotanical methods. In contrast to paleobotanical studies of pre-Pleistocene deposits, the evolutionary changes and extinction of species have a more limited use in the Pleistocene and the emphasis is placed primarily on climatic changes and species composition of the vegetation.



GSC 1967

FIGURE XII-15. Stages in the deglaciation of Wisconsin ice.

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Plate XXIX. Devonian and Carboniferous Plants

(All figures natural size unless otherwise stated)

Figure 1 *Psilophyton princeps* var. *ornatum* Dawson. Lower Devonian; Gaspé Peninsula; 13048.

Figure 2 *Archaeopteris jacksoni* Dawson. Frasnian; Gaspé Peninsula; 13050.

Figure 3 *Aneimites acadica* Dawson. Tournaisian; Nova Scotia; 720.

Figure 4 *Lepidodendropsis corrugata* (Dawson) Bell. Tournaisian; Nova Scotia; 15098.

Figure 5 *Calamites suckowi* Brongniart. Westphalian B; Nova Scotia; 14920.

Figure 6 *Alethopteris serli* (Brongniart) Goeppert. Early Westphalian D; Nova Scotia; 14996.

Figure 7 *Linopteris obliqua* (Bunbury) Zeiller. x2, early Westphalian C; Nova Scotia; 14945.

Figure 8 *Lepidodendron aculeatum* Sternberg. Westphalian B; Nova Scotia; 14937.

Figure 9 *Ptychocarpus unitus* (Brongniart) Zeiller. x2, early Westphalian D; Nova Scotia; 1370.

Figure 10 *Asterophyllites equisetiformis* (Schlotheim) Brongniart. Early Westphalian D; Nova Scotia; 3003.

Plate XXX. Mesozoic and Cenozoic Plants

(All figures natural size)

Figure 1 *?Dictyophyllum* sp. Lower Jurassic; Vancouver Island; 15099.

Figure 2 *Cladophlebis virginiensis* Fontaine forma *martiniana* Dawson. Neocomian-Barremian; British Columbia; 13446.

Figure 3 *Athrotaxites berryi* Bell. Aptian or early Albian; Alberta; 13462.

Figure 4 *Cladophlebis yukonensis* Bell. Neocomian-Barremian; Alberta; 13440.

Figure 5 *Nilssonia californica* Fontaine. Aptian or early Albian; Alberta; 13510.

Figure 6 *Araliaephyllum westoni* (Dawson) Bell. Albian; Alberta; 13541.

Figure 7 *Sagenopteris williamsii* (Newberry) Bell. Aptian or early Albian; Alberta; 13474.

Figure 8 Cone, *Sequoia* sp. Santonian-Campanian; Alberta; 5092.

Figure 9 Seed, *Carpites* sp. Santonian-Campanian; Alberta; 5098.

Figure 10 *Metasequoia occidentalis* (Newberry) Chaney. Paleocene; Alberta; 5991.

Figure 11 *Ilex? mammillata* Bell. Santonian-Campanian; Alberta; 1249.

Figure 12 *Artocarpus* sp. Maestrichtian; Alberta; 14856.

Figure 13 Cone, *Larix laricina* (DuRoi) K. Koch. Pleistocene; Mackenzie River Delta; 3586.

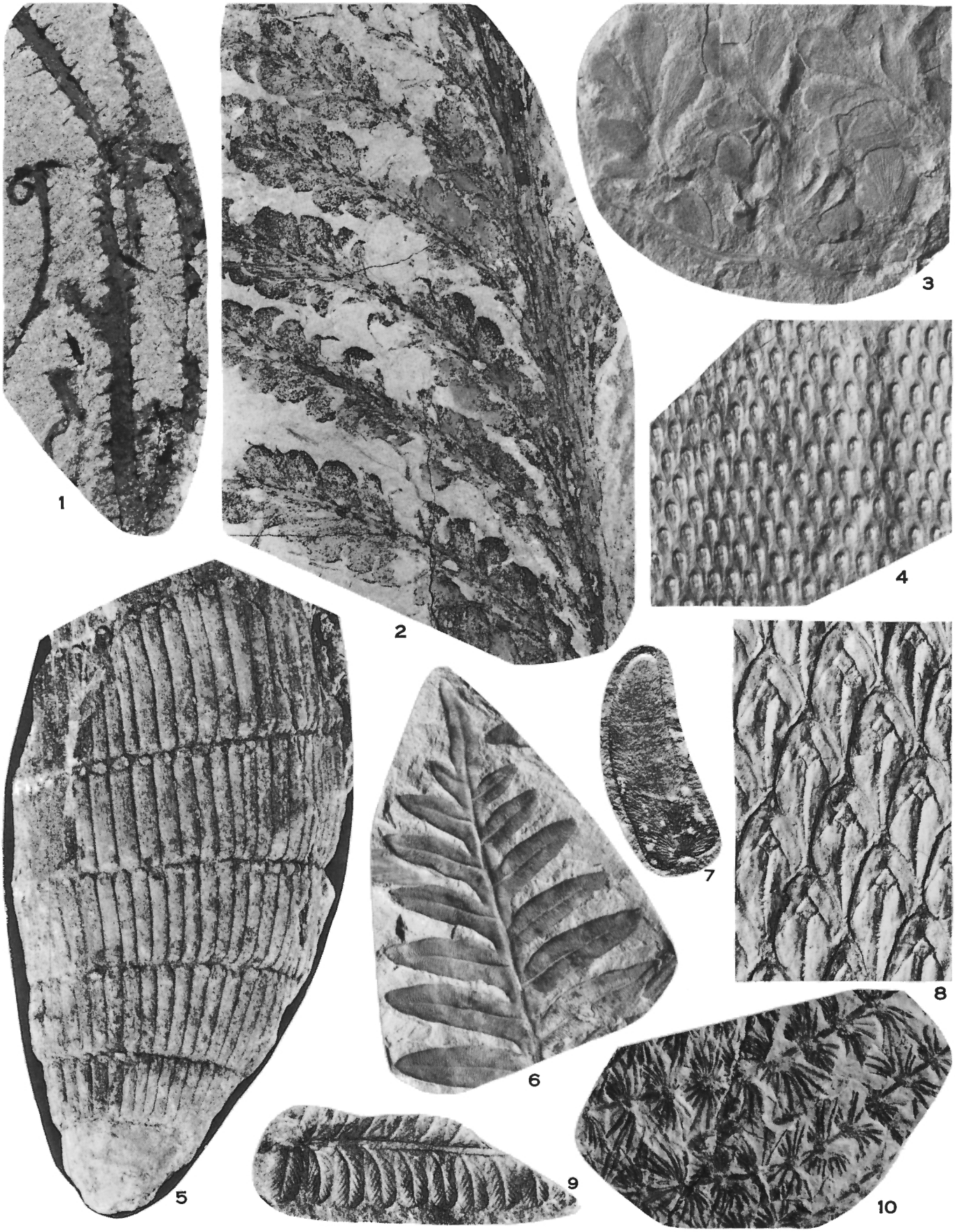
Figure 14 *Viburnum asperum* Newberry. Paleocene; Alberta; 14838.

Figure 15 *Pseudoprotophyllum boreale* (Dawson) Hollick. Cenomanian; Alberta; 5398.

Figure 16 *Osmunda macrophylla* Penhallow. Paleocene; Alberta; 6163.

Figure 17 Cone of *Pinus contorta* Dougl. Pleistocene (late-glacial); Vancouver Island; 3556.

Figure 18 *Dryas drummondii* Richards. Pleistocene (late-glacial); Vancouver Island; 3573.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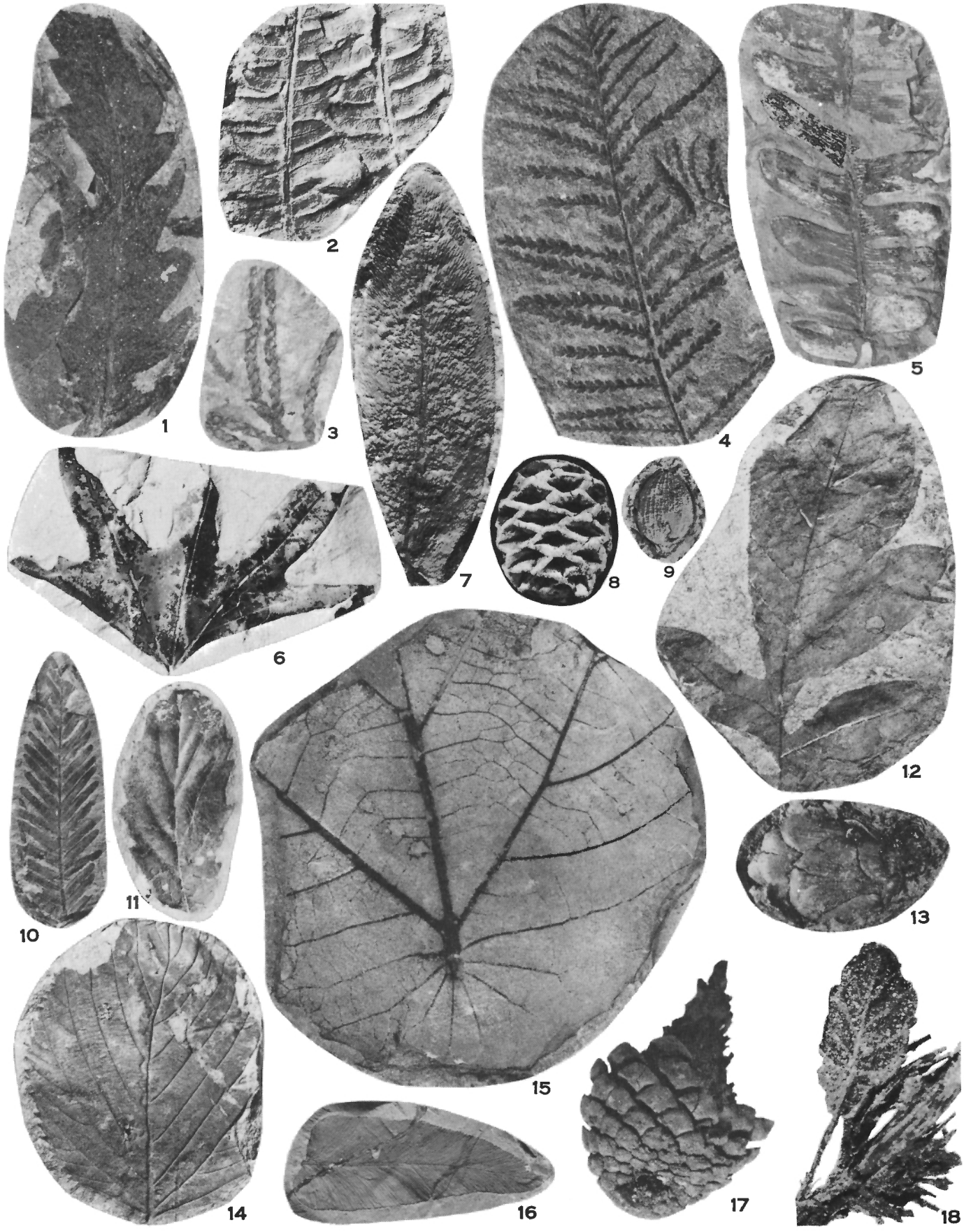
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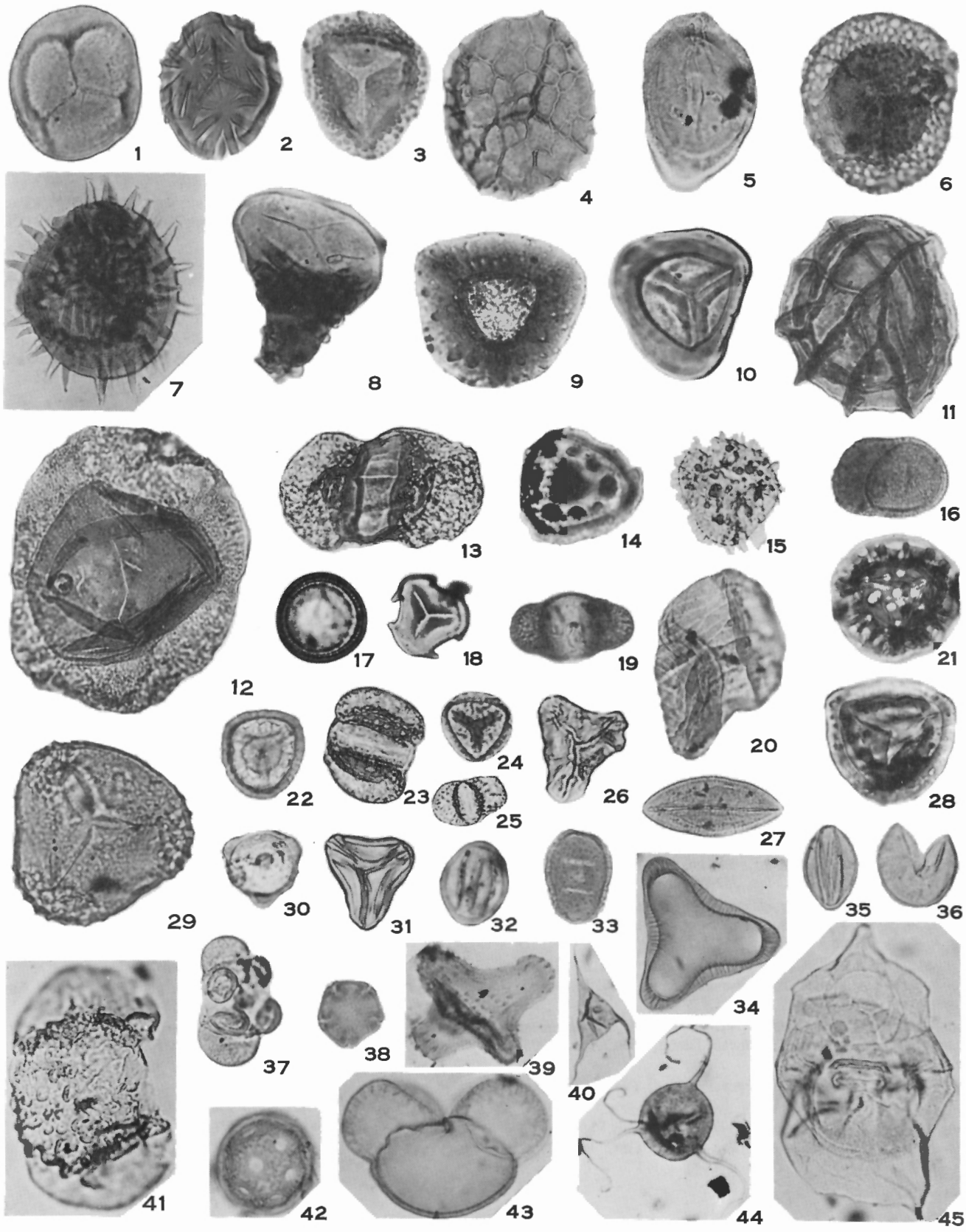
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PLATE XXXI. Palynomorphs

(Fig. 7 x250; others x500)

Figure 1 *Retusotriletes* sp. Lower Devonian; Gaspé Peninsula, GSC loc. 6583; 15064.

Figure 2 *Emphanisporites* sp. Lower Devonian; Gaspé Peninsula, GSC loc. 7094; 15065.

Figure 3 Unidentified spore. Lower Devonian; Gaspé Peninsula, GSC loc. 6583; 15066.

Figure 4 *Reticulatisporites ?emsiensis* Allen. Lower Devonian; Gaspé Peninsula, GSC loc. 6583; 15067.

Figure 5 *Archaeoperisaccus* sp. Middle Frasnian; Great Slave Lake, N.W.T., GSC loc. 30425; 15068.

Figure 6 *Hymenozonotriletes lepidophytus* Kedo. Late Famennian; southern Ontario, GSC loc. 7057; 15069.

Figure 7 *Hystricosporites* sp. Frasnian; Prince Patrick Island, GSC loc. 7026; 15070.

Figure 8 *Lophozonotriletes cristifer* (Luber) Kedo. Late Frasnian or early Famennian; Bathurst Island, GSC loc. 59041; 13625.

Figure 9 *Densosporites spitsbergensis* Playford. Viséan; South Nahanni River, N.W.T., GSC loc. 5004; 15071.

Figure 10 *Murospora aurita* (Waltz) Playford. Viséan; South Nahanni River, N.W.T., GSC loc. 5004; 15072.

Figure 11 *Vestispora costata* (Balme) Bhardwaj. Early Westphalian C; New Brunswick, GSC loc. 7570; 15073.

Figure 12 *Potonieisporites* sp. Lower Permian; Prince Edward Island, GSC loc. 7571; 15074.

Figure 13 *Protohaploxypinus* sp. Lower Permian; Prince Edward Island, GSC loc. 7571; 15075.

Figure 14 *Lophozonotriletes rarituberculatus* (Luber) Kedo. Tournaisian; Nova Scotia, GSC loc. 2531; 15076.

Figure 15 *Neoraistrickia* sp. Mid-Permian; Melville Island, GSC loc. 5221; 15077.

Figure 16 *Torispora securis* Balme. Westphalian C; Nova Scotia, GSC loc. 7574; 15078.

Figure 17 *Classopollis classoides* Pflug emend. Pocock and Jansonius. Lower Jurassic; Ellesmere Island, GSC loc. 6660; 13769.

Figure 18 *Triancoraesporites communis* Schulz. Norian or Rhaetian; Ellesmere Island, GSC loc. 4905; 13691.

Figure 19 *Klausipollenites vestitus* Jansonius. Norian or Rhaetian; Ellesmere Island, GSC loc. 4905; 13722.

Figure 20 *Vittatina* sp. Upper Permian; Peel River area, N.W.T., GSC loc. 5279; 15079.

Figure 21 *Limbosporites lundbladii* Nilsson. Norian or Rhaetian; Ellesmere Island, GSC loc. 4905; 13689.

Figure 22 *?Taurocusporites* sp. Lower Jurassic; Ellesmere Island, GSC loc. 6660; 13763.

Figure 23 *?Pinuspollenites vancampoi* Danzé-Corsin and Laveine. Lower Jurassic; Ellesmere Island, GSC loc. 6660; 13781.

Figure 24 *Camarozonosporites rudis* (Leschik) Klaus. Norian or Rhaetian; Ellesmere Island, GSC loc. 4905; 13682.

Figure 25 *Vitreisporites pallidus* (Reissinger) Nilsson. Lower Jurassic; Ellesmere Island, GSC loc. 6660; 13775.

Figure 26 *?Leiotriletes lineatus* Bolkhovitina. Lower Cretaceous; Alberta, GSC loc. 5925; 15084.

Figure 27 *Cycadopites* sp. Mid-Permian; Melville Island, GSC loc. 5221; 15080.

Figure 28 *Contignisporites ?cooksonii* (Balme) Dettmann. Lower Cretaceous; Richardson Mountains, GSC loc. 26909; 13835.

Figure 29 *Trilobosporites* sp. Lower Cretaceous; Alberta, GSC loc. 5925; 15081.

Figure 30 *Leiotriletes venustus* Bolkhovitina. Upper Cretaceous; Saskatchewan, GSC loc. 6658; 15082.

Figure 31 *Gleicheniidites senonicus* Ross. Lower Cretaceous; Alberta, GSC loc. 5925; 15083.

Figure 32 *Eucommiidites troedssonii* Erdtman. Lower Cretaceous; Nova Scotia, GSC loc. 6327; 13215.

Figure 33 *Wodehouseia* sp. Late Upper Cretaceous; Devon Island, GSC loc. 4723; 15085.

Figure 34 *?Proteacidites* sp. Late Upper Cretaceous; Devon Island, GSC loc. 4723; 15086.

Figure 35 Unidentified angiosperm pollen. Early Miocene; Vancouver Island, GSC loc. 7572; 15087.

Figure 36 Pollen of Taxodiaceae. Early Miocene; Vancouver Island, GSC loc. 7572; 15088.

Figure 37 Unidentified gymnosperm pollen. Upper Cretaceous; Saskatchewan, GSC loc. 6658; 15089.

Figure 38 Unidentified angiosperm pollen. Late Upper Cretaceous; Devon Island, GSC loc. 4723; 15090.

Figure 39 *Aquilapollenites amplus* Stanley. Late Upper Cretaceous; Devon Island, GSC loc. 4723; 15091.

Figure 40 *Veryhachium* sp. Late Famennian; southern Ontario, GSC loc. 7057; 15092.

Figure 41 *Rugubivesiculites* sp. Upper Cretaceous; Saskatchewan, GSC loc. 6658; 15093.

Figure 42 Pollen of Caryophyllaceae. Pleistocene; Cape Breton Island, GSC loc. 7575; 15094.

Figure 43 *Pinus* sp. Pleistocene; Cape Breton Island, GSC loc. 7575; 15095.

Figure 44 *Baltisphaeridium* sp. Late Famennian; southern Ontario, GSC loc. 7057; 15096.

Figure 45 *Deflandrea* sp. Upper Cretaceous; Ellef Ringnes Island, GSC loc. 7573; 15097.

MARINE PLEISTOCENE FAUNAS

by F. J. E. Wagner

Unlike other periods, the part of the Pleistocene from which fossils have been obtained was not of sufficient duration for evolutionary changes to have taken place. Therefore, the faunal assemblages are of little value stratigraphically. They do, however, show regional differences, as between Atlantic and Pacific, and they are also useful as indicators of water depths and water temperatures in late Pleistocene time.

The areal aspects of the faunas, as shown by the most common species, are set forth in the explanations accompanying PLATES XXXII and XXXIII. Arctic areas were apparently populated mainly through migration of species from the North Atlantic and have few species in common with western Canada. The cosmopolitan species indicate boreal or subarctic water temperatures, rather than high-arctic temperatures. Of the other species illustrated, *Natica clausa, Neptunea lyrata, Nuculana fossa, Nuculana minuta, Portlandia arctica, Astarte borealis, Astarte montagui striata,* and *Mya pseudoarenaria* are particularly typical of boreal temperatures, and *Buccinum cyaneum, Neptunea despecta tornata, Chlamys hindsi, Saxidomus giganteus, Mya arenaria,* and *Balanus hameri* are found in cool temperate waters. In western Canada, faunal assemblages indicate an amelioration of the hydroclimate between about 12,000 to 14,000 years ago and the present. A similar warming trend is shown by the faunas of eastern Canada, but covering the shorter time span of the Champlain Sea (approximately 11,500 years B.P. to between 8,000 and 9,000 years B.P.) Fossil assemblages from the region of Boothia Peninsula to eastern Back River area in the arctic mainland suggest a shallowing of the water from a possible maximum depth of 200 feet for deposits about 23,300 years old that now lie 640 feet above present sea level to a possible minimum depth of 15 to 20 feet for deposits now 70 feet above sea level and about 3,690 years old. Similar changes in depth, that is deeper at first and then becoming shallower, have also been noted for eastern Canada.

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PLATE XXXII. Marine Pleistocene Fossils

(All figures natural size unless otherwise stated)

COSMOPOLITAN[1]

Figure 1 *Serripes groenlandicus* (Bruguière). Grande Rivière du Chêne, Quebec; 20156.

Figures 2, 3 *Hiatella arctica* (Linné). Chevalier, Quebec; 20164.

Figure 4 *Balanus crenatus* Bruguière. Atlantic Coast (Recent); 20191.

Figure 5 *Macoma calcarea* (Gmelin). Richelieu River, Quebec; 20158.

Figure 6 *Mytilus edulis* Linné. St. Joseph du Lac, Quebec; 20149.

Figure 8 *Mya truncata* Linné. Grande Rivière du Chêne, Quebec; 20161.

Figures 11, 12 *Hemithiris psittacea* (Gmelin). St. Nicholas, Quebec; 20131.

EASTERN CANADA

Figure 7 *Lepeta caeca* Müller. St. Nicholas, Quebec; 20132.

Figure 9 *Mya arenaria* Linné. St. Remi, Quebec; 20160.

Figure 10 *Balanus hameri* (Ascanius). St. Philomène, Quebec; 20192. (*See also* Pl. XXXIII, fig. 6).

Figure 25 *Tethya logani* Dawson. Montreal, Quebec; 20130. (x2)

Figure 31 *Neptunea despecta tornata* (Gould). Grande Rivière du Chêne, Quebec; 20140.

[1] All areas of marine submergence in eastern Canada (Champlain Sea submergence in Ontario and Quebec plus the Atlantic Provinces), Tyrrell Sea submergence (ancestral Hudson Bay), Arctic regions and western Canada (coastal British Columbia).

EASTERN CANADA AND ARCTIC CANADA

Figure 13 *Buccinum tenue* Gray. Ste. Geneviève, Quebec; 20138.

Figure 14 *Buccinum cyaneum* Bruguière. St. Janvier de Joly, Quebec; 20137.

Figure 15 *Trichotropis borealis* Broderip and Sowerby. Montreal, Quebec; 20134. (x2)

Figure 16 *Balanus balanus* (Linné). Atlantic Coast (Recent); 20190.

Figure 19 *Cylichna alba* Brown. Montreal, Quebec; 20143. (x2)

Figure 20 *Tachyrhynchus erosum* (Couthouy). St. Nicholas, Quebec; 20133.

EASTERN CANADA, ARCTIC CANADA, AND TYRRELL SEA

Figure 17 *Protelphidium orbiculare* (Brady). Montreal, Quebec; 20126. (x55)

Figure 18 *Islandiella teretis* (Tappan). Alexandria, Ontario; 20112. (x45)

Figures 21, 22 *Portlandia arctica* (Gray). Ste. Geneviève, Quebec; 20148.

Figures 23, 24 *Nuculana pernula* (Müller). Ste. Geneviève, Quebec; 20145.

Figure 26 *Mya pseudoarenaria* Schlesch. St. Janvier de Joly, Quebec; 20163a.

Figures 27, 28 *Macoma balthica* (Linné). St. Joseph du Lac, Quebec; 20157.

Figures 29, 30 *Astarte montagui striata* (Leach). St. Janvier de Joly, Quebec; 20151.

PLATE XXXIII. Marine Pleistocene Fossils

(All figures natural size unless otherwise stated)

EASTERN CANADA, ARCTIC CANADA, AND WESTERN CANADA

Figure 4 *Natica clausa* Broderip and Sowerby. Ste. Geneviève, Quebec; 20135.

Figure 5 *Natica pallida* Broderip and Sowerby. Ste. Geneviève, Quebec; 20136.

ARCTIC CANADA AND TYRRELL SEA

Figures 11, 12 *Astarte borealis* Schumacher. Bathurst Island, Northwest Territories; 20997.

ARCTIC CANADA AND WESTERN CANADA

Figure 13 *Neptunea lyrata* (Gmelin). Vancouver Island, British Columbia; 20998.

Figure 15 *Nuculana minuta* (Fabricius). Whatcom co., Washington, U.S.A.; 20999. (x2)

WESTERN CANADA

Figure 1 *Saxidomus giganteus* (Deshayes). Point Roberts, British Columbia; 21005.

Figures 2, 3 *Macoma incongrua* (Martens). Surrey municipality, British Columbia; 21006.

Figures 7, 8 *Protothaca staminea* (Conrad). Redondo Beach, California, U.S.A.; 21004.

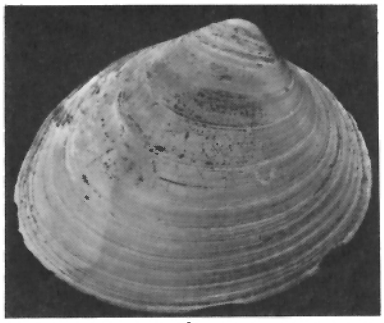
Figures 9, 10 *Chlamys hindsi* (Carpenter). Point Roberts, British Columbia; 21002.

Figure 14 *Trichotropis cancellata* Hinds. Surrey municipality, British Columbia; 21000. (x2)

Figure 16 *Nuculana fossa* (Baird). Vancouver Island, British Columbia; 21001.

Figure 17 *Clinocardium nuttalli* (Conrad). Point Roberts, British Columbia; 21003.

Figure 6 *Balanus hameri* (Ascanius). Eastern Canada (*see* Pl. XXXII, fig. 10).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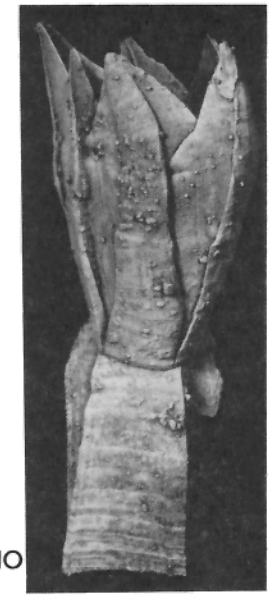
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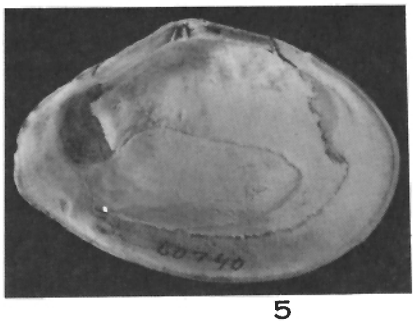
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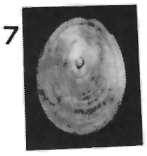
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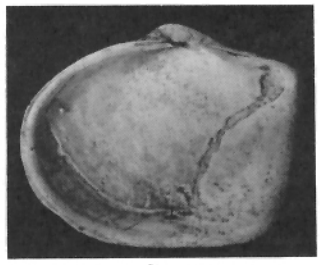
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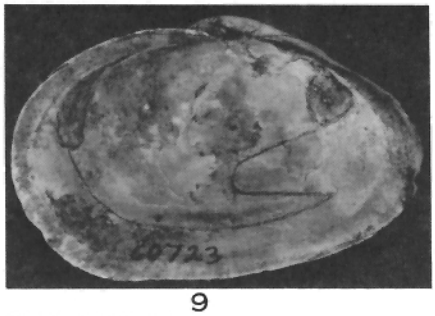
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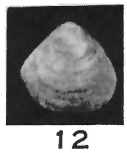
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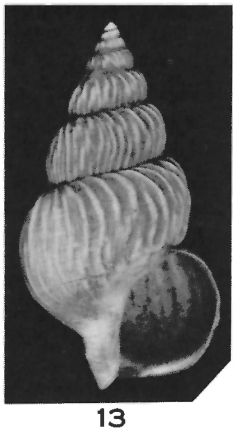
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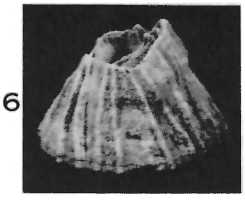
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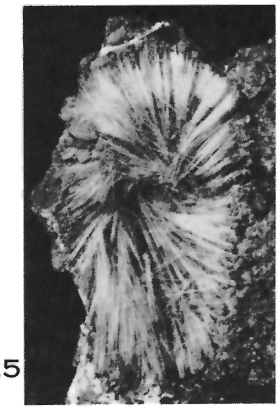
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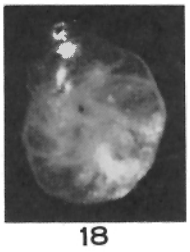
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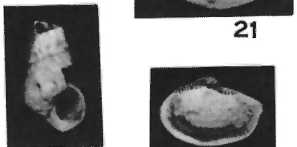
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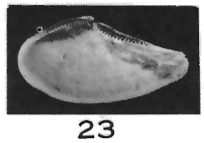
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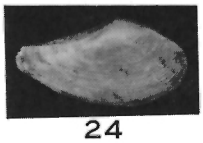
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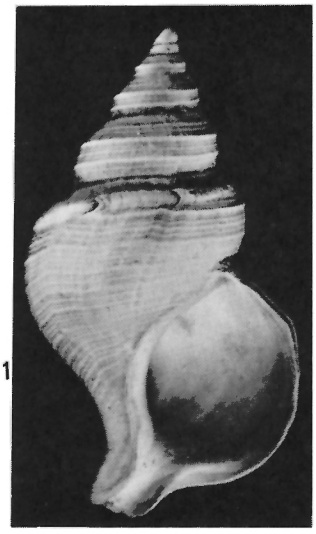
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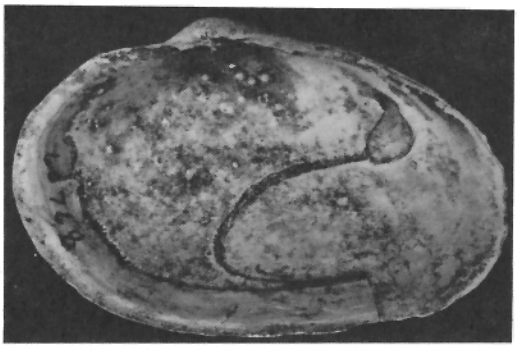
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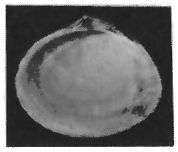
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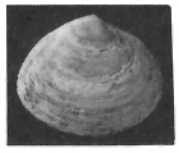
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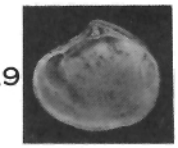
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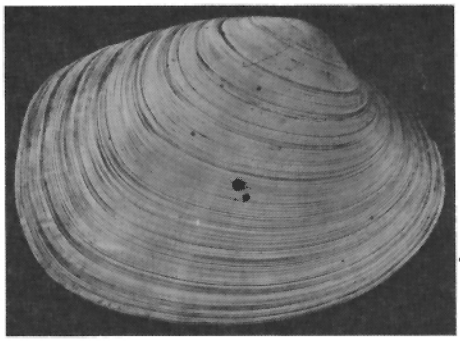
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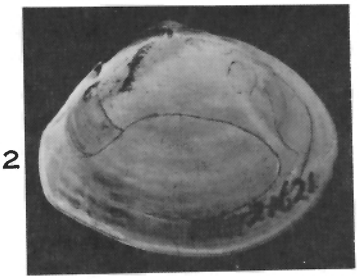
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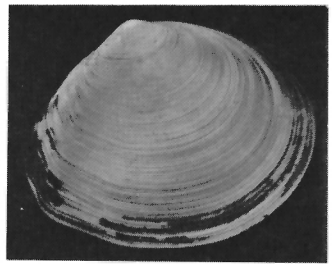
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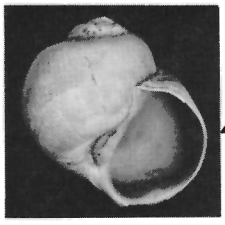
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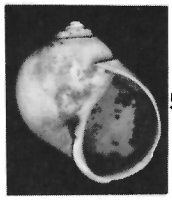
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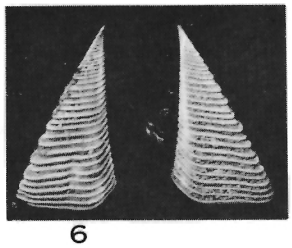
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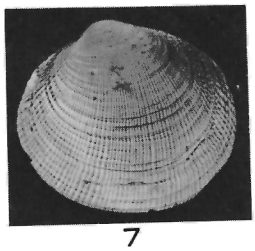
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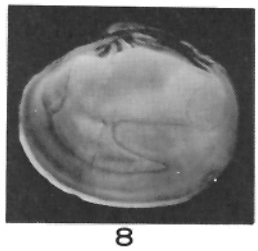
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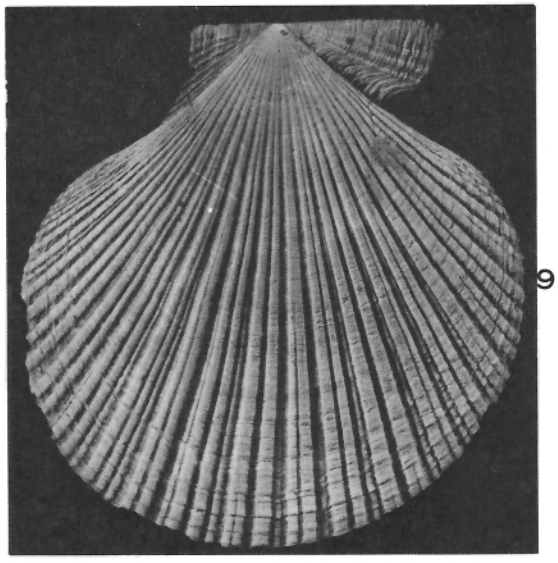
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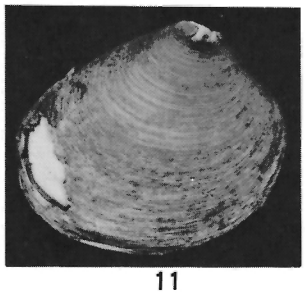
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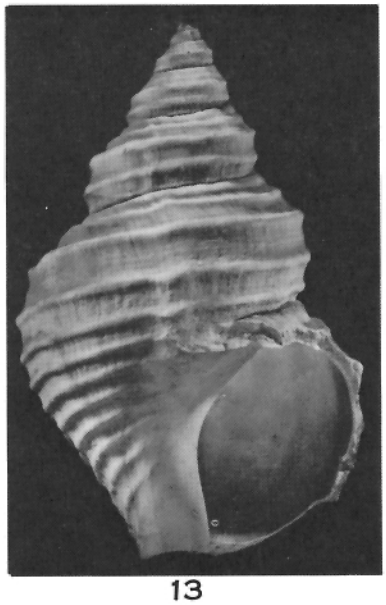
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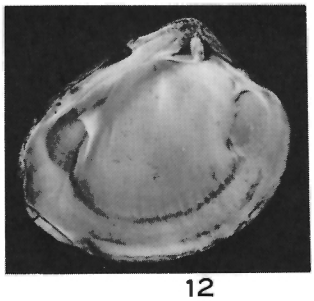
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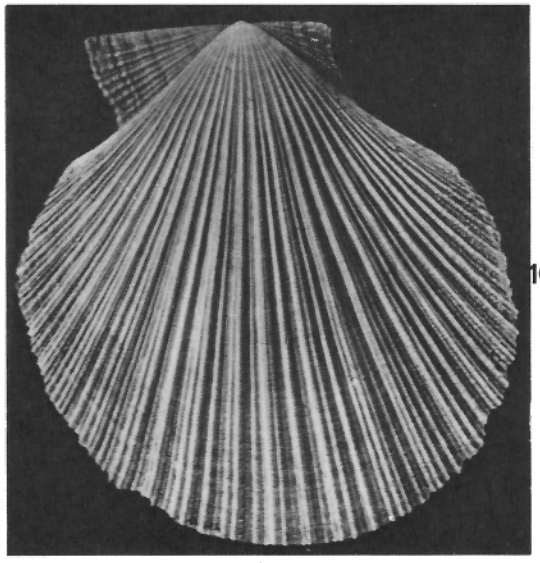
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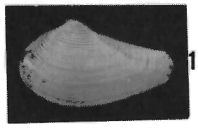
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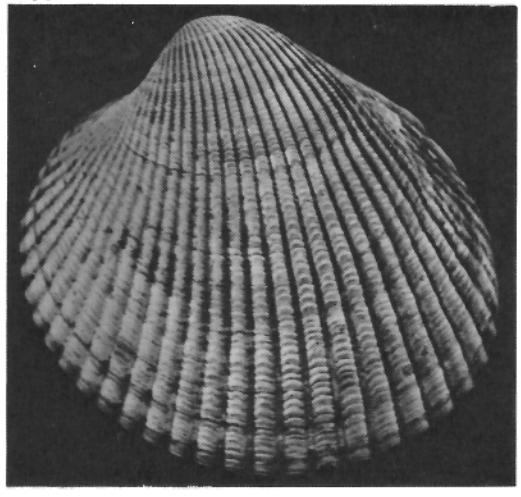


XII. Quaternary Geology of Canada

Introduction 676

Older Glacial and Non-Glacial Record 679

Classical Wisconsin and Postglacial Events 705

Economic Considerations 756

Selected References 758



INTRODUCTION

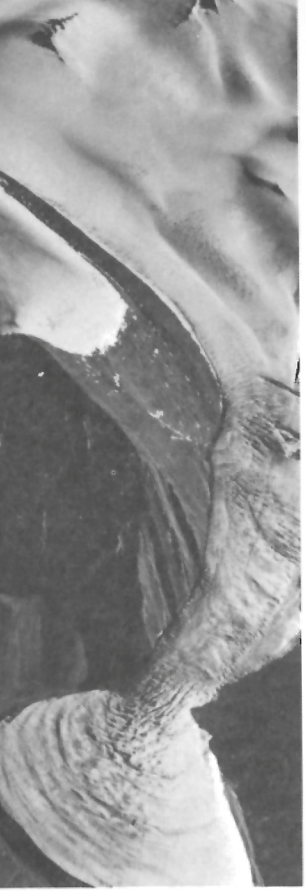
The term Quaternary refers to about the last million years of the earth's history and is set apart from the Tertiary period by reason of the climatic changes that gave rise to successive glaciations of vast regions, and to a general lowering of snowlines throughout the world. Some recent oceanographical and paleontological data suggest that the climatic changes actually may have begun as early as 3 million years ago.

The geology of the glacial and non-glacial deposits that mantle the bedrock and various aspects of the landscape that are attributable to Quaternary events are discussed in this chapter. The term Quaternary in the writer's opinion, may be interchanged with Pleistocene as applied to Canada. This term is frequently used in Canada as it more readily conveys the glacial and climatic connotations, and accordingly is generally used here. The Pleistocene also includes the present, very short, non-glacial interval which is commonly termed Recent. In Canada, Recent may be regarded as comprising the last 7,000 years—the period following dissipation of the major part of the last mainland ice sheet.

At present, about 10 per cent of the earth's land surface is covered by glacier ice whereas during former glaciations as much as 30 per cent was under ice and permanent snowfields. About 97 per cent of Canada has been glaciated, and hence this country contains more glaciated terrain than any other; at present about one per cent remains under glacial cover—in the Queen Elizabeth Islands, Baffin Island, and in the mountains of western Canada. An area of about 70,000 square miles in the western Yukon, in the shadow of the coastal mountains, escaped glaciation altogether. Two elongate areas along the mountain front west of Mackenzie River in Northwest Territories, comprising about 4,000 square miles, are also thought to be partly unglaciated, but have been little studied. A small area in the Foothills of southwestern Alberta stood higher than the adjacent interior ice sheet, but harboured elongate valley glaciers from the mountain ice on its western side. Parts of the southern Interior Plains close to the International Boundary and near the southern terminus of the interior ice sheet apparently stood higher than the surrounding glaciers and escaped glaciation.

The organic remains preserved in the surficial deposits are mainly those of modern species. Study of their stratigraphic position and geographic location relative to their present distribution reveals information on former migration of plants and animals. Extinction and evolution of species occurred in the Quaternary, but not to the same degree as in earlier periods.

Man made his appearance during the Quaternary and his development has been controlled to a large degree by the climatic conditions so characteristic of the period. It is generally believed that it was the lowering of sea level, consequent upon the amount of water incorporated in continental ice, that allowed Asiatic tribes to migrate to



XII
Quaternary Geology of Canada

V.K. Prest

Viking Ice Cap and glaciers, Ellesmere Island, Northwest Territories.

North America, mainly by land, via the Bering Sea land-bridge (north of Aleutian Islands chain). Later, as the glaciers waned, these people left the unglaciated *refugia* in Alaska and Yukon Territory, migrating eastward along the arctic coast and southward into the interior of the continent. Some of these people—the early North American Indians—camped along glacial lakes and spillways that have long since disappeared.

The Quaternary includes four major glaciations each of which occupied a period of about 100,000 years. From oldest to youngest these are known as Nebraskan, Kansan, Illinoian, and Wisconsin. They were separated by longer-term, interglacial intervals—the Aftonian, Yarmouth, and Sangamon—when climates were as warm or warmer than now and when the continent must have been largely ice free. Under such conditions a rich and varied flora and fauna must have occupied most parts of Canada. Few deposits, however, can be assigned with certainty to these interglacial intervals though all are probably represented in some known occurrences. Interglacial sediments buried by glacial drift have been reported from the Maritimes, Great Lakes, Interior Plains, Cordillera, and Arctic; all the intervals must be represented in the stratigraphy of Porcupine Plain in northern Yukon Territory, though this has as yet been little studied.

Each of the four major glaciations was interrupted by relatively short-term non-glacial intervals or interstades. These were times when the climate ameliorated and the glaciers receded extensively from peripheral zones but presumably continued to occupy part of the mainland. Plants and animals tolerant of a cool climate migrated into the newly deglaciated areas, only to have their progeny displaced by the re-advancing glaciers at a later date. In Canada such interstades have been recognized only within the Wisconsin Glaciation. The best known occurrences are in Lake Erie region of Ontario and around Strait of Georgia, British Columbia.

There is no generally accepted classification scheme for Wisconsin age sediments, but they may be subdivided on the basis of glacial and non-glacial deposits. By means of radiocarbon (C^{14}) dates of organic materials contained in some deposits, reliable correlation between scattered sites dating back to about 45,000 years have been made, but the detailed geological and biological studies necessary to support the age determinations are usually lacking. Reliable age-datings are generally restricted to the last 25,000 or 30,000 years B.P.[1]. Investigations in the Lake Erie region indicate a mid-Wisconsin interstade occupying a period of about 20,000 years. This has enabled subdivision of the Wisconsin Glaciation into two main stages—early and late. The latter is commonly termed the 'Classical' Wisconsin. Fluctuations of the ice fronts during the mid-Wisconsin interstade complicate the stratigraphic record. This interstade was brought to a close by a major ice-frontal advance

[1] B.P., "Before Present", refers to the year 1950. This is to be understood in connection with all datings that follow, though it will not always be repeated.

during which the glaciers returned to the vicinity of the limit reached during the earlier Wisconsin Glaciation. The late or Classical Wisconsin climax occurred about 20,000 years ago, and as mentioned earlier most ice had disappeared from the mainland by 7,000 years B.P. The time transgressive stratigraphic relationships recognized

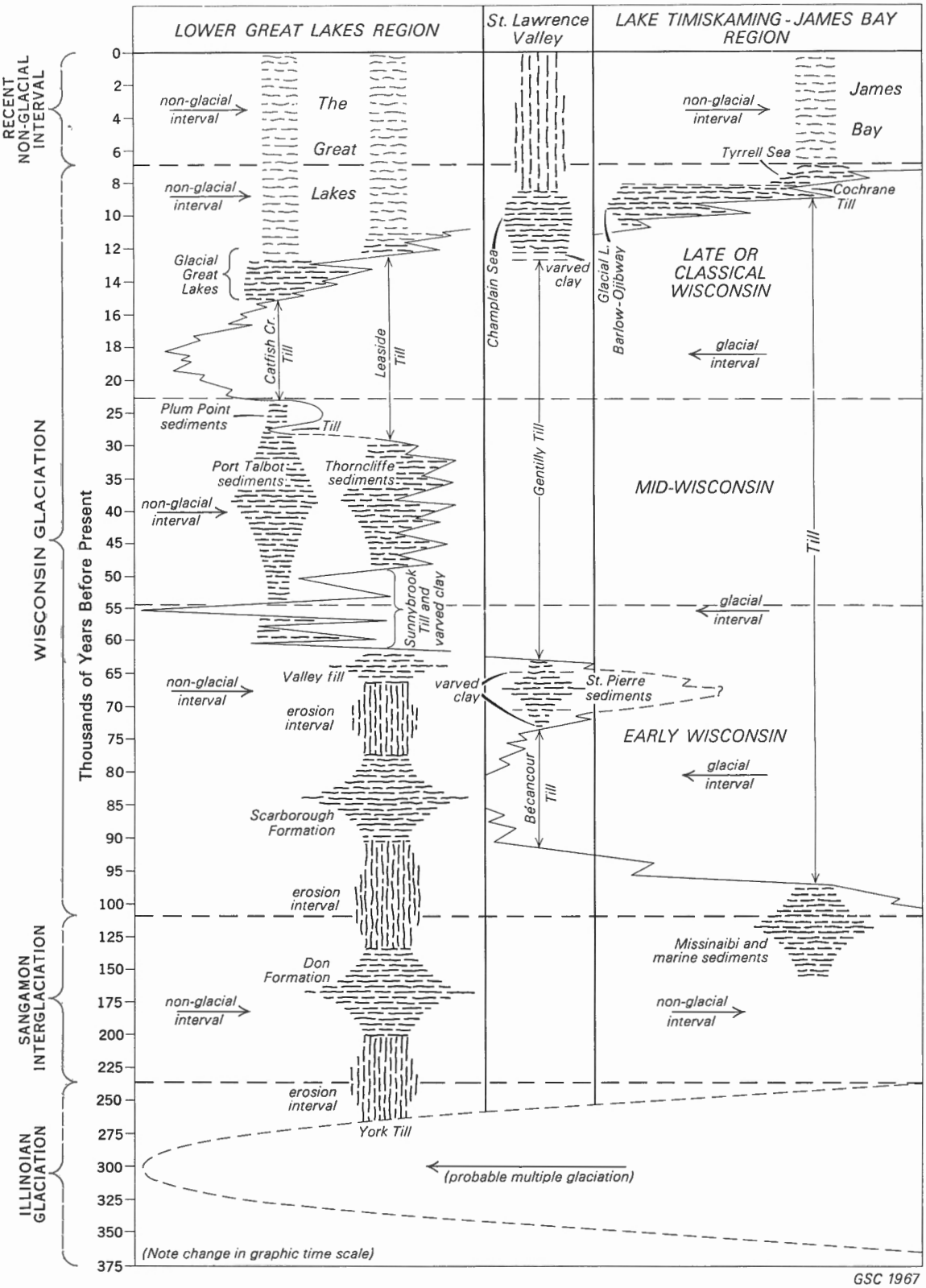


FIGURE XII-1
Time-stratigraphic relationships of late Pleistocene in east-central Canada.

in east-central Canada during the late Quaternary are illustrated diagrammatically in Figure XII-1; similar relationships are no doubt applicable to the whole of the Quaternary and to all parts of Canada.

Because there are uncertainties regarding Wisconsin events and their implications in different parts of Canada, and because many samples submitted for radiocarbon analyses have been given "greater than" datings, it is not yet possible to discuss interglacial deposits separately from Wisconsin, or other interstadial deposits. Furthermore, because of the general lack of paleontological or other data sufficiently accurate to identify specific interglacial deposits, the associated till sheets cannot be identified. Thus all deposits that predate the climax of the late, main, or Classical Wisconsin are discussed together, followed by an account of the last glaciation, and of deglaciation. In some parts of the Arctic and the Cordillera, glacial conditions have prevailed to the present and are manifested in the form of ice caps and mountain glaciers; for convenience, fluctuations of these ice masses are discussed along with Wisconsin deglacial events though strictly speaking they are post-Wisconsin.

THE OLDER GLACIAL AND NON-GLACIAL RECORD

This section deals with the Canadian record of the Quaternary prior to the climax of the Classical Wisconsin Glaciation; it includes glacial, interglacial, and interstadial deposits and the events that caused them or otherwise modified the landscape. Locally the deposits may be preglacial and represent events occurring in the earliest Quaternary.

Organic remains form a small but vitally important part of the sedimentary record, and are given considerable attention here. As a result of recognition of the environment and age of the non-glacial sediments the glacial sediments above or below them may be tentatively assigned to a particular glaciation. The deposits themselves and pertinent aspects of the physiography are treated on a regional rather than temporal basis.

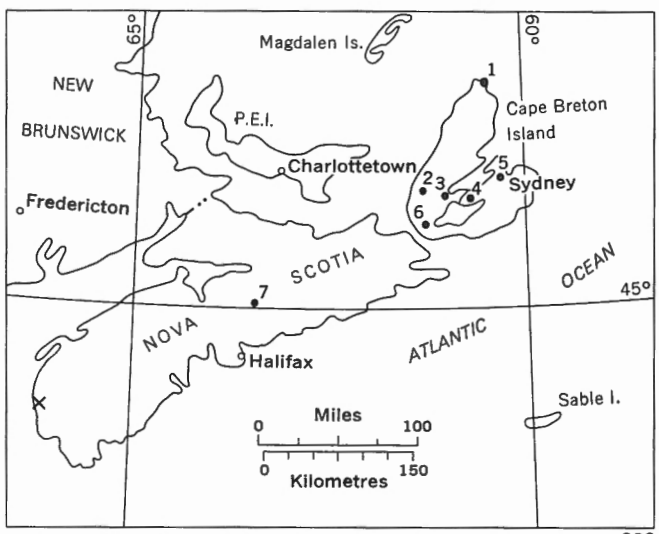
Appalachian Region

The physiography of the Appalachian Region reflects some aspects of the older glacial and non-glacial intervals but these are, in general, effectively masked by the more obvious fresh landforms and deposits of the last glaciation. Numerous estuaries represent the drowned parts of former river systems that may have been operative during times of lower sea level in Quaternary time. R. H. MacNeill (Acadia Univ.) believes many of the streams in Nova Scotia are re-excavated preglacial channels, and also that some till-mantled coastal terraces were probably cut during preglacial times. A grey sandstone regolith is found in parts of eastern New Brunswick both at the surface and beneath a mantle of red, sandy till. It is formed of soft Pennsylvanian sandstone and siltstone, up to 25 feet thick, and argues against extensive modification of much of the topography at this time. The highlands of New Brunswick show surprisingly little evidence of glaciation; erratics appear to be absent over large areas and the bedrock is soft and deeply weathered.

Interglacial and interstadial deposits together with associated glacial deposits that predate the Classical Wisconsin have been recognized on Cape Breton Island, and one deposit is known on the Nova Scotia mainland; none has been recognized in Newfoundland, Prince Edward Island, New Brunswick, or the Gaspé.

Buried Organic Deposits (Fig. XII-2)

Bay St. Lawrence. Unconsolidated materials on the northeastern end of Cape Breton Island form a seacliff up to 150 feet high and comprise stratified sediments, with some organic materials, between stony till-like layers. Towards the southwest, bevelled bedrock is exposed beneath the drift. Up to 20 feet of hard-packed sand-gravel 'till' occurs at the base of the drift section. This is overlain by as much as 4 feet of interbedded fine gravel and silt, or by a boulder layer believed to be a washed-surface of the till. The stratified unit and boulder layer is overlain by a dense organic layer, a few inches to a foot thick, followed by 2 feet of stratified silt and as much as 30 feet of interbedded sand and gravel grading upward into cobble gravel. The silt beneath the organic layer was



1. Bay St. Lawrence 2. Hillsborough 3. Whycocomagh 4. Benacadie 5. Leitches Creek 6. Inhabitants River 7. Milford
× Old' shell site: shells in till, Cape St. Mary

FIGURE XII-2. Location of buried organic deposits in Nova Scotia.

found to contain spores derived solely from Mississippian rocks. The organic bed represents a detrital sediment and contains pollen of alder, birch, black and white spruce, jack pine, balsam fir; a trace of juniper, willow, blue beech and walnut; and a variety of shrubs, herbs, grasses, ferns, mosses, and fungi (Mott and Prest, 1967). A piece of wood identified as tamarack has yielded a radiocarbon dating of >38,300 years (GSC-283)[1]. The immediately overlying silt beds contain similar pollen and spores but higher strata are barren. Nearby, E. H. Muller noted a lens of tan, silty clay, 12 feet thick at its maximum and 200 feet long. Overlying this lens and the cobble gravel (reported above) is a sandy boulder till some 30 to 90 feet thick, which becomes increasingly coarser upward; boulders as much as 8 by 5 feet occur near the surface.

In the clay lens are fragments of *Megayoldia thracaeformis,* a marine mollusc inhabiting waters colder than those surrounding Cape Breton today. The clay also contains pollen of alder and birch indicative of a cool climate but with pine, some oak, and a trace of basswood; otherwise the assemblage is similar to that of the detrital organic layer. Dinoflagellate cysts of Quaternary age and spores from the nearby Mississippian rocks are also present. The fossil record and radiocarbon date on the buried organic materials clearly indicate a pre-Classical Wisconsin cool period; this may be very early Wisconsin or perhaps late Sangamon.

Hillsborough. In southwestern Cape Breton Island between Mabou and Hillsborough from 8 to 12 feet of dull red, compact, clayey till overlies some 5 feet of stratified silty to clayey sediments that contain streaks and thin lenticular beds of carbonaceous material, and a basal layer of peat, silt, and wood a few inches to two or more feet thick. The organic layer rests on up to 18 inches of silt which in turn rests on an uneven ortstein layer developed on the surface of a highly oxidized sand and gravel more than 11 feet thick. The base of the exposed section at road level is about 10 feet above the level of the river and the sea.

The silty sediments and the peat bed (Mott and Prest, 1967) contain five pollen zones with a different assemblage predominating in each. The whole assemblage indicates a forest cover similar to that of the Boreal Forest Region rather than that of Cape Breton today. Wood from the base of the peat layer is dated at >51,000 years (GSC-370). An early Wisconsin or other interstadial interval is indicated but correlation with the Bay St. Lawrence site is not justifiable.

Whycocomagh. Buried organic beds in a highway-cut in the village of Whycocomagh, Cape Breton Island, were re-examined by the writer following an anomalous radiocarbon dating obtained on wood from this site. Gravelly till-like material 10 feet thick overlies 5 feet of stratified sediments ranging from fine gravel to silt and including a few inches of silty peat with scattered wood. The organic layers rest on 15 inches of partially oxidized clay and silt overlying 10 to 16 feet or more of stony, clay till that rests on an irregular bedrock surface. The stratified sediments extend for more than 50 feet, pinching out to the west as both bedrock and till mantle rise towards the surface. Pollen in the organic layer (Mott and Prest, 1967) is characterized by an assemblage dominated by alder, birch, spruce, and pine, and similar to part of the Hillsborough pollen diagram. Wood from the organic layer has an age of >44,000 years (GSC-290). The sites are only 15 miles apart at opposite ends of a through-going valley with a low drainage-divide; both sites lie close to sea level. They are believed correlative.

Benacadie. Intertill stratified sediments occur in a shore-cliff south of Benacadie at the entrance to East Bay, Bras d'Or Lake. Some 50 feet of reddish, stony, clay till overlies 5 feet of well-bedded clayey silts and a lower 20 feet of sand-gravel till. The clay-silt beds have a high percentage of pollen of pine (probably jack pine) and sedge, with lesser amounts of birch, alder, and grass (Mott and Prest, 1967). The assemblage has more affinities with the Bay St. Lawrence pollen diagrams than with those of the nearer Whycocomagh and Hillsborough sites. Another occurrence of intertill stratified sediments was found by Terasmae and Mott near Derby Point 4 miles northwest of Benacadie. Beneath the level of the shore road, 2 feet of reddish, clayey till overlies a few feet of silty clay, sand with plant detritus, and sandy gravel that rest on about 10 feet of slumped till down to sea level. The intertill silty clay did not carry pollen.

Leitches Creek. Part of a drill-log and some samples from a borehole at Leitches Creek, west of Sydney, that penetrated some 190 feet of overburden, indicate the presence of two layers of organic-bearing sediments each underlain and overlain by till. Preliminary pollen studies suggest that the lower organic layer is of interglacial age and that the upper layer correlates with the Hillsborough interstadial interval.

Inhabitants River. Buried organic deposits were reported in 1868 by J. W. Dawson from Inhabitants River, Cape Breton Island. A hard, peaty bed with roots and branches of coniferous trees rests on grey clays and underlies about 20 feet of till. The site has not been relocated.

Milford. A unique deposit of buried organic material found as overburden was removed at the gypsum quarry 2½ miles south of Milford Station, Nova Scotia. Beech nut *(Fagus* sp.), hickory nut *(Carya* sp., cf. *C. aquatica),* bayberry seed *(Myrica pennsylvanica),* and a beaver-cut stick were collected from a sinkhole in the gypsum. The

[1] These letters refer to the radiocarbon laboratory responsible for the age dating and the number is the sample reference number. GSC, Geological Survey of Canada; Gx, Geochron. Laboratories; Gro and GrN, Groningen, Netherlands; I, Isotopes Inc.; L, Lamont Geological Observatory; S, University of Saskatchewan; Y, Yale University; W, U.S. Geological Survey.



PLATE XII-1

Karst topography on gypsum at site of interglacial deposits, Milford, Nova Scotia. The sinkholes contained a mixture of glacial and non-glacial sediments including organic materials dated at >38,000 years B.P. The gypsum surface is ice scoured and overlain by a complex of tills and fossiliferous sediments visible in bluff in background.

wood was dated at >33,800 years (GSC-33). The following notes on the quarry site and its environs are based on observations by W. Take (Nova Scotia Museum).

The surface of the gypsum displays a karst topography (Pl. XII-1). In the base of depressions there is usually a mixture of slumped till and glaciofluvial materials, overlain by unfossiliferous brown to grey clay and sphagnum peat containing gastropods and patches of slumped till. Overlying the peat unit is highly fossiliferous grey clay and sandy clay. The thin basal part of the clay is characterized by abundant macrofossils of white pine and rare hemlock; the latter increases upward and is evenly distributed. Logs, beaver-sharpened sticks, cones, insects, mollusca, amphibian and mammalian remains were collected from the stratified clays. An erosional unconformity separates the clay from plant-bearing sands—mainly spruce, fir, and rare white pine cones.

The sinkhole deposits and the glacially scoured surface of the bedrock are evenly truncated and overlain by a compact, grey, gypsiferous till. In places this till carries woody detritus; in two places it has a boulder pavement developed on it. The till is generally overlain by sediments grading from unfossiliferous grey clay through fossil-bearing silty clay, into highly fossiliferous grey sand with abundant twigs of coniferous trees as well as cones of spruce and rarely of fir. Overlying these sediments are two very similar tills, separated in valley bottoms by orange-brown gravelly sands. The basal part of each till, and especially of the lower one, is gypsiferous. The tills change from grey near the base to grey-brown at the top. In upland sections the combined till is a uniform brown colour, does not contain gypsum but is deeply oxidized. At one locality it has a well-developed soil-profile.

Lying on a grey-brown till about 35 feet above sea level in some valleys is an orange-brown to red-brown laminated clay with vertical rootlets in its upper part suggestive of a marsh environment. At somewhat higher elevations, a black clay carries mollusc remains, white pine cones, hemlock bark and cones, and red oak acorns. A distinctive red-brown clay till overlies the stratified sediments in low areas, and the grey-brown tills on higher ground. This is followed by a complex of red clays, red till, minor glaciofluvial sand and gravel, and a thin sand-gravel till unit, the last strictly confined to the valleys. Glaciofluvial sand and gravel, and stratified clay and silty clay, both with the modern soil profile or a cultivated surface, complete the sequence of deposits.

Take considers the Milford deposits to represent parts of the Kansan, Illinoian, and Wisconsin Glaciations, as well as the Yarmouth and Sangamon interglacial intervals. Certainly interglacial as well as interstadial fossil assemblages are indicated, but the implications of the karst topography have precluded definite correlations.

St. Lawrence Lowlands

Major features of the physiography of the St. Lawrence Lowlands, such as Niagara Escarpment and the bordering highlands of the Canadian Shield and Appalachian Region, predate the Quaternary and, except for local over-deepening of valleys and some glacial scour, were little changed by glaciation. The great morainal belts in southwestern Ontario bear testimony to the direct effects of glaciation. Drift thickness in the Oak Ridges interlobate moraine reaches over 1,000 feet, and there is some evidence of a buried drainage system north of Lake Ontario beneath what is now a major ridge overlooking the lake. The St. Lawrence and Ottawa River valleys are filled by 100 to 200 feet of marine sediments, and in an area of high bedrock knobs north of Oka-sur-le-Lac a channel containing about 400 feet of drift rests on rotted bedrock some 175 feet below present sea level. Thus the flat valley bottoms bear little resemblance to the more mature drainage systems of interglacial and preglacial times.

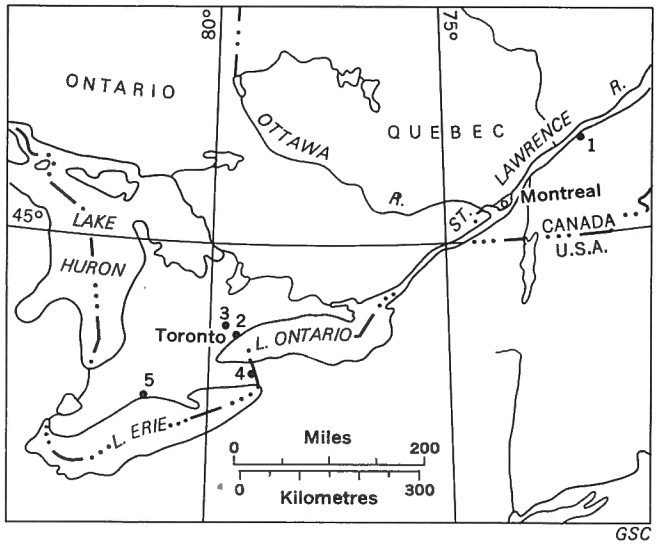
In the St. Lawrence Lowlands the only undoubted interglacial deposits are those of the Toronto area where the Don Formation is assigned to the Sangamon interglacial interval, and the basal York Till is thought to represent an Illinoian glaciation (Karrow, 1967). Near Trois-Rivières, Quebec, and Toronto and London, Ontario,

numerous occurrences of buried organic materials appear to represent non-glacial intervals within the span of the Wisconsin glacial period.

Buried Organic Deposits (Fig. XII-3)

Trois-Rivières. The Quebec part of the St. Lawrence Lowlands near Trois-Rivières has been studied in some detail by Gadd (1960, *in press*). The St. Lawrence Valley drainage system appears to have been blocked in very early Wisconsin time, giving rise to a glacial lake and deposition of reddish varved clays (Table XII-1). These clays were later overridden by the glacier and the Bécancour Till, a brick-red, somewhat sandy, clay till, was deposited. This till is known over a wide area south of the St. Lawrence, but the southern limit reached by the glacier is unknown. The till derives its rich red colour from the Ordovician, Queenston, red shales. Most stones in the till, however, are Precambrian types and hence the glacier that deposited the till came from the Canadian Shield, presumably from the Laurentian Highlands. However, neither the red till nor a correlative grey till has been recognized north of St. Lawrence River.

The type section of the St. Pierre sediments is near St. Pierre les Becquets, Quebec (Gadd, 1960). The name 'St. Pierre' is given to the intertill sediments and to the interval during which they were deposited. The sediments overlie the red Bécancour Till and consist of sand and silty sand with lenticular and discontinuous beds of highly compressed peat and some disseminated organic matter. The sand unit has a maximum thickness of 25 feet, but at the type section is only 13 feet thick and includes three layers of peat. The upper and thickest peat bed where observed along a ravine near St. Pierre les Becquets was 1.75 feet thick and was traced for over 200 yards. Each peat layer begins with a gyttja and grades upward through *Carex* peat into *Sphagnum* peat with abundant tree and other plant remains, the whole suggestive of periodic flooding of the lowlands.



1. Trois Rivières area
2. Toronto
3. Woodbridge
4. Niagara Falls
5. Port Talbot

FIGURE XII-3. Location of buried organic deposits in St. Lawrence Lowlands.

The St. Pierre sediments occur for about 50 miles along the south side of St. Lawrence River from Pierreville to Deschaillons and inland for about 20 miles to Ste. Brigitte (Pl. XII-2). They have also been found on the north side of the river, at Les Veilles Forges, about 6 miles northwest of Trois-Rivières (Gadd and Karrow, 1960). According to P. F. Karrow, it is very likely that the organic-bearing sediments reported by A. P. Coleman at Donnacona 50 miles to the east are the equivalent of the St. Pierre sediments. Some peaty sediments of uncertain stratigraphic position and non-organic sediments between



PLATE XII-2

Section of postglacial sediments, Gentilly Till, and St. Pierre sediments on Bécancour River, north of Aston Junction, Quebec. Sediments are probably underlain by red Bécancour till exposed downstream. (A) St. Pierre sand and pebbly sand >60 feet, (B) Gentilly Till 15 feet, (C) Champlain Sea clay 24 feet, (D) alluvial sand 6 feet.

two tills elsewhere in the region suggest that the St. Pierre sediments are even more widespread. The whole appears to represent an old fluvial system that is slowly being exposed by the St. Lawrence River and its tributaries.

Palynological studies of St. Pierre sediments reveal that spruce, pine, birch, and alder are the main tree types present, and that oak, beech, maple, elm, ash, and hickory together form only 2 to 5 per cent of the tree pollen in the middle and warmest part of the sequence. Hemlock is noticeably absent. The pollen spectrum is similar to that of the early postglacial assemblages in the St. Lawrence Lowlands, except for the absence of hemlock, and it is similar to that of the boreal forest today (Terasmae, 1958, p. 20). Beetle wings are fairly common in the peat beds and a few ostracods have been noted. Terasmae concludes that the climate remained fairly constant throughout most of the St. Pierre interval with sub-arctic conditions prevailing near the bottom and top, and that a relatively short span of time is involved.

Samples from Pierreville and St. Pierre were dated at Groningen by an isotopic enrichment method at 67,000 ± 2,000 years and 64,000 ± 1,000 years, and a sample from Donnacona by conventional techniques at >44,470 years (Y-463). Thus, St. Pierre sediments are inferred to represent an early Wisconsin non-glacial interval. The St. Pierre sediments are overlain by the Lake Deschaillons varved clay, 70 feet thick, deposited in a proglacial lake; these in turn are overlain by Gentilly Till deposited by the advancing ice sheet. This ice cover persisted throughout the greater part of Wisconsin time until final recession of the ice sheet and incursion of the sea into St. Lawrence

TABLE XII-1 *Composite section of Pleistocene deposits, St. Lawrence Lowlands (by N. R. Gadd)*

	(Max. known thickness, feet)
Bog deposits	20
Low terrace sands, el. ca. 100′	10
High terrace sands, el. ca. 300′	10
Champlain Sea sand	25
St. Narcisse Till	15+
Champlain Sea clay	100
Gentilly Till	15
Lake Deschaillons varved clay	70
St. Pierre sediments	25
Red-banded varved clay	5
Bécancour Till	55
Red varved clay	5
Bedrock (Queenston red shale)	—

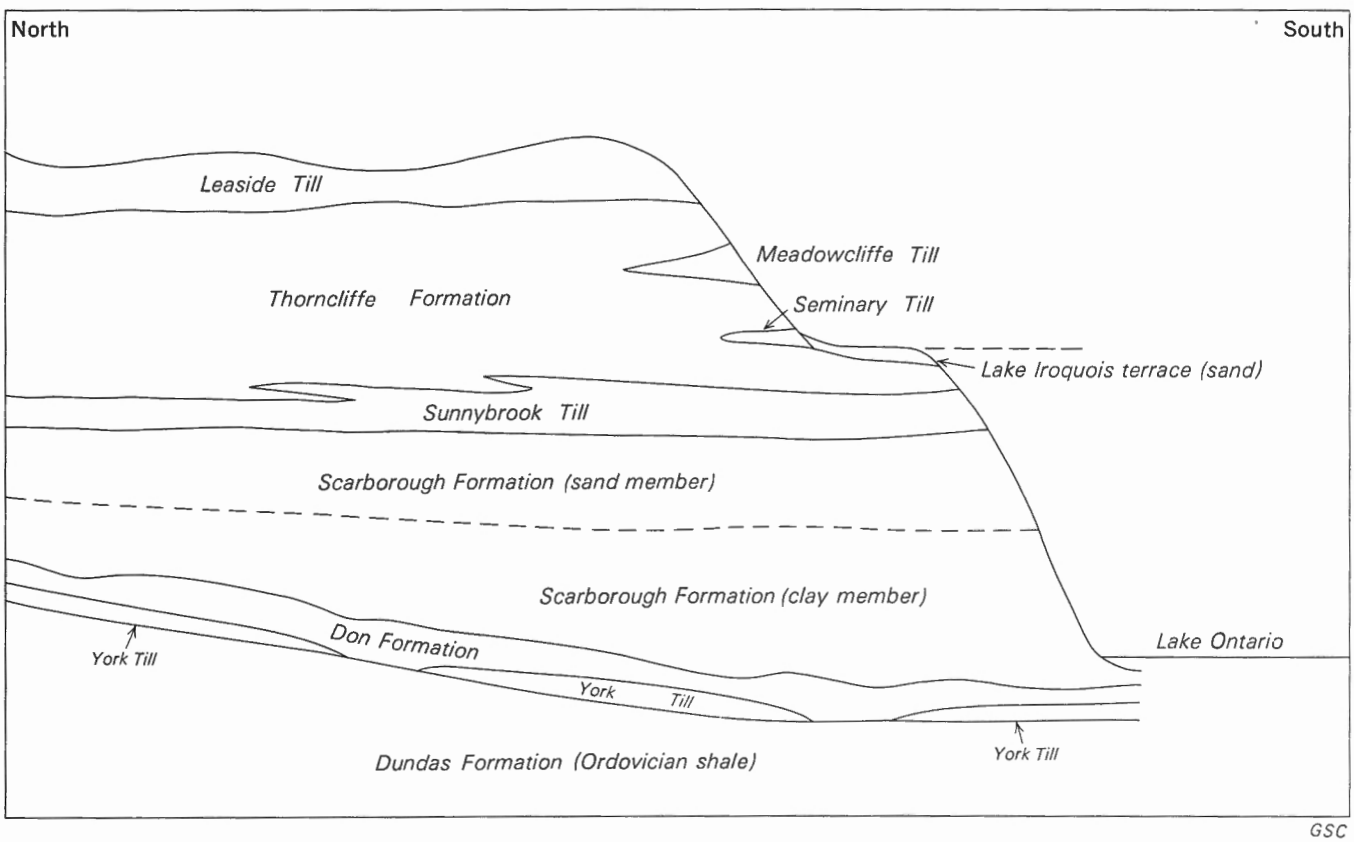


FIGURE XII-4. Schematic diagram showing stratigraphic relationships of Pleistocene deposits at Scarborough Bluffs, Toronto, Ontario (after Karrow, 1964).

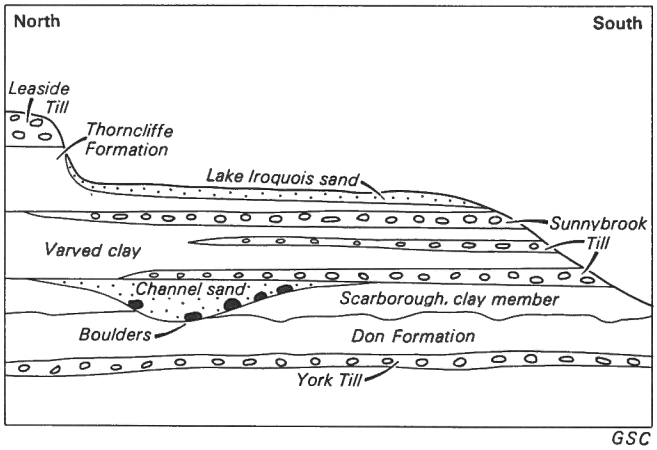


FIGURE XII-5. Schematic diagram showing stratigraphic relationships of Pleistocene deposits in Don Valley brickyard, Toronto, Ontario (after Karrow, 1964).

Lowlands almost 12,000 years ago. Subsequently, differential uplift of the land caused regression of the sea, and the present drainage system was established.

Toronto. The most famous buried, non-glacial Pleistocene deposits in Canada are those of the Toronto area. They have been known for over a century and have been under more or less continual observation in the Don Valley brickyard operations since 1889. Recent studies have done much to clarify the chronology recorded by the complex deposits. The non-glacial organic-bearing beds have been encountered in the extensive subway excavations and other construction projects of the last decade. The sequence of non-glacial deposits in the region is shown schematically in Figures XII-4 and 5.

In the Don Valley brickyard the basal York Till is one to four feet thick and rests on Ordovician strata. It is overlain by fossil-bearing stratified sediments of the Don Formation (Karrow, 1964, 1967). An appreciable hiatus occurred between the deposition of the York Till and the Don Formation as the basal parts of the latter contain pollen of a warm flora (Terasmae, 1960, p. 33).

The Don Formation consists of up to 25 feet of generally well stratified clay and sand displaying some crossbeds and cut-and-fill structures, and containing scattered remains of a wide variety of plants and animals. Noteworthy among the plants, in that they do not range so far north today, are southern white cedar, blue ash, osage orange, iron oak, chestnut oak, and black locust. In all, some 44 taxa of plants have been identified from macroscopic remains. Most of these have been recognized also as pollen, plus an additional 28 taxa, including pollen of sweet gum, that are not found in the Toronto area today (Terasmae, 1960). Terasmae also identified some 20 species of diatoms, and noted the presence of freshwater sponge spicules from the middle part of the Don Formation, which indicate lake, stream, and bog habitats. Animal remains recovered include the shells of some 40 species of pelecypods and gastropods (including a few land snails), part of a catfish, and bones of groundhog, deer, bison, bear, and giant beaver. Terasmae states: "An ecological-climatological interpretation of all evidence supplied by the fossils from the Don beds suggests that the annual mean temperature at the time of their deposition reached a maximum probably 5°F warmer than the present." In recent years investigations by Karrow and others have shown that wood is not common and leaves and vertebrate remains are very rare.

The Don Formation occurs some 60 feet above the level of Lake Ontario (el. 246 feet) in the brickyard but its upper surface is below lake level at Scarborough Bluffs. The lower part was deposited at the mouth of a river as it entered a lake in the Ontario basin. Lowering of the lake level is recorded by the character of the diatom assemblage in the middle part of the beds and by the sandy nature of the upper part of the formation. The upper sandy beds furthermore had been leached and weathered (Terasmae, 1960; Karrow, 1964) prior to the deposition of the overlying cool-climate Scarborough Formation. In the Don Valley brickyard the Don Formation is locally separated from the overlying Scarborough Formation by a layer of hard, compact, non-calcareous sand with some pebbles and cobbles. The formational contact was also encountered in borings at Scarborough Bluffs 15 feet below lake level. The Don Formation clearly represents part of an interglacial interval. It is generally assigned to the Sangamon but exact dispositon must await further work.

The Scarborough Formation (Karrow, 1964) is best known from the exposures on Scarborough Bluffs where it comprises a lower clayey-silt unit about 100 feet thick and an upper sandy unit about 50 feet thick, but it is also present in the Don Valley brickyard. There, the Scarborough clay unit is less than 25 feet thick and seemingly devoid of fossils although a few plant detrital seams have been noted between the thin clay layers. This deposit is very finely bedded and is regarded as a deep water deposit. The clay unit at Scarborough Bluffs includes peaty layers in places half an inch or more thick. These have yielded many small fossils including diatoms, and the leaves, seeds or spores of some 15 plants and, notably, the wing covers and other chitinous parts of some 72 species of beetle (Coleman, 1933). The beetles are reported as mostly extinct species but a re-study of the beetle content of the Scarborough plant detrital layers is needed. Terasmae has identified 41 plants from the clay unit, which indicates a boreal forest cover. The overlying sandy unit also contains plant detrital layers, sparse ostracods, and molluscs. Terasmae reports a pollen assemblage of boreal forest species similar to those found in the lower clay unit. He concludes that the climate during Scarborough times was perhaps 10°F cooler than now. The sands were deposited by southeast-flowing waters and presumably represent a delta formed in a lake (Lake Scarborough) that stood some 200 feet higher than the present lake. A piece of

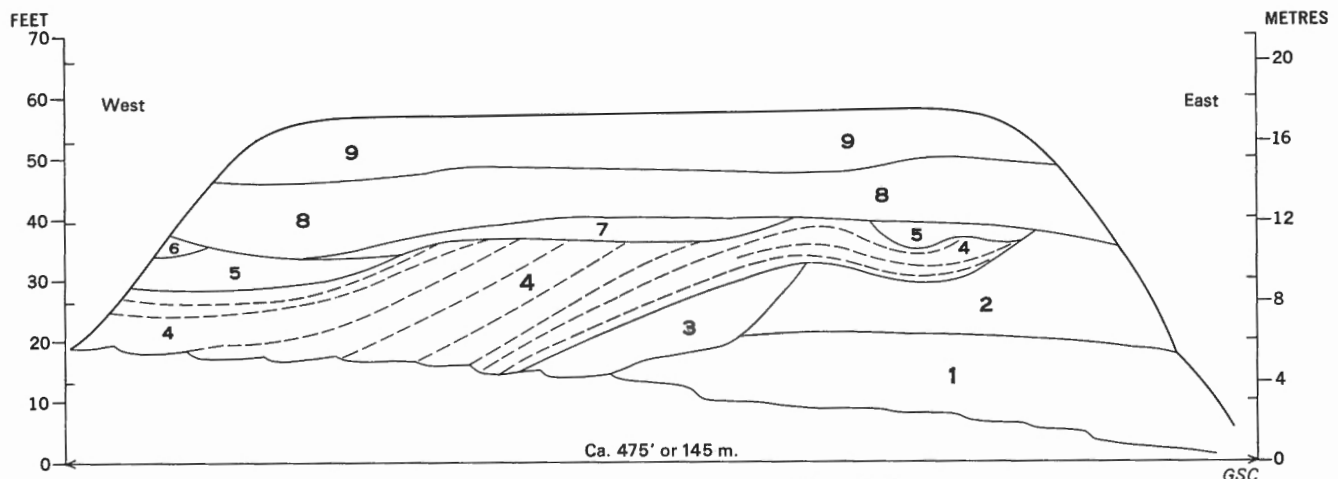
wood from near the top of the Scarborough Formation at Scarborough was dated at >52,000 years (Gro-2555), hence the precise age of the beds remains unknown.

The writer considers the Scarborough beds to be older than the St. Pierre sediments. Lake Scarborough formed after a lengthy erosional interval during which the Don beds were scoured and weathered. Lake Scarborough may be attributed to plugging of the drainage system by an advancing glacier, and the biotic record seems to support this view. The St. Pierre sediments represent an old river system with associated flood plain deposits, which required a through-flowing St. Lawrence Valley drainage system. Thus the Scarborough Formation has to be either younger or older than the St. Pierre sediments, and based on other evidence noted below it is considered to be older.

Organic matter is also present in post-Scarborough deposits. During and following the lowering of the Scarborough lake, deep valleys were cut into the Scarborough Formation, and in places into the Don Formation. Rising lake levels again brought about deposition of sandy sediments within these valleys. Boulders and cobbles, and occasional balls of till occur in the bottom of a channel in the Don Valley brickyard, which may indicate a significant time gap. The organic materials found in these sediments have been in part re-deposited from the older beds but some are indigenous to this stage of sedimentation. Of special interest is the first occurrence noted in Canada of the distinctive gastropod *Hendersonia occulta*. Coleman (1933) records water-worn "vertebrae of bison, part of a lower jaw of a bear, and a horn of an extinct deer, *Cervalces borealis*," and bits of ivory from mammoth or mastodon; these are from the Christie Street sand pits which P. F. Karrow correlates with the above-mentioned valley-fillings.

Subsequent to a mid-Wisconsin glacial invasion into the Lake Ontario basin and deposition of the Sunnybrook Till and varved clays, another period of non-glacial sedimentation occurred. The sediments formed at this time contain sparse plant remains and comprise the Thorncliffe Formation (Karrow, 1964), which consists of stratified clay, silt, and sand believed to have been deposited in both lakes and streams. Near Lake Ontario two wedges of till interfinger with the Thorncliffe Formation as a result of short-term expansions of the westward-moving ice. Both the sparse plant fragments and the till lenses (Seminary and Meadowcliffe Tills) indicate a rather cold climate. Jack pine and spruce were the dominant tree types with minor tamarack, oak, and birch; non-arboreal pollen was rather abundant. A small sample of plant material from the Thorncliffe beds has been dated at 38,900 ± 1,300 years (GSC-271). Glaciation, foreshadowed by the interfingering tills, then engulfed the region during the classical Wisconsin with deposition of the Leaside Till. Vegetal and animal life did not return until late glacial and early postglacial times.

Markham and Woodbridge. A few miles north of Toronto old organic materials have been observed beneath till. At Markham, a peat ball found in a gravel pit and dated at >34,000 years (W-194) is thought to have been derived from Thorncliffe beds or the Scarborough Formation. At Woodbridge, five discrete tills and a few feet of bedded silt, the whole comprising up to 35 feet, rest on some 25 feet of clayey silt with streaks and lenses of



ILLINOIAN
1. *YORK TILL: Clayey sand till*
2. *Sand, gravel*

EARLY WISCONSIN?
3. *Colluvial silt, minor organic matter*

EARLY WISCONSIN
4. *Stratified sediments: Clay, silt, sand, gravel, compressed peat, wood (C^{14} dated > 49,700 years)*
5. *SUNNYBROOK TILL: clayey till*

MID-WISCONSIN
6. *Gravel*

LATE (CLASSICAL) WISCONSIN
7. *WENTWORTH TILL: Gravelly till*
8. *HALTON-LEASIDE TILL: Silty till*
9. *Clayey till*

FIGURE XII-6. Wisconsin and Illinoian drift in railway-cut, Woodbridge, Ontario (after Karrow, 1965).

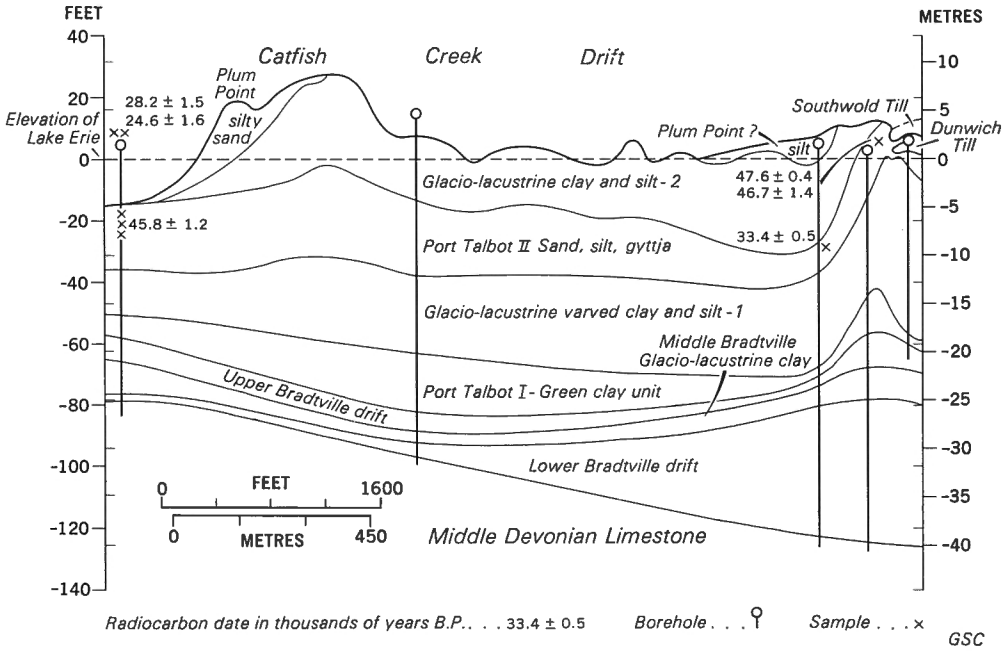


FIGURE XII-7
Generalized profile through Pleistocene deposits near Port Talbot, Ontario (after Dreimanis, et al., 1966).

peat and wood (Fig. XII-6). Another 14 feet of sand and gravel and 9 feet of till of Illinoian age are exposed at the base of the cut (Karrow, 1965). Some peat balls have been seen in the clayey till that overlies the peaty sediments. Palynological studies reveal that a northern boreal forest grew in the area at that time, with jack pine, black and white spruce, and birch the main tree types. This assemblage and a radiocarbon dating of >49,700 years (GSC-203) suggest a correlation with the Scarborough Formation, though a post-Scarborough age is also possible.

Niagara Falls. Silt in the buried St. David's channel near Niagara Falls has yielded a spore and pollen assemblage that is indicative of the northern boreal forest zone. Pollen studies by Terasmae substantiate the early report of fragments of spruce wood, from a boring over the channel, at a depth of 186 feet. Correlation of these silt deposits with other non-glacial beds in Ontario is not yet conclusive, but they may well be mid-Wisconsin.

Port Talbot. Buried organic deposits indicative of non-glacial conditions were reported by A. Dreimanis in 1951 from the Lake Erie shore south of London. As investigations progressed both along the shore and by drilling a complex sequence of deposits was indicated (Fig. XII-7). Radiocarbon datings on organic materials appeared to fall into main groupings, namely, about 48,000 to 44,000 years B.P. and about 28,000 to 23,000 years B.P. The indicated non-glacial intervals were thought to be separated by a glacial period represented by a till sheet (Southwold) found overlying the 'older' sediments in several places. The two intervals were named Port Talbot and Plum Point by Dreimanis (1957, 1958).

The main organic-bearing deposit at Port Talbot is exposed at the base of a 100-foot bluff overlooking Lake Erie. Contorted beds of silt, clay, calcareous gyttja, and scattered peat balls are overlain by glaciolacustrine clay and silt, organic-bearing silt and silty sand of Plum Point unit, and an uppermost drift complex of the main or Classical Wisconsin Glaciation. The gyttja contains larch, spruce, and water plants (*Potamogeton, Menyanthes,* and *Najas*) (Dreimanis, *et al.,* 1966). Part of a mastodon tusk was found in a clayey bouldery gravel some 600 feet east of the gyttja site at the same stratigraphic horizon. Some 17 species of ostracods have been recorded from the clayey silts both above and below the gyttja horizon. Mollusc shells from both the silts and the gyttja appear to have been crushed by the glacial action that contorted the beds but three genera have been recognized. The pollen assemblage from the gyttja horizon indicates that pine (mostly jack pine), spruce, larch, and birch were the dominant trees in the region. The base of the gyttja includes also large pollen grains of either white or red pine. The organic record indicates that the climate was cooler than that at present along the Lake Erie shore.

In boreholes at Port Talbot, beneath stratified sediments dating 47,700±1,200 years (GSC-217), up to about 30 feet of brownish buff varved silt and clay with a sparse pollen content overlies a few feet of greenish clay and silt with abundant pollen of jack pine, spruce, oak, and non-arboreal pollen. Mineralogical characteristics of the green clay and silt suggest a gap in the record after deposition of the basal, reddish (Bradtville) till, which rests on bedrock some 130 feet below surface of Lake Erie. Thus the Port Talbot non-glacial interval may extend back 50,000 years B.P. The lower Bradtville Till has been subdivided into three units of very similar lithol-

ogy with the lower two separated by a glaciolacustrine clay. The initial advance of Wisconsin ice into Erie basin, from the east, appears to have removed all evidence of interglacial or older glacial deposits in Port Talbot area. The initial advance was followed by a retreat stage during which a glacial lake occupied the basin. Two later encroachments of the ice formed the middle and upper Bradtville Till sheets. The time occupied by these early Wisconsin events is unknown but possibly 5,000 to 10,000 years.

Recent radiocarbon dates of 33,400±500 and 38,000±1,500 (GrN-4238, 4272) indicate that the time gap between the Port Talbot and Plum Point non-glacial intervals was short. It now appears that one long-continued interval, a mid-Wisconsin interstade, may be represented by the intertill stratified sediments and their contained organic matter. Southwold Till, formerly considered to intervene, is now correlated with the younger Catfish Creek Till which overlies the Plum Point sediments (Dreimanis, *et al.*, 1966). Thus, non-glacial conditions may have prevailed in the Port Talbot area from about 50,000 years to about 24,000 years ago. The ice front was, however, not too far away during this long period of sedimentation; the eastern end of Erie basin was blocked by ice on two occasions during which glacial lake clays were deposited and also ice of northern derivation was nearby when the Dunwich Till was deposited at some time during the Port Talbot interval.

Hudson Bay Lowland

The history of Pleistocene events in Hudson Bay Lowland has been dealt with by Lee (1968a). Important deposits bearing on the early Pleistocene record have recently been obtained by Craig and McDonald (1968).

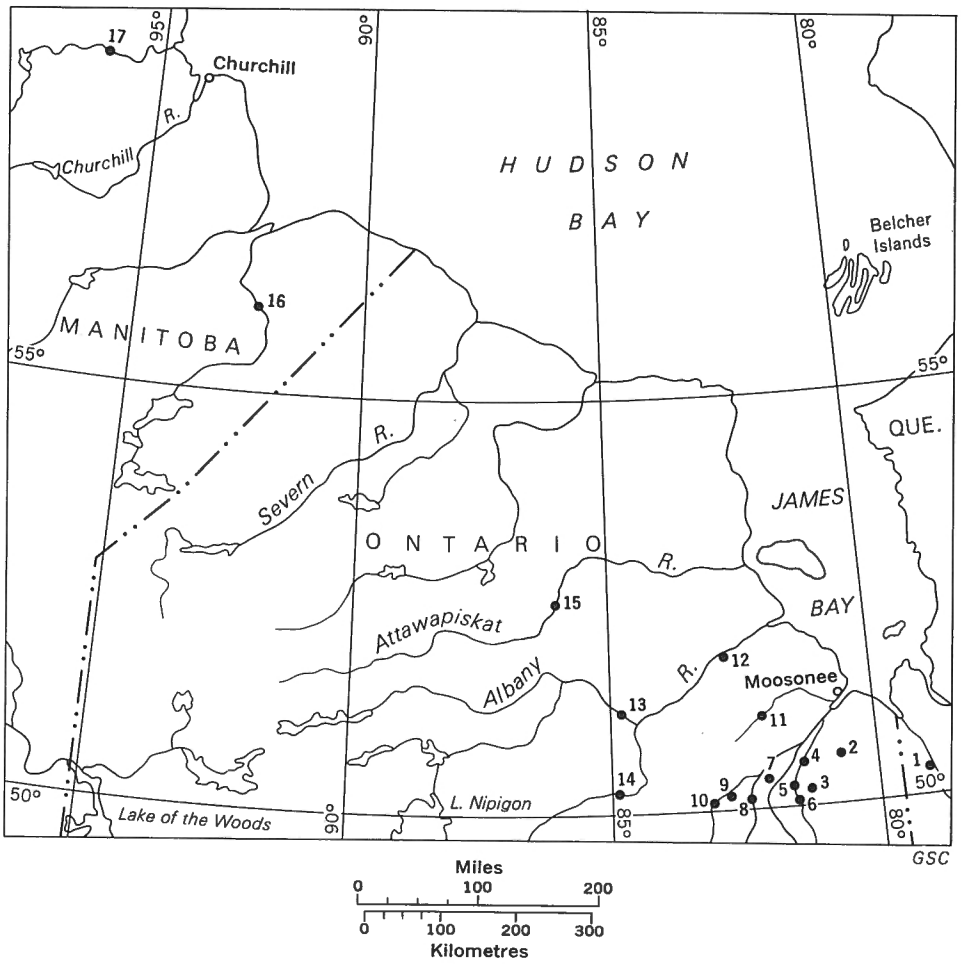


FIGURE XII-8
Location of buried organic deposits in Hudson Bay Lowland.

1. Harricanaw River
2. Nettogami River
3. Little Abitibi River
4. 5. 6. Abitibi River
7. Campbell Lake
8. 9. 10. Missinaibi and Opasatika River
11. Kwataboahegan River
12. 13. Albany River
14. Kenogami River
15. Attawapiskat River
16. Gods River
17. Seal River

Buried Organic Deposits (Fig. XII-8)

In the Hudson Bay Lowland there are numerous records of organic deposits lying below or between glacial tills. The published records are confusing in that the term lignite has been applied not only to Lower Cretaceous lignite deposits but also to Quaternary compressed peat and to sands, of both glacial and non-glacial affinities, containing detrital lignite. Eliminating the occurrences that are probably Cretaceous and the sands with detrital lignite grains of Quaternary age, a great many occurrences remain which, in the writer's opinion, record Quaternary non-glacial intervals in the Lowland. The occurrences are commonly cited as of interglacial age and indeed their wide distribution lends credence to this concept, but with present knowledge of Wisconsin interstades, a younger setting must be kept in mind where there is no strong evidence to the contrary. The more important sites exclusive of new ones found by Craig and McDonald are described below from east to west:

Harricanaw River. Two miles below Seven Mile Island, rhythmically bedded silt and sand with some thin vegetal layers underlie a marine clay beneath river sand and gravel. The vegetal layers were dated at >42,000 years (Y-1165).

Nettogami River. On the Nettogami River the lignite reported by Bell (1904) is most probably Pleistocene compressed peat. He refers (p. 161) to thick beds of black clay-shale with a great many very thin seams of a rather peaty lignite and to boulder clay containing striated pebbles both above and below the organic beds.

Little Abitibi River. Sections along the lower part of Little Abitibi River and along the adjacent Abitibi River reveal dark grey, compact and jointed silty clay beneath till and Tyrrell Sea deposits. Thin peaty laminae yielded pollen of black spruce, jack pine, and birch, together with varied non-arboreal pollen and spores of ferns and mosses. Farther up Little Abitibi River, silty and carbonaceous beds in the lower part of a 70-foot section of sand yielded much the same pollen assemblage. Wood from the lower part of the sand unit gave an age of >43,600 years (GSC-435). These deposits probably correlate with the Missinaibi beds on the Missinaibi River.

Abitibi River. The occurrence of buried Pleistocene organic deposits in the Moose River basin, including sites along Abitibi River (Bell, 1904; Wilson, 1906), may appear doubtful because of the Lower Cretaceous lignite found along Abitibi, Mattagami, and Missinaibi Rivers. Study of borehole data and excavations at Otter Rapids, however, revealed plant detritus in intertill sediments and indicated correlation with Pleistocene beds along Missinaibi River (Terasmae and Hughes, 1960). Furthermore at the Onakawana lignite workings, close to the river sections observed by Bell and Wilson, buried organic deposits of Pleistocene age were encountered in a shaft and in some 116 exploratory drillholes (Martison, 1953). Two till sheets separated by interglacial or interstadial sand, gravel, and clay are clearly present.

On Abitibi River 8 miles below Otter Rapids the writer has observed contorted beds of compact, dark grey, stony clay, containing broken or fractured marine shells, beneath younger drift. Eleven species of foraminifers, three of ostracods, and two of pelecypods were identified along with indeterminate species of forams, a pelecypod, and sponge spines, and spicules. Most of the species live in Hudson Bay at present. The stony marine clay overlies an oxidized quartzose sand, with a bed of differentially rotted igneous boulders. The marine deposits and the oxidized sand are considered to represent an interglacial interval. They are overlain by a Wisconsin clayey till and Tyrrell Sea deposits. Half a mile downstream the till rests on several feet of gravel containing limestone and igneous rocks; although the gravel appears to overlie the interglacial stony clay it is presumed to be much younger. Coleman (1941) also reported peaty clay and marine shells from interglacial sediments along Abitibi River, as well as shells from crumpled clay beds beneath till at Moose Factory.

Campbell Lake. A drillhole at Campbell Lake (Wawa Lakes) intersected 725 feet of drift resting on Paleozoic rocks (Hogg, *et al.*, 1953). J. Satterly reports that microfauna were recovered from samples taken at intervals between 262 and 645 feet—a section that includes two tills and two or three stratified clayey units. The fauna are of Pleistocene age and, as they were largely marine foraminifers, it is probable that the beds are interglacial. The great depth of overburden, in a section that includes both tills and marine beds, appears to indicate the presence of both preglacial and interglacial valleys of the Mattagami River. Microfauna identified in research laboratories, Shell Oil Co., are as follows:

Nonion grateloupi (d'Orbigny)
Elphidium gunteri (Cole)
Cibicides concentricus (Cushman)
Quinqueloculina lamarckiana (d'Orbigny)
Quinqueloculina seminulum (Linnaeus)
Discorbis sp.
Elphidium discoidale (d'Orbigny)
Siphonina cf. *pulchra* Cushman
echinoid spines
fragments gastropods and pelecypods
unidentified ostracod and plant spores
Cibicides pseudoungeriana (Cushman)
Globigerina bulloides d'Orbigny
Discorbis orbicularis (Terquem)

Missinaibi and Opasatika Rivers. Along Missinaibi River, above its junction with Opasatika River, lignite visible at low water was first noted by Bell (1879, p. 4C). Some of the lignite is undoubtedly the equivalent of the lignite of the Lower Cretaceous Mattagami Formation, but Bell's

description suggests that at least some of the occurrences are Pleistocene intertill compressed peat and wood. A section 9 miles above the mouth of Opasatika River is as follows:

Feet	
0 - 10	Hard, drab clay with striated pebbles and small boulders, and holding rather large valves of *Saxicava rugosa, Macoma calcarea,* and *Mya truncata.*
10 - 15	Hard, lead-coloured clay with yellow seams and spots, and red, grey, drab, and buff layers.
15 - 21	Lignite, made up of laminae of moss and sticks.
21 - 22	Clay with spots of lignite.
22 - 62	Unstratified drift full of small pebbles.
62 - 65	Yellowish stratified sand and gravel.

Three miles farther up the river Bell records:

0 - 45	Blue clay with pebbles, some striated.
45 - 47½	Lignite, made up principally of sticks and rushes.
47 - 127	Yellow weathering grey clay with pebbles, some striated.

Important observations were made also by J. M. Bell (1904) and J. Keele (1921), but the occurrence of both Cretaceous and Pleistocene deposits remained unproven until the area was restudied and pollen analyses were made by R. Auer (McLearn, 1927).

Terasmae and Hughes (1960) recognize five main Pleistocene units: (1) a lower drift; (2) a middle drift consisting of till and glaciofluvial sand and gravel; (3) layers of peat, organic silt, and clay, termed the Missinaibi beds; (4) an upper drift consisting mainly of till; and (5) marine clay, sand, and silt. Locally this sequence is eroded and overlain by the fluviatile deposits of the terraces along the rivers. Wood from Missinaibi beds gave a radiocarbon age of >53,000 years (Gro-1435). Palynological studies by Terasmae (1958) indicate a climate similar to the present or slightly cooler and possible correlation with the St. Pierre beds of St. Lawrence Lowlands. The inferred climate is unlike that represented by the Toronto Don beds, even allowing for the difference in latitude between the two areas, and hence the Missinaibi beds, if not representative of a late part of the Sangamon Interglaciation, must have been deposited during an early part of the Wisconsin glacial interval.

Kwataboahegan River. Wilson (1906) found solid peaty material in the bed of this river some 65 miles above its mouth:

> The mass where examined was six feet thick and it can be traced along the river for 430 feet. It is a dark brown colour and breaks off into lumps two to three feet thick. It burned slowly in the camp fire but left a large quantity of ash. Thin layers of the same material are exposed in the bank intercalated with the clay for several miles up the river.

A section, 60 miles above the mouth, is described by J. M. Bell (1904, p. 168) as follows:

> The seam, which has a maximum width of two feet six inches, outcrops almost continuously along the edge of the river for 450 feet in a bank 40 feet high. Though compact and hard it is never pure and is for the most part mixed with clay. Above it lies about 25 feet of hard, blue clay surmounted by six feet of shell-bearing post-glacial material. Below the seam is a hard stony clay containing many shells. This is of great scientific interest as it is the only point in the Moose Basin where interglacial shells are known to occur. The lignite itself is both arenaceous and argillaceous. It consists of thin layers of indurated moss with partings of clay and sand. It burned with considerable difficulty in the camp fire leaving a large residuum of clay and sand.

Though Martison (1953) did not examine the Kwataboahegan organic deposits, he considered them of Pleistocene age, and noted that high ash content is not characteristic of the Cretaceous lignites at Onakawana.

Albany River. Here till overlies a few feet of blue and brown clay (reported as leached) with two 2-inch beds of 'lignite.' Williams (1921) reported one bed to be "composed mostly of moss" and the other as "containing compressed roots." Nearby in a 90-foot section on the north bank of the river the till cover is 50 feet thick and rests on similar clay though no 'lignite' was observed. The Pleistocene age of these materials was corroborated by Terasmae and Hughes (1960). They conclude that the spore and pollen assemblages are entirely unlike those of the lignite of the Cretaceous Mattagami Formation, known on Mattagami, Abitibi, Missinaibi, and Opasatika Rivers.

Farther down Albany River 10 feet of stratified sand rests on 20 feet of pebble clay, presumably till, that in turn rests on 20 feet of thin-bedded clay, peat and moss, and 2 feet varved clay and pebble clay beneath which are 8 feet of pebble clay possibly a till. The bedded clay, peat, and moss sequence was considered by Martison to be interglacial.

Kenogami River, a tributary of Albany River, intersects a preglacial river channel cut into Silurian limestone. R. Bell (1887, p. 38) reported a basal till overlain by a 6- to 8-foot bed of soft lignite, containing many flattened stems of small trees and succeeded by 30 to 40 feet of rudely stratified red and grey drift with rounded boulders and many pebbles. This is most probably interglacial.

Attawapiskat River. Well-stratified clay with silt and sand laminae and sparse plant fragments occurs beneath 6 to 10 feet of sand and gravel and 12 to 15 feet of clayey till. The plant fragments are dated at >35,800 years (GSC-83). It is likely that the organic-bearing sediments are related to the Missinaibi beds.

Gods River. On Gods River, formerly called the Shamattawa, Tyrrell (1913) reports that intertill sand and gravel and the basal part of the overlying till contain moss and wood, partly altered to lignite. He recognized two tills over a wide area, in places separated by a striated boulder pavement. He noted the intertill sand and gravel also along Hayes River but could find no organic remains.

Seal River. A buried non-glacial deposit carrying organic materials was reported by Taylor (1961). This occurrence is on the Precambrian Shield, 85 miles west of the Paleo-



PLATE XII-3. Deep meltwater channel between unglaciated Cypress Hills and Wisconsin end moraine complex, southwestern Saskatchewan. Vertical airphoto of meltwater channel, about 700 feet deep, now occupied by Adams Creek. Road leads north to Maple Creek, Saskatchewan. Scale 1 inch to 3,000 feet.

zoic rocks, very close to the limit reached by the Tyrrell Sea. Beneath sandy till typical of Shield areas, is a fluviatile bouldery gravel with a matrix of goethite. Two 6-inch layers in the gravel contain casts and impressions of leaves and twigs partly replaced by, and set in a matrix of, goethite. The plant-bearing layers represent the replacement of a woody peat composed of moss, sedge, grass, and shrubs including herbaceous plants such as leather-leaf, dwarf lamel, and bear-berry. The plant assemblage is similar to that of the area today, and is believed to have been deposited during an interglacial interval.

Interior Plains

Early Pleistocene Events

Pediment surfaces. In late Tertiary the climate of the Interior Plains is inferred to have been arid to semiarid, and a series of pediplains was developed along the mountain front and around smaller uplands to the east (Gravenor and Bayrock, 1961; Parizek, 1964; Barton, *et al.,* 1965). With the change to a cooler and moister climate at the close of the Tertiary and in earliest Quaternary extensive valley erosion, locally accompanied by alluviation, left the present uplands as remnants of the pediplain system. Some of these uplands that escaped glacial modification are Cypress Hills (Pl. XII-3) and Wood Mountain near the International Boundary. Other uplands such as Missouri Coteau in southern Saskatchewan and Hand Hills in east-central Alberta have been extensively modified. There is some disagreement as to the degree of the effects of glaciation on the gross topography of the Plains. Contour maps of the bedrock surface give a general picture of the preglacial topography but as much as 1,000 feet of drift is known in some valleys in Saskatchewan, indicating that considerable modification of the topography has taken place.

Drainage systems. At the end of the Tertiary period a mature, dendritic drainage system existed on the Interior Plains (Fig. XII-9). A good account of early work on buried valleys dating back to that of Dawson in 1885 is given by Stalker (1961). He reports also that the glaciers responsible for the numerous till sheets recognized on the southern plains had the over-all effect of forcing rivers to follow more southerly courses. In a general way the major elements of the present-day drainage reflect the trend of the old valleys, but there are many divergences as a result of disruption and modification of the preglacial rivers. For instance, near the town of Peace River Hen-

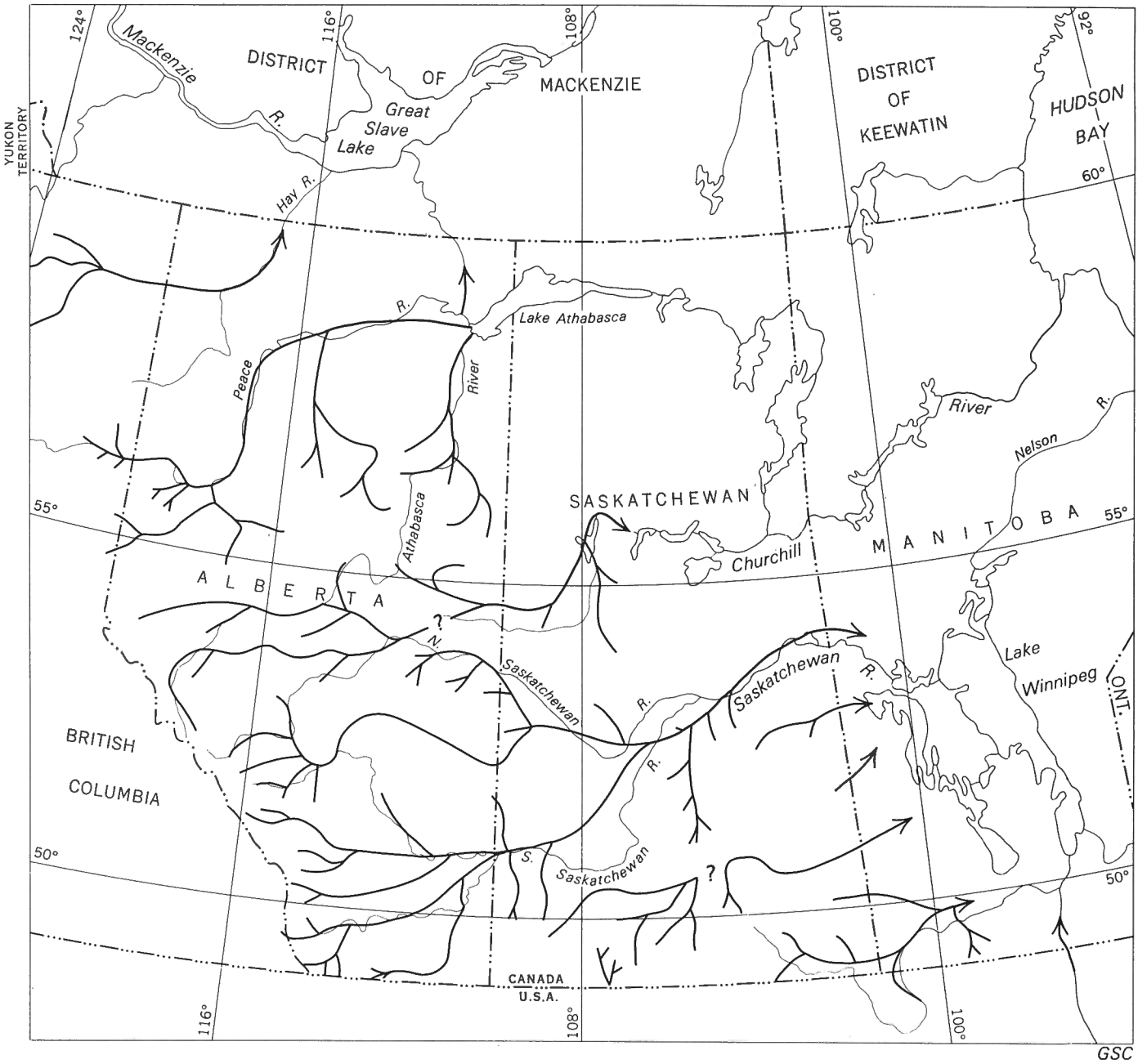


FIGURE XII-9. Inferred preglacial drainage systems of the Interior Plains (modified from available sources).

derson (1959a) records 800 feet of drift in a preglacial channel 3 miles southeast of the present river, and also shows that the preglacial Smoky River joined Peace River some 20 miles southwest of the present junction. In central Alberta the preglacial Red Deer River followed a markedly different course from that of the present, north of Red Deer; this preglacial channel is now occupied in places by Battle River. Similarly the preglacial South Saskatchewan River system in both Alberta and Saskatchewan was quite different from that of the present (Stalker, 1961; Christiansen, 1967). The preglacial Milk River in southern Alberta was probably a tributary of the South Saskatchewan rather than Missouri River, and farther east this latter river may have flowed across southeastern Saskatchewan into Manitoba and thence presumably northward to Hudson Bay rather than joining the Mississippi as at present (Meneley, 1957). The course of the preglacial Qu'Appelle River in south-central Saskatchewan is in doubt (Kupsch, 1964).

On the basis of shape and contained sediments Stalker has identified both preglacial and interglacial valleys in southern Alberta. He has noted a very dark grey till in

the lower parts of all preglacial valleys. In the bottom of their channels the preglacial valleys are characterized by the presence of the Saskatchewan gravels, free of stones derived from the Canadian Shield. Gravel is uncommon in the interglacial valleys, and where present contains some Canadian Shield stones, although they may be very rare. The preglacial valleys in southern Saskatchewan are generally 4 to 10 miles wide and have gently sloping sides, whereas interglacial valleys are a mile to 2 miles wide with steeply sloping walls (Christiansen, 1967). Christiansen records that the preglacial valleys are filled with 50 to 1,000 feet of drift. They may be apparent at the surface where the drift is thin, but are completely obscured where it is thick.

Saskatchewan gravels. The occurrence of preglacial Saskatchewan gravels on the Interior Plains is of great interest and their age has long been controversial (Westgate, 1965). These buried gravels are recognized in Alberta, Saskatchewan, and Manitoba, occurring as alluvial terraces and benches below the Tertiary (Miocene–Pliocene) pediment surfaces and as lower-level channels or valley fills. They are very extensive beneath the drift mantle whether this be a few feet or more than 1,000 feet thick; generally they are a few feet to a few tens of feet thick, varying laterally from sand to sand mixed with coarse gravel. Henderson (1959a), however, reports more than 100 feet of gravel in Peace River area. Westgate found that 98 per cent of the gravel-sized material in southeastern Alberta is made up of quartzite, argillite, and chert, the remainder being arkose, limestone, a green porphyry, and local bedrock. Their source is the Cordilleran Region to the west, also earlier deposited Cordilleran gravels that capped the pediment surfaces of the Interior Plains. Over most of the Plains they rest unconformably on Cretaceous rocks. Frost-action structures have been observed beneath Saskatchewan gravels by J. A. Westgate in southern Alberta, and by Westgate and Bayrock (1964) in central Alberta. In places the gravels are undisturbed by the frost structures. A periglacial environment prevailed during deposition of at least the early part of the unit. Westgate notes the occurrence of bones of a woolly mammoth *Mammuthus primigenius*, and of a horse (*Equus* sp.) somewhat smaller than the present-day horse. He considers that the bulk of the Saskatchewan gravels probably range in age from early to late Pleistocene. Stalker, however, considers them to be early Pleistocene.

Proglacial deposits. Fine-grained sediments including varved sediments have been noted resting on Saskatchewan gravels in major preglacial valleys and directly on bedrock in their tributaries. Thicknesses up to 150 feet have been recorded. They have formerly been included with the Saskatchewan gravels as there is no evidence of a prolonged break in sedimentation prior to their deposition. Westgate (1965), however, treats them as a separate stratigraphic unit—the Wolf Island sediments. They were deposited in cold, quiet, proglacial lakes rather than in swift-flowing

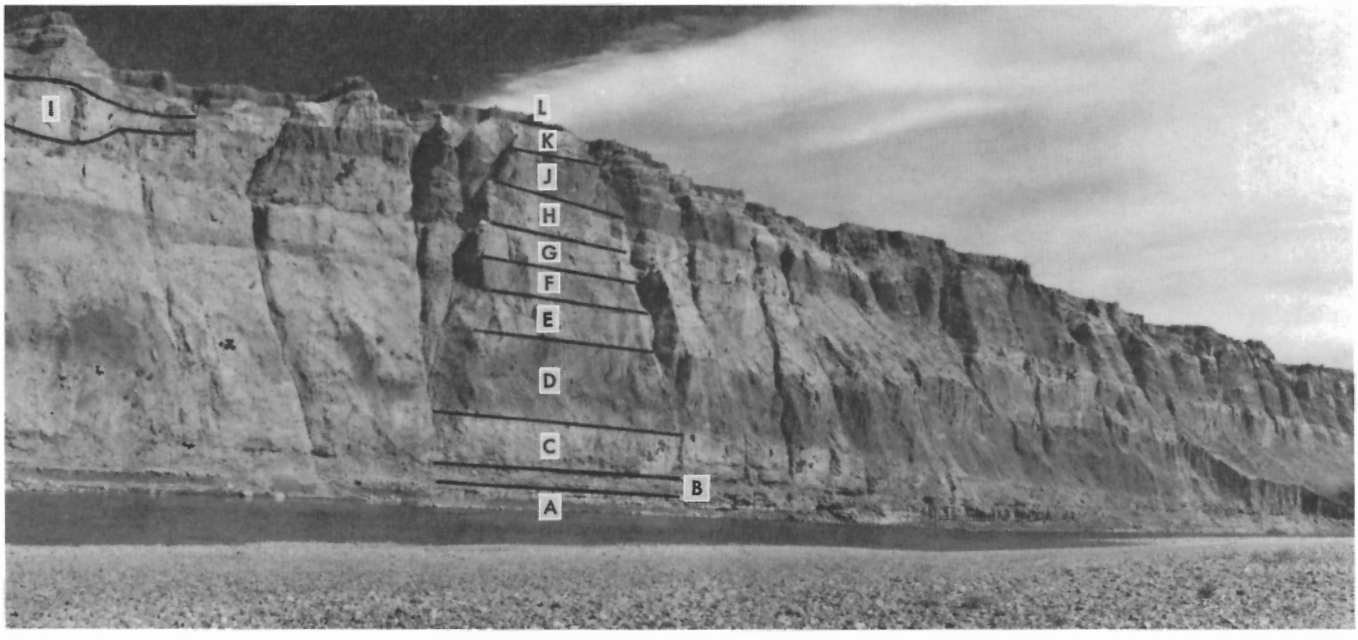


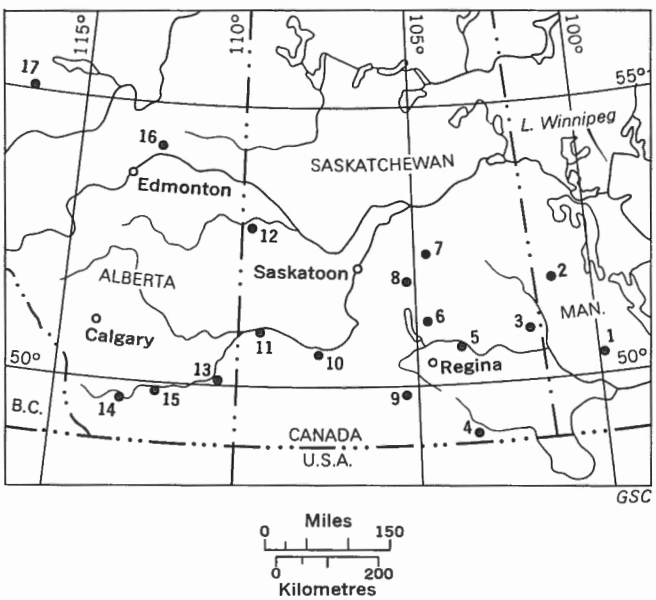
PLATE XII-4. Section of non-glacial and glacial sediments on Oldman River at Brocket, Alberta.

L. Sandy loam; modern soil.
K. Sand; lake and outwash.
J. Varved clay, silt and fine sand; glacial lake.
I. Till (Buffalo Lake); Laurentide.
H. Sand, silt, clay; lacustrine or alluvial; possibly mid-Wisconsin.
G. Till (unnamed), medium light brown; Laurentide?
F. Till (Brocket), dark brown; Laurentide.
E. Till (Maunsell), light bluish grey; Laurentide.
D. Till (Labuma), dark brown to black; Laurentide.
C. Till (Albertan), light grey; Cordilleran.
B. Gravel (no Canadian Shield stones); mostly outwash.
A. Bedrock (Willow Creek Formation); Paleocene.

streams, and appear to reflect a blockage of the northeast-flowing rivers by encroachment of Laurentide ice.

Early glaciations. Very little is known as to the course or extent of pre-Classical Wisconsin glaciers. The generally flat character of the Plains limits observations of Pleistocene stratigraphy to the relatively few sections along entrenched rivers. In Saskatchewan great use has been made of drilling rigs and side-hole samplers to establish the stratigraphy; this practice gives great promise of the establishment of a firm chronology in a region that might otherwise remain an enigma. In many places multiple till sheets, with and without intervening stratified sediments, have been recognized (Pl. XII-4). There is as yet little agreement as to the precise ages of the tills or sediments. Opinions vary from those who believe that all stages of the Pleistocene are represented to those believing in only a Wisconsin glaciation. Some writers place the limit of the Classical Wisconsin drift south of the 49th parallel while others contend that it lies well north of Cypress Hills. Only recently have radiocarbon datings proved the existence of Pleistocene age organic-bearing sediments that predate the Classical Wisconsin. Extensive oxidized zones are recognized in some thick till sections; these appear to represent periods of near-surface exposure during both interglacial and intraglacial times.

In the vicinity of Del Bonita Hills and Cypress Hills an old drift topography occurs at higher altitudes than the surrounding more youthful hummocky terrain. Westgate (1965) terms the higher and older drift the Elkwater drift and considers it correlative with the most extensive and southernmost drift sheet in Montana. He recognizes five Laurentide drift sheets in the Foremost–Cypress Hills area of southeastern Alberta and attributes end moraines to each. The ice sheets generally advanced southeast but with some variations due to lobing in the marginal zones. He considers the oldest drift sheet to be post-Sangamon. Stalker (1963), however, believes that most, if not all, the pre-Wisconsin glacial periods are represented by tills in southern Alberta.



1. Riding Mountain
2. Duck Mountain
3. Churchbridge
4. Outram
5. Fort Qu'Appelle
6. Gregherd
7. Spalding
8. Lanigan
9. Spring Valley
10. Swift Current
11. Prelate Ferry
12. Marsden
13. Medicine Hat
14. Lethbridge area
15. Taber
16. Smoky Lake
17. Goose River

FIGURE XII-10. Location of buried organic deposits in Interior Plains.

Buried Organic Deposits (Fig. XII-10)

Organic materials in stratified deposits beneath one or more till sheets are known from many places on the Interior Plains.

Riding Mountain. Buried organic materials were found in a highway-cut on the north side of Minnedosa River valley (Klassen, *et al.*, 1967). Three till units are separated by silt and sand layers, the whole comprising 61 feet (Pl. XII-5). This complex overlies a 5-foot silt layer that contains plant remains and rodent bones. The silt unit rests on a foot of limonite-stained gravel with small limestone pebbles weathered to a powder. Beneath is 16 feet of dark grey, shale-rich till, the surface of which also appears to be weathered in places. The sediments are almost devoid of pollen, presumably due to oxidation. Bone fragments are of arctic ground squirrel *(Citellus undulatus)* and a large vole (*Microtus* sp.). Plant fragments gave an age of >31,300 years (GSC-297). Several boreholes on the south side of Riding Mountain encountered stratified sediments beneath two tills and overlying a third. Wood chips from a hole near Inglis, farther northwest, were encountered in clay at a depth of 196 to 212 feet and were dated at >30,000 years (GSC-218). Bedded silts 75 feet thick occur beneath the clay unit and rest directly on shale bedrock. It may be that the Riding Mountain buried deposits are interglacial.

Duck Mountain. In 1964 R. W. Klassen re-examined the locale of interglacial sediments reported by Tyrrell (1892, p. 116). The section on Rolling (Roaring) River was heavily slumped but the Heart Hill section was clear. Two or more tills totalling 54 feet overlie 8.5 feet of silty beds containing small pelecypods, gastropods, ostracods, plant fragments, and seeds. The silt beds overlie 10 feet sand and gravel, and 15 feet sand; the lower 58 feet down to river level was heavily slumped. The upper 1.5 feet of organic-bearing silt is brown rather than grey, suggestive of a weathered zone. The contact with the overlying till is gradational over some 6 inches. The basal inch or two of the silt unit is a clayey sand, cemented with limonite. It contains pelecypods and much carbonaceous material including plant fragments. Some of the latter gave a radiocarbon dating of >38,000 years (GSC-284).

The silt unit was sampled at 2½-inch intervals and examined for both shells and pollen. Seven species of

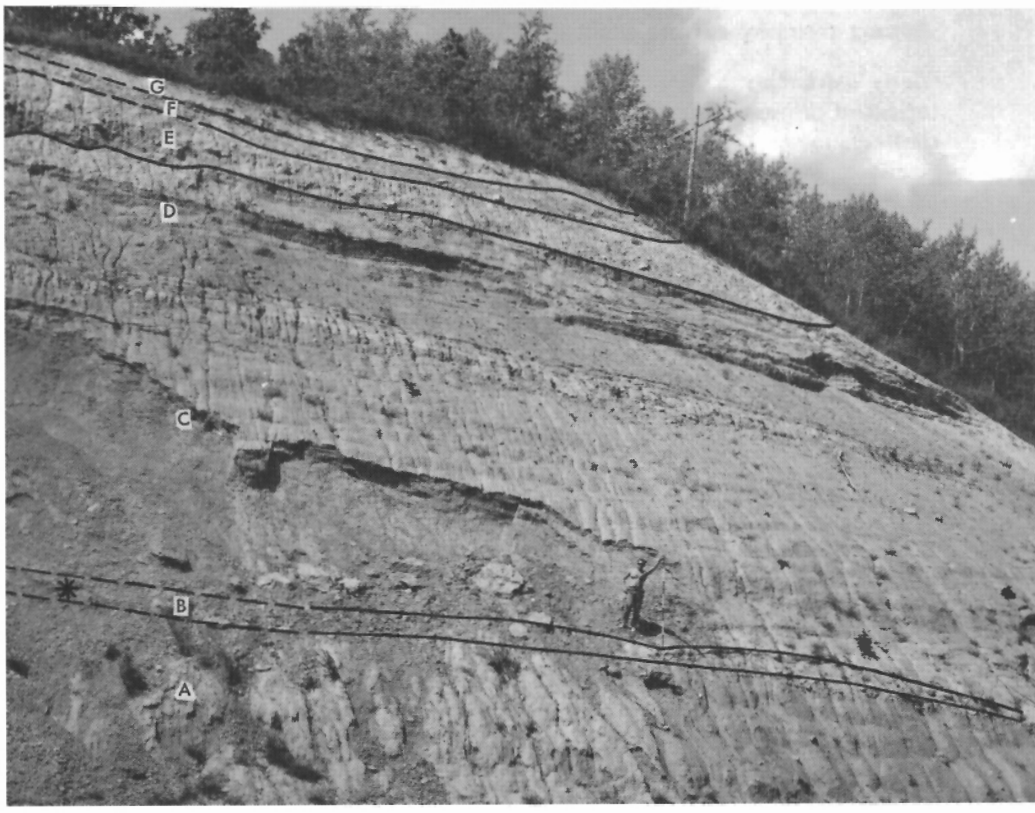


PLATE XII-5
Section of glacial and interglacial deposits on Minnedosa River, north of Minnedosa, Manitoba. A, till; B, silt with bones (collection site); C, till; D, silt, sand, gravel; E, till; F, sand and gravel; G, till.

pelecypods, three gastropods, and ten ostracods have been identified. The ostracods and the pollen indicate a moist climate about as warm as the present, which supports the interglacial interpretation of Tyrrell. The fossil assemblage also indicates a cool–warm–cool sequence; this sequence was terminated by glaciation and deposition of till.

Churchbridge. Tyrrell (1892, p. 142E) recorded the log of a well at Churchbridge that appears to indicate some 32 feet of drift, mainly till, overlying more than 235 feet of clayey and sandy sediments. A piece of wood was found at a depth of 200 feet and identified as a species of larch, *Larix churchbridgensis.* Elsewhere in the area as much as 165 feet of till is reported overlying sand or gravel.

Outram. Christiansen and Parizek (1961) report wood chips from probable lacustrine sediments at a depth of 170 feet from a test-hole in southeastern Saskatchewan. The section from the top down is, light, olive-brown calcareous till, 10 feet; grey, calcareous till with two clayey beds, 92 feet; stratified, oxidized and leached sand rich in organic matter, 8 feet; grey, calcareous till, 20 feet; grey, interbedded sand, silt and clay (from which wood chips were obtained at depth 170 feet), 71 feet; grey, calcareous till, 15 feet; stratified sediments, 144 feet; light brownish grey, calcareous till, 8 feet; gravel, 3 feet; pale yellow, kaolinitic, calcareous till, 4 feet; sand, 2 feet; and light brownish grey, calcareous till, 1 foot +. The wood chips were dated at 27,750 ± 1,200 years (S-96), which suggests correlation with the Prelate Ferry interval.

Fort Qu'Appelle. Fossiliferous sediments beneath 200 feet of drift were reported by Christiansen (1960). He reports that upstream from Fort Qu'Appelle a till overlies 10 to 25 feet sand and 20 to 40 feet of exposed gravel; down the valley towards Lebret, terraced valley-fill is at least 120 feet thick. Vertebrate fossils in the sediments, including bones and teeth of bison, mammoth, horse, wolf, and bear, are tentatively assigned to the Sangamon by L. S. Russell. Christiansen thinks that there is an older drift beneath these sediments.

Gregherd. Wood chips from sand beneath till, in a well at a depth of 300 feet, were dated at >30,000 years (S-111). The overlying till is considered late Wisconsin but the age of the non-glacial sediments is unknown.

Spalding. A boring has yielded wood chips from the lowest of three tills at a depth of 221 feet. The wood was dated at >34,000 years (S-127). It probably represents an interglacial deposit.

Lanigan. Wood from a depth of 540 feet in a drift-filled valley and only 20 feet above bedrock was encountered during sinking of a mine shaft. It was dated at >42,000 years (GSC-632), and is believed to be of Pleistocene age.

Spring Valley. Layers and lenses of peat associated with silty-clay beds occur beneath a till mantle in the hummocky ice-thrust moraine of the Dirt Hills. The peat was dated >38,000 years (GSC-790) and pollen indicates Pleistocene age.

Swift Current Creek. Wickenden (1931) reported a multiple till section, with intervening sediments and organic matter, in Swift Current Creek valley near its junction with South Saskatchewan River. From the surface downward are thin lacustrine clays; 20 to 70 feet of till; 35 feet white sand and gravel with some poorly preserved plant material; 3 to 45 feet dark grey till; and up to 225 feet sand and gravel with some poorly preserved peat and plant detritus. The last sediments rest either directly on bedrock or on a brown to yellowish till that locally displays a weathered zone, 8 feet thick, with poorly preserved plant materials and rootlets. On the basis of the great thickness of the lower gravels, the presence of plant materials, and the weathered zone on the basal till, Wickenden regarded the lower gravels as deposited during a long, moist interglacial interval. Christiansen (1959) has linked the two upper tills to end moraines in the area and assigns the upper unit of sand and gravel to a proglacial, very late part of the Wisconsin. He regards the lower stratified sediments as earlier proglacial Wisconsin deposits.

Prelate Ferry. David (1966) reported a buried soil beneath some 120 feet of drift including three tills from South Saskatchewan River valley north of Prelate. From the surface downward are lacustrine silt and clay, 16 feet; oxidized, calcareous till, 13 feet; stratified sand, 26 feet; mainly oxidized very calcareous till, 39 feet; stratified sand, silt, and a thin basal marl-like layer, 24 to 25 feet; paleosol, 3 to 4 feet; oxidized, very calcareous, sandy clay-loam till, 7 feet; stratified and crossbedded silt, sand, and gravel, 8 feet; and a basal, oxidized, calcareous till, 72 feet. The marl-like layer, only 6 to 8 inches thick, provides a widespread and useful horizon marker for locating the paleosol. The buried soil has been dated at 20,000 ± 850 years (S-176). David considers the paleosol to have developed during a non-glacial interval that he named the Prelate Ferry interval. Its beginning is unknown but it ended about 20,000 years ago when advancing Classical Wisconsin glaciers led to burial of the soil, first by stratified sediments and later by till.

Marsden. Christiansen (1965) reports a soil, buried beneath a few feet of till, about 125 miles north of the Prelate Ferry paleosol exposures. The dating of 21,000 ± 800 years (S-228) suggests that this soil was formed also during the Prelate Ferry interval.

Medicine Hat. Four miles north of Medicine Hat, in South Saskatchewan River valley, Stalker records about 100 feet of intertill sediments exposed in a small buried valley cut in till. Wood fragments and dark carbonaceous layers occur, dated at 24,490 ± 200 and 28,630 ± 800 years (GSC-205, 578). Two tills of differing lithologies overlie the stratified sediments and are considered to represent the Classical Wisconsin. Beneath the sediments in the buried valley are other tills and intervening sediments with abundant wood, bones, and shells which Stalker considers to be Sangamon or older. Wood from near river level at a site 3 miles north of town and beneath 220 feet of drift including several tills was dated >46,700 years (GSC-543). He considers the containing sediments to be of Yarmouth age.

Lethbridge. Multiple till sections have been described by Stalker (1963) on Oldman River. At one site 10 miles west of Lethbridge wood and cones of black spruce were found in the basal part of intertill sand, silt, and gravel. The wood has been dated as >54,500 years B.P. (GSC-237). Two tills occur beneath these sediments at several places along the river and at one place the basal till shows evidence of weathering. Stalker considers these stratified sediments to be present beneath an upper till as far north as Edmonton and to be of Sangamon age. The two lower tills are equally extensive.

At the Kipp section on Oldman River about 7 miles west of Lethbridge, the humic part of a wood sample from the intertill sediments is dated at >37,000 years (L-455A). Some 24 feet of intertill sediments includes a basal, consolidated, white sand that forms a prominent horizon marker in the area. The sediments are overlain by 32 feet of dark, compact till of unknown age and this in turn by some 163 feet of alluvial sediments within which wood fragments have been found. Another 45 feet of sediments overlying the alluvium includes varved clays and minor till lenses and appears to be glacial deposits. These sediments are overlain by 50 feet of sediments carrying some snail shells that may have been deposited relatively close to the ice front. Similar intertill sediments occur near Taber 30 miles east of Lethbridge where wood fragments have been dated at >32,000 years (S-65).

Smoky Lake. A log of spruce was found in till at a depth of 24 feet near Smoky Lake some 60 miles northeast of Edmonton. This wood was dated at >31,000 years (S-92).

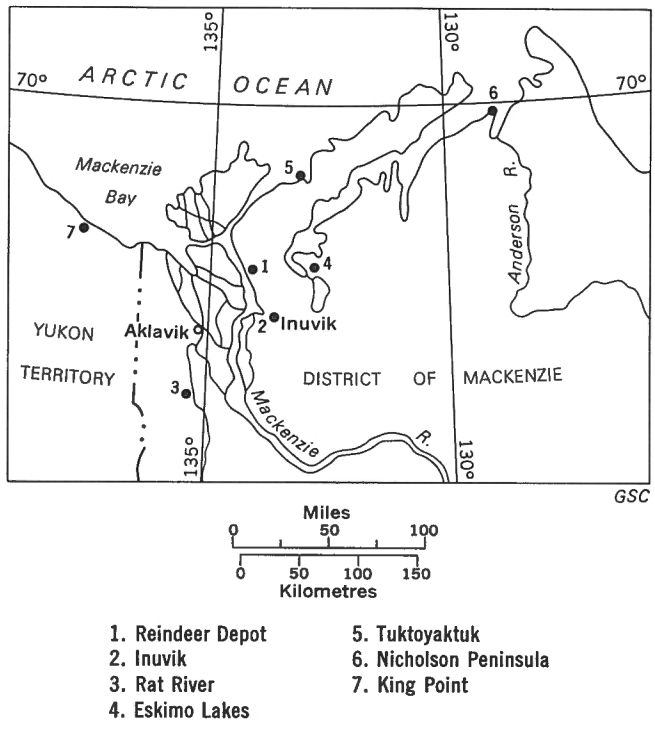
Goose River. The northernmost occurrence of buried wood in the Interior Plains is from Goose River northwest of Edmonton. There alluvium and till overlie a crossbedded sand unit in which wood was found that dated >42,500 years (GSC-501).

Mainland Arctic Coastal Plain

A 10-to-20-mile-wide belt of older glaciated terrain on the eastern side of the Cordillera west of Peel River is continuous northward onto the Yukon Coastal Plain. This belt extends westward from Mackenzie Delta to the

Alaskan boundary (Fyles, 1966) and has subdued hummocky topography as far west as Herschel Island, but beyond this area, where the zone narrows to only a few miles, glacial landforms are extremely rare. East from Kay Point, 20 miles southeast of Herschel Island, the older glaciated terrain is bordered on the northeast by a narrow zone of Wisconsin drift. Beneath these deposits interglacial sediments have been seen on headlands between Kay Point and Mackenzie Delta. The sediments grade, in general, from silt in the west to gravel in the east; Fyles considers them to represent a single stratigraphic unit.

In the Mackenzie Delta region, Mackay (1963) found evidence for glaciation of the main and near-shore islands, and possibly also the off-shore islands. Tuktoyaktuk Peninsula was partly glaciated; evidence of glacial action is lacking for the northeastern part. Farther east on Cape Bathurst and Baillie Island, an extensive low-lying plain is underlain by silt and sand beneath which is a marine clay with rare wood (Fyles, 1966). No clear distinction between postglacial and interglacial sediments could be made. This area and part of Tuktoyaktuk Peninsula are shown as outwash (including alluvium) on the Glacial Map of Canada (1253A, *in folio*) pending clarification of the age of the deposits. Possibly older glacial as well as interglacial deposits are present along the northern edge of the Mackenzie Delta complex.



1. Reindeer Depot
2. Inuvik
3. Rat River
4. Eskimo Lakes
5. Tuktoyaktuk
6. Nicholson Peninsula
7. King Point

FIGURE XII-11. Location of buried organic deposits of the mainland Arctic Coastal Plain.

Buried Organic Deposits (Fig. XII-11)

Near Reindeer Depot on Mackenzie River 10 feet of peat and silt underlies some 200 feet of drift (Porsild, 1938). Cones of larch were found although the site is 50 miles beyond its present northern limit. Pollen includes abundant birch and alder, much white and black spruce, jack pine, tamarack, and willow, and some ericaceous shrubs, small herbaceous plants, grasses, sedges, and ferns. J. Terasmae considers the pollen assemblage to represent an interglacial deposit, probably the Sangamon. The peat gave an age of >42,000 years (L-522A).

Wood from a delta-kame complex at Inuvik gave an age of >39,000 (GSC-29) and was evidently derived from older non-glacial deposits. Some 60 miles southwest of Inuvik on Rat River, a beaver-chewed stick near the base of a 40-foot section of silt with organic layers and overlying an older till was dated >38,600 years (GSC-120). Evidence elsewhere along the river suggests that Laurentide ice, possibly in Classical Wisconsin time, overrode the site and extended several miles to the west.

Buried organic materials, dated at >50,900 years (GSC-329), occur southwest of Eskimo Lakes where 7 feet of modern peat overlies 76 feet of gravel that rests on at least 46 feet of organic silt with two thin peaty layers in its upper part. O. L. Hughes regards the gravel as outwash in front of a moraine left by the last major ice sheet. According to Terasmae, the pollen content of the peat is unlike that of the present and probably represents an interglacial interval. Fyles (1966) reports a great variety of materials in the southwest part of Eskimo Lakes area where organic matter was noted in three different lithologies.

A non-glacial interval is indicated by the occurrence of abraded wood from a pingo at Tuktoyaktuk dated at >33,000 years (L-300A). Also flattened wood from 20-foot depth in contorted sands, believed to be ice shoved, at the north end of Nicholson Peninsula gave a date of >35,200 years (GSC-34). O'Neill (1924) observed thin peat beds in silty sediments along the arctic coast east of the mouth of Mackenzie River, and he mentions an earlier report by Sir John Franklin of "poor lignite" in similar sediments on nearby Pullen Island.

The interglacial deposits along the arctic coast west of Mackenzie Delta contain a great variety of organic materials. Fyles (1966) reports wood up to a foot in diameter from gravels near the delta, and smaller pieces of wood, peat, freshwater shells, rare bones, and tusks from the silts farther west. Thin beds of marine clay and a few ice-wedge casts complicate the succession. An important site along this coast is at King Point 4 miles east of a moraine believed by O. L. Hughes to mark the western limit of the Classical Wisconsin ice sheet. There up to 3 feet of peat overlies 8 to 15 feet of silt and sand that rests on 20 to 30 feet of Wisconsin till. Beneath this till a stony clay, containing marine shells, grades downward into organic-bearing silts. Plant materials collected from these silts, 2 feet above the base of the seacliff, gave a radiocarbon dating of >51,100 years (GSC-151-2).

The change from plant-bearing silt to shell-bearing stony clay probably reflects a period of marine inundation as a result of the encroaching Wisconsin ice sheet.

Canadian Shield

Mainland

The mainland Shield region has been subjected to intense scouring action by a number of glaciations, but the gross physiography has probably changed little since the end of the Tertiary period. Stream valleys along rugged parts of the Quebec and Labrador coasts, however, have been markedly modified by valley glaciers stemming, in the main, from the interior ice sheet. During interglacial intervals stream erosion probably contributed to valley deepening along these same coasts but elsewhere was relatively ineffective.

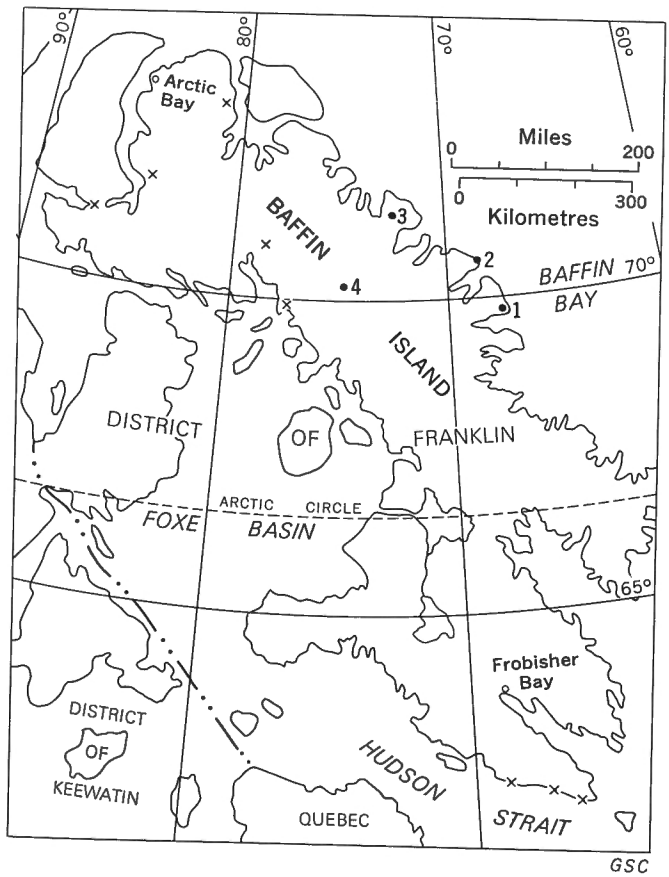
In the Torngat and Kaumajet-Kiglapait Mountains of northern Labrador, an early Torngat Glaciation is believed responsible for widely scattered erratics on the highest peaks, and the later Koroksoak Glaciation for high-level lateral moraines, kame terraces, and a felsenmere trimline well below the mountain tops (Wheeler, 1958; Ives, 1958a, b, 1960a; Tomlinson, 1959; Løken, 1962). A still lower trimline, referred to the Saglek Glaciation is regarded by Andrews (1963) as the probable upper limit of the Classical Wisconsin ice sheet. At that time large parts of the mountains projected above the ice.

The last major glaciation was generally effective in removing the deposits of older glacial and interglacial intervals; buried organic deposits are rare in the Canadian Shield. They are, however, more likely to be found in peripheral areas than in the interior, as for example the Seal River occurrence described under Hudson Bay Lowland. Old deposits may occur along some Labrador coastal lowlands where the adjacent rugged terrain shielded them from the glaciers issuing along the fiords, as on Baffin Island.

Baffin Island

The gross physiography of Baffin Island throughout the Pleistocene was probably much the same as now. In Tertiary time an old erosion surface in central and southern Baffin Island, preserved as concordant hill-tops, was tilted, northeast side up (Goldthwait, 1950). A system of dendritic and parallel valleys resulted, and the drainage divide gradually migrated westward. Glaciers stemming from the interior developed deep outlet valleys and speeded the westward migration of the drainage-divide. The resulting broad strip of mountains along the northeast coast rise to elevations of about 6,000 feet in the Penny Highlands. Bird (1954) reports that lower erosional platforms can be recognized in western Baffin Island down to a 600-foot surface cut in relatively weak rocks. He regards the widespread horizontal surface developed on rolling Precambrian rocks as probably exhumed from beneath a cover of Paleozoic rocks of which remnants survive around Foxe Basin; stripping of this mantle may have been accomplished mainly by glaciers. He states that on Brodeur Peninsula, in northwestern Baffin Island, a well-preserved surface at 1,000 to 1,200 feet surrounds a central upland at about 2,000 feet. These surfaces had reached a late-mature to old-age stage before being elevated. These high erosional surfaces and the main tilted plateau of Baffin Island have persisted throughout the Pleistocene with but little modification by successive glaciations.

Old organic deposits (Fig. XII-12). Information on the pre-Classical Wisconsin history is limited though the occurrences of old organic materials are widespread. Information on earlier Pleistocene events has been obtained from terrain along the coast that was not recently glaciated and from organic deposits beneath drift in the interior. Baffin Coastal Lowland is several miles wide east of Barnes Ice Cap and does not appear to have been covered by ice in the last 50,000 years, though there is some evidence of an older glaciation (Løken, 1966). The last ice sheet supported active glaciers in the main fiords



1. Cape Aston
2. Cape Christian
3. Bruce Mountains
4. Isortoq River

'Old' shell sites: shells winnowed from pre-Classical Wisconsin deposits . . . . ×

FIGURE XII-12. Occurrences of 'old' organic materials on Baffin Island.

of the northeast coast and in a few places formed piedmont glaciers on the coastal lowland. Near Cape Aston, south of Clyde, a delta was built at an altitude of 262 feet, while outlet glaciers filled a nearby fiord and meltwater channels spilled over towards Cape Aston. Marine shells in the deltaic sediments, at elevation 200 feet, are indicative of a cold water environment. They were dated at >54,000 years (Y-1703). Beneath the delta, 25 feet of gravel overlies stony, shell-bearing materials that represent marine deposits ploughed-up by a glacier. Old non-glacial deposits also occur interbedded with glacial deposits in the coastal cliffs near Cape Christian, again in a protected position. Marine shells from a near-shore deposit lying between the two uppermost glacial deposits have a radiocarbon age of >50,000 years (Y-1702). In Bruce Mountains northeast of Barnes Ice Cap flattened pieces of willow and associated moss were found on a low moraine ridge; these appeared to have been winnowed from the till. They are dated at >39,600 years (GSC-209) and may be interglacial.

In the interior of Baffin Island, pre-Classical Wisconsin organic deposits have been found in several places. Fine plant detritus >40,700 years old (GSC-427) occurs in several layers of contorted fluvial sands along Isortoq River only 10 miles west of northern Barnes Ice Cap (Terasmae, *et al.*, 1966). Leaves of *Dryas, Vaccinum,* and *Ledum* suggest a climate at least as warm as the present and thus presumably interglacial. Andrews reports that the beds were contorted by ice flowing westward towards Foxe Basin. Marine shells from western Baffin Island, 100 feet above the postglacial marine limit, were dated at 30,320±820 years (GSC-528); this date may prove to be minimal.

In northern Baffin Island shells indicative of pre-Classical Wisconsin marine events were obtained from varied materials both above and below the postglacial marine limit: finite dates of about 34,200 and 35,400 years (GSC-184, 188) were obtained from Borden Peninsula, and a dating of >30,580 years on shells (GSC-189) from southern Brodeur Peninsula.

In southern Baffin Island shells found in till have given finite dates of about 30,200 and 34,800 years (GSC-414, 426). Blake (1966) regards these as minimum dates and concludes that Hudson Strait was open during the last interglacial or some early Wisconsin interstadial, and that subsequently, during an ice advance, material was scraped from the sea bottom and deposited along the south coast of Baffin Island. Old shells, washed from till, have been found also in postglacial beach deposits (GSC-468).

Some finite radiocarbon age-datings from Baffin Island may indicate an interstadial interval the lower limit of which is unknown. This is not yet proven, but neither do the numerous infinite age-datings prove that the relevant deposits are interglacial. It appears that both types of interval may be represented by the organic-bearing deposits of Baffin Island. However, undoubted glacial till overlying a fossiliferous deposit has yet to be found.

Arctic Islands

The nature of glaciation and the physiography of the arctic islands differ greatly from one region to another. Except for part of Banks Island, the southern islands were overridden by the Laurentide Ice Sheet so that the early Pleistocene record was removed or lies buried by the younger drift and fresh glacial features characterize the landscape. Some of the western islands beyond the northwestern limit of the Wisconsin Laurentide Ice Sheet were covered by an older Laurentide ice sheet. The western part of the Queen Elizabeth Islands is a lowland and includes areas of old and young glacial deposits, which may be related to both regional and local glaciers, and also some areas lacking obvious evidence of former glacial activity (Fyles and Craig, 1965). The main, eastern part is a region of mountains and plateaux that developed their own ice cover and have been extensively scoured. There is, nevertheless, some evidence of the older Pleistocene record.

In the eastern Queen Elizabeth Islands the mountain ranges, plateaux, and lowlands had assumed their present form by the beginning of the Quaternary. High-level terraces and pediment-like surfaces, with associated wood-bearing alluvium on Ellesmere and Axel Heiberg Islands are regarded as remnants of a system of mature preglacial valleys that may perhaps be early Pleistocene. Subsequent fluvial and glacial erosion have carved fiords and inner valleys as much as 1,800 feet below the old valley levels and in some lowland areas underlain by soft rocks have locally reduced the preglacial deposits to hilltop remnants. In the western islands, however, the landscape was reduced to a northwest-sloping surface of low relief when the latest Tertiary and early Quaternary sediments of the Beaufort Formation accumulated on Arctic Coastal Plain. Islands and straits (formerly valleys) comprising the western part of the archipelago are believed to be younger than the Beaufort Formation (Craig and Fyles, 1960) and to be the result of faulting and repeated cycles of fluvial and glacial erosion (Morley and Fortier, 1956; Thorsteinsson and Tozer, 1961).

Preglacial Deposits

Unconsolidated, wood-bearing alluvial sediments of the Beaufort Formation occupy most of the Arctic Coastal Plain from Banks Island to Meighen Island and are believed to extend seaward beyond the present coast; they occur also as isolated patches on hilltops to the east. These beds are tentatively regarded as late Tertiary to early Quaternary. Fyles (1965) has suggested that certain high-level terrace deposits on Axel Heiberg and Ellesmere Islands are also Beaufort equivalents (Pl. XII-6). The Beaufort Formation on Ellef Ringnes and Borden Islands consists of a lower unit, mainly brown silt with



PLATE XII-6
High-level terraces with Beaufort-type sediments overlying Cretaceous bedrock, eastern Axel Heiberg Island, Northwest Territories. View southward from Buchanan Lake. Only the uppermost light coloured material, up to 40 feet thick, is considered Beaufort type. Surface is strewn with a few glacial erratics.

peaty layers, and an upper unit of sand or fine gravel with driftwood. Only the upper unit has been observed on Prince Patrick and Banks Islands where logs of wood up to 2 feet in diameter have been observed.

The Beaufort Formation characteristically contains pollen of spruce, pine, birch, alder, various herbaceous plants such as the ericaceae, and spores of ferns and moss (Fyles and Craig, 1965). Many samples also contain small amounts of pollen of hemlock and of the temperate climate hardwoods such as hazel, beech, elm, hornbeam, and oak. It is thought that much of the hardwood pollen may be secondary, derived from nearby Tertiary strata, but some is definitely not. The pollen from many species other than the hardwoods comprise an imposing array of plants that grew in what is now a treeless, barren country. The assemblage indicates a climate not only much warmer than the present but also warmer than might be expected in any of the interglacial intervals and may be representative, at least in part, of preglacial earliest Quaternary.

On Meighen Island, which has an ice cap at present, the wood-bearing sands of the Beaufort Formation overlie unconsolidated, marine clay and silt (Thorsteinsson, 1961). On the basis of pollen content, which indicates a warm climate, the marine unit is Tertiary, but the sand unit may be early Quaternary. The sand underlies a boulder veneer presumed to be of glacial origin (Fyles and Craig, 1965). The Beaufort Formation on Arctic Coastal Plain of Borden Island is particularly interesting in that hardwood pollen is abundant and the remoteness of known Tertiary rocks suggests a primary rather than a secondary plant assemblage. The wood-bearing sands on Ellef Ringnes Island are believed to be inland, high-level parts of the Beaufort Formation (St. Onge, 1965). The organic-bearing sediments which occur as high-level terrace deposits over large parts of Ellesmere and Axel Heiberg Islands are probably the temporal equivalents of the Beaufort Formation of the coastal plain as they have a similar pollen assemblage. These deposits, characteristically gravel, sand, and silt, have thicknesses up to 200 feet and commonly lie 300 to 2,000 feet above the valleys. They appear to be remnants of broad mature valleys. They commonly contain wood and locally exhibit buried soil profiles, and enclose layers of moss or peat, and peaty beaver-pond deposits with gnawed wood (Fyles and Craig, 1965).

On northwestern Banks Island the Beaufort Formation lies at the surface (Fyles and Craig, 1965). In the higher, eastern part of the island it is only a few feet thick and rests on what appears to be an eroded Tertiary sand. In the lower central part the deposits are thicker and their base is not exposed, as on Ballast Brook where 300 feet of sand and gravel with much wood is underlain by 10 feet of peat which in turn rests on sand and silt containing plant detritus.

Early Glacial and Non-glacial Events

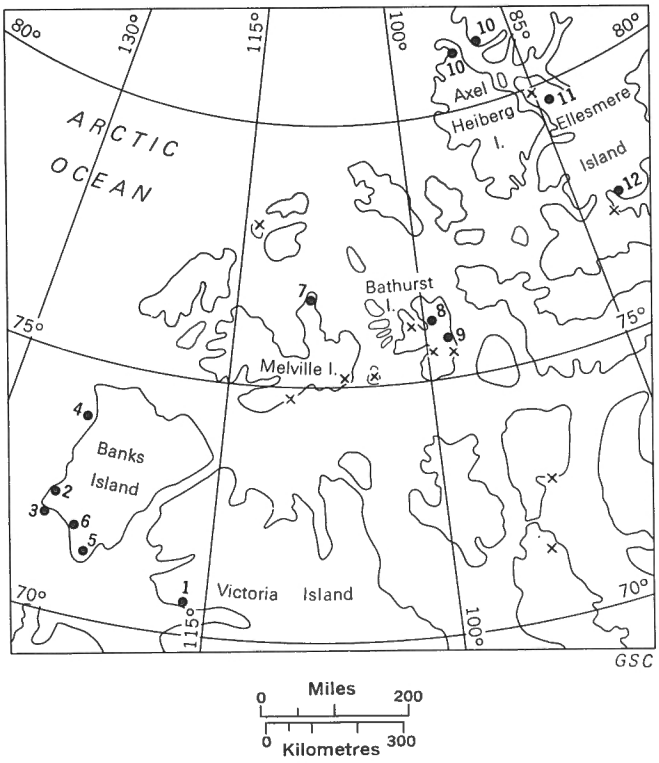
Terraces and plains within the areas underlain by Beaufort Formation consist of reworked sand and gravel. On Prince Patrick Island, most if not all of these terraces are younger than the northward-trending faults cutting the Beaufort Formation (Fyles, 1965). They indicate alluviation during some pre-Classical Wisconsin non-glacial period.

Parts of Prince Patrick, Eglinton, and Melville Islands, and most of northern and western Banks Island, bear

deposits of an older glaciation, and particularly of far-travelled erratics from the Canadian Shield. Study of buried organic deposits has shown that both interglacial and interstadial intervals are probably represented.

Buried Organic Deposits and 'Old' Shell Occurrences (Fig. XII-13)

Victoria and Banks Islands. Fyles (1963a) reports subtill organic-bearing sediments from the north side of Prince Albert Sound in western Victoria Island. Locally gravels with knob-and-kettle topography rest on 30 feet of till overlying 150 feet of stratified sediments that in turn rest on some 50 feet of dense till. In the lower part of the intertill sediments, 20 feet of thin-bedded silt contains closely packed mats of leaves and other remains of small plants which were radiocarbon dated at about 28,000 years (I-GSC-30). One of the plant layers contains sparse pollen of grasses, sedges, and other herbaceous plants. Fyles considers these intertill deposits to have been deposited by rivers and in lakes prior to the last glacial invasion under climatic conditions not greatly different from the present. Similar sediments, lacking organic remains, lie beneath till throughout large parts of northwest Victoria Island and the adjoining coastal region of Banks Island and may be correlative.



1. Prince Albert Sound, Victoria Island
2. Worth Point, Banks Island
3. Duck Hawk Bluff, Banks Island
4. Bernard Island, off Banks Island
5. Nelson Head, Banks Island
6. Masik Valley, Banks Island
7. Barrow Dome, Melville Island
8. Stuart River, Bathurst Island
9. Goodsir Inlet, Bathurst Island
10. Nansen Sound, Ellesmere and Axel Heiberg Islands
11. Slidre River, Ellesmere Island
12. Makinson Inlet, Ellesmere Island

'Old' shell occurrences; various islands. . . . . ×

FIGURE XII-13. Location of buried organic sites and 'old' shell occurrences on the arctic islands.

At Worth Point, on western Banks Island, beyond the inferred limit of the Classical Wisconsin Laurentide Ice Sheet, an uncompressed peat at the top of a 100-foot exposure of till and stratified sediments has yielded a date of >49,000 years (GSC-367). The peat contains pollen of birch as well as of plants now growing in the area. A stony layer beneath the peat is considered to be colluvium, for beneath this layer are pond silts with layers of only slightly compressed peat and moss. The peat contains wood of small trees, and some beaver-gnawed sticks. The wood gave an infinite radiocarbon age (I-GSC-19). These deposits are believed to be interglacial. At Duck Hawk Bluff, near the top of a 125-foot shore cliff, peat and willow stems occur in silts beneath colluvium or possibly till; the organics also gave an infinite age (GSC-238). The peat contains pollen of alder, willow, spruce, and herbaceous plants and is considered to be an interglacial deposit. The silts rest on an older till that overlies the Beaufort Formation. Another very similar section occurs on Bernard Island, farther north off the west coast of Banks Island. There moss peat from lacustrine deposits contained pollen of conifers as well as of tundra vegetation like that now found in the area. The peat gave an infinite radiocarbon age (I-GSC-28) and the deposit is believed to be interglacial interval.

On southern Banks Island, near Nelson Head, willow wood was found in a sand, gravel, and silt unit beneath 150 feet of till and glaciofluvial gravel and sand. This was dated >41,600 years (GSC-222). The overlying drift dates from the last major glaciation but the organic deposits may be either interstadial or interglacial. In Masik River valley, beneath till, Fyles also has observed spruce wood and layers of moss and peat associated with silt, sand, and gravel. He tentatively considers these organic deposits to date from an interglacial interval.

Melville Island. On northern Melville Island, where little vegetation grows, 2.5 feet of moss and sand was found 6 feet below the surface of a flat-topped hill, at about elevation 300 feet. The deposit appears to predate both glaciation and dissection of the lowland but to post-date the formation of the higher level terraces tentatively considered to be of Beaufort age. The moss, dated at >38,600 years (GSC-422), is probably interglacial.

Shell fragments from raised beach deposits in southeastern Melville Island that were expected to be post-glacial yielded a radiocarbon date of about 34,050 years (GSC-154). The date is probably minimal and relates to a pre-Classical Wisconsin non-glacial interval.

Also on adjacent Byam Martin Island a shell sample from an outwash deposit was presumably derived from older materials as it yielded an infinite age-dating (GSC-357).

Bathurst Island. On Stuart River, in northern Bathurst Island, postglacial marine silty-clay overlies a terrace within which 4 feet of peat overlies 20 feet of gravel of local derivation. This rests on the eroded surface of a remotely derived white sand possibly of Tertiary age. The gravels are presumed to represent a period of aggradation during some pre-Wisconsin interval during which the sea level was higher than at present. This was followed by a relatively long period of stable conditions during which 4 feet of peat formed in the valley. The peat has been dated at >36,000 years (GSC-165).

Near Goodsir Inlet in east-central Bathurst Island, peat dated at >35,000 years (GSC-178) was found beneath till and as pods within it. Blake (1964) regards the peat to be interglacial although a Wisconsin interstadial is possible. He reports that the till is similar to that from which marine shells have been collected at sites above the postglacial marine limit. Finite datings of about 35,900 and 33,940 years (GSC-212, 378) on shells from sites in central part of island are thought to be minimal.

Ellesmere and Axel Heiberg Islands. J. G. Fyles has observed terrace deposits that are at lower levels than the Beaufort terraces, and he tentatively considers these to be interglacial. In the vicinity of Slidre Fiord on west-central Ellesmere Island he found, associated with a gravel terrace, sedge and moss peat that have been dated at >41,200 years (GSC-268). Boulders on the gravel terrace are probably an indication that it has been glaciated. Glaciation is believed responsible also for the dissemination of old marine shells on similar upland surfaces south of Slidre Fiord and elsewhere on Ellesmere and Axel Heiberg Islands both above and below the postglacial marine limit; these have yielded radiocarbon dates in the 20,000 to 40,000 year range, which, due to possible contamination by old shells, are considered minimal. They provide evidence, however, of marine events prior to the last major glaciation.

Along Nansen Sound, on both Ellesmere and Axel Heiberg Islands, fossiliferous marine sediments, considered to be interglacial, occur well above the highest postglacial marine features and locally they appear to be overlain by glacial drift (Fyles and Craig, 1965). The shells have yielded radiocarbon datings ranging between 35,000 and 40,000 years (GSC-65, 113, 149), which are considered minimal.

Sandy moss peat exposed in the base of a meltwater channel near the head of Makinson Inlet in southern Ellesmere Island overlies bouldery gravel and is overlain by boulders. The peat was dated at >36,400 years (GSC-140), and along with the gravel and boulders is believed to be a remnant of a high-level terrace that may be of early glacial or interglacial age.

Shells from the sandy ground surface at the marine limit on Swinnerton Peninsula, on the southwest side of the inlet, were radiocarbon dated at 29,800±200 years (GSC-134). These shells are believed to record a marine episode prior to the last glaciation.

Cordilleran Region

The Pleistocene history of the Cordillera is somewhat more varied than that of the continental interior in that parts remained unglaciated and other parts were glaciated during only one or more periods prior to the last major glaciation, presumed to be the Classical Wisconsin. The latter, though not everywhere as extensive as older glaciations, did cover the greater part of the Cordillera. Geomorphological studies are limited but the presence of buried valleys and thick sequences of buried glacial and non-glacial sediments do bear testimony to major drainage changes resulting from successive glacial and non-glacial intervals. The record of older Pleistocene events may be preserved in its entirety in the unglaciated parts of the Yukon or in adjoining areas that were covered only by the older glaciers. At present data are sparse. The older Pleistocene record is also fragmentary in the region covered by the last major glacier complex, but locally there is much data bearing on the older events.

Unglaciated Areas

Unglaciated areas in the Cordilleran Region are shown on the Glacial Map of Canada (Map 1253A). A small area in southwestern Alberta lies above and beyond the limit of Cordilleran valley glaciers from the west and Laurentide Ice Sheet from the east. Cordilleran glaciers were restricted in their development on the dry, eastern flank of the Rocky Mountains. The rugged terrain of this unglaciated area is unsuited to the preservation of non-glacial sediments that record early Pleistocene history and climate. Unglaciated areas of very irregular outline also occur in parts of Mackenzie Mountains and Liard Plateau. These existed between the Cordilleran and Laurentide glaciers because of a combination of climatic and topographic factors. There is no information on the Pleistocene stratigraphy of these regions.

By far the largest unglaciated area in Canada is in western Yukon. This area remained unglaciated throughout Pleistocene time as it lies in a dry belt east of the high St. Elias Mountains where most of the moisture in Pacific air masses is precipitated. During glacial intervals western Yukon was both dry and cold and sustained flora and fauna less varied than that of the dry and warm interglacial intervals or that of the present. The record of glacial, interglacial, and interstadial intervals must be sought therefore in the non-glacial sedimentary sequence. To date there is little specific information.

O. L. Hughes states that on Porcupine River, in Old Crow Basin, modern peat mantles 29 feet of silt and clay

which he refers to a period of meltwater discharge across Richardson Mountains from a Laurentide ice sheet. Beneath this unit is some 144 feet of brown sand and gravel from which wood was radiocarbon dated at >41,300 years (GSC-199). The older unit also contains pollen indicative of an interglacial interval. This sedimentary sequence records the present non-glacial interval, an older glacial interval, and a still older interglacial interval, but the precise chronology remains unknown.

Throughout most of the unglaciated Klondike area O. L. Hughes has found that the base of the modern peat ranges in age from about 9,000 to 11,000 years B.P., whereas the subsurface organic materials associated with silt and gravel, encountered in placer operations are generally too old to date by radiocarbon methods. At one site on Hunker Creek, however, wood from the base of 20 feet of silty peat was dated 9,520±130 years (GSC-73), whereas wood from 4 feet below this in frozen silt gave a finite date of ca. 30,800 years (GSC-88). Sediments of this mid-Wisconsin interstadial interval were no doubt deposited in many parts of the unglaciated Yukon.

Older Glaciated Areas

In many parts of the Cordillera there is evidence that one or more early glaciations reached higher altitudes than the last major glaciation or extended beyond its outer limits (Map 1253A). Fyles (1963b) reports that elevation of the uplands of southern Vancouver Island and consequent development of narrow stream channels took place prior to deposition of any of the surficial deposits found there at present. These materials and the glacial features are believed to date from mid-Pleistocene time. Pleistocene glaciers overrode the region and extensively modified the existing valleys and elsewhere smoothed and rounded the major topographic features. The area bears evidence of at least two separate glaciations. In the Okanagan Range of Cascade Mountains in south-central British Columbia, ice at one time overrode mountains up to 8,500 feet whereas the last ice sheet did not reach over 7,500 feet. In the Foothills of southwestern Alberta an area of older glaciated terrain surrounding the unglaciated area lies about 500 to 1,000 feet above the Wisconsin limit of both Cordilleran and Laurentide ice.

In Mackenzie and Selwyn Mountains most of an elongate, partly unglaciated region bears evidence of one or more early Pleistocene Cordilleran as well as Wisconsin Cordilleran glaciations. The early glaciers originating in western Mackenzie Mountains and eastern Selwyn Mountains apparently filled the intermontane valleys to higher levels than did the Wisconsin glaciers. The fringe zone of older glaciations is thus very complex and at present remains undifferentiated (Map 1253A). The region lies above and beyond the western limits reached by Laurentide Ice Sheet. The Cordilleran glaciers, however, on flowing northward towards the Arctic Red and Peel Rivers encountered the older, southward-moving, Laurentide Ice Sheet. In this area a zone of older glaciated terrain is both extensive and distinct, but the interrelations of the Cordilleran and Laurentide ice masses are uncertain.

A zone of older glaciated terrain, some 10 to 20 miles wide, lies west of Peel River between the unglaciated part of the Yukon and the western limit of Wisconsin Laurentide Glaciation. Glacial landforms are relatively uncommon in this area, but areas of subdued hummocky terrain occur.

In southwestern Yukon, north of St. Elias Mountains, an area of greatly variable width separates unglaciated terrain from areas to the east and south that were glaciated during the Classical Wisconsin. Several glaciations have been recognized in this fringe zone (Denton and Stuiver, 1967; Bostock, 1966). It appears also that older glacier complexes in the interior mountainous parts of the Yukon attained greater thickness than the Wisconsin ice. Some peaks in the dry central part of the Yukon that protruded above the Wisconsin glacier surface were covered by one or more older glaciers and bear a fringe of older glaciated terrain. Near Kluane Lake on the northeast side of St. Elias Mountains, on the other hand, two older glaciations, the Shakwak and Icefield, were not so extensive as the Kluane or Classical Wisconsin Glaciation. The older glaciers extended 70 to 75 miles northeast of the present Pacific–Interior ice divide, and reached the eastern part of Kluane Lake basin. The last major ice sheet filled Kluane basin and extended 90 to 100 miles northwest along the valley to the vicinity of Snag, Yukon, with an ice tongue extending 20 miles northwards down White River (Bostock, 1952). There, an older glacier was responsible for drift extending another 20 miles down the river. Correlation between the older drifts of Kluane area and that of White River valley has not been made, but apparently the latter is older than either of the former and probably is pre-Illinoian.

In central Yukon two clearly defined glacial limits are evident and may be readily correlated and traced (Bostock, 1966). They are considered to mark the limits of the last and an earlier glacier advance. In addition, older glacial drift, including till and erratics, and modified glacial landforms have been found beyond the limit of the earlier advance and may record two still earlier glaciations. In Stewart Valley the McConnell was the last major glaciation. The youngest of the older drift units, the Reid extends westward down the valley at least 40 miles beyond the McConnell limit; it displays distinct glacial features, including an end moraine. West of the Reid End Moraine, in lower Stewart Valley, glacial landforms are generally lacking but grey, silty to clayey till with mostly fresh stones and scattered erratics represents the older Klaza Glaciation.

The Klaza glaciers extended at least 25 miles beyond the Reid limit. In some other valleys also, subdued glacial landforms related to the Klaza glaciers have been observed. In lower Stewart Valley the oldest glacier extended about 25 miles beyond Klaza glacial limit and reached to within

20 miles of Yukon River; it is referred to as the Nansen Glaciation. It is expressed by a great thickness of drift that has a hummocky to undulating surface or is locally terraced. This drift blocked the channel of a creek, which then cut a canyon 300 feet deep in bedrock. In some valleys Bostock noted that the Nansen drift is extensively weathered; the stones in the drift are rotted, and usually both till and gravels are brown. Till beneath a lava flow, 3 miles downstream from Fort Selkirk on Yukon River, is believed to be related to the Nansen Glaciation but may be older. On the north side of Stewart River valley, the Nansen glacier appears to have reached the altitude of a low pass in the valley side so that meltwaters flowed northward depositing a distinctive gravel train along the valley of Australia Creek and Indian River as far as their junction with Yukon River about 40 miles below the mouth of Stewart River.

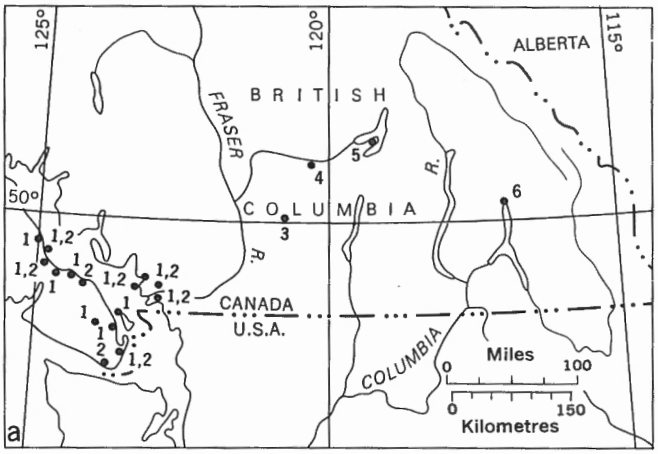
Bostock states that in upper Stewart Valley the profiles of the ice surfaces of the major glacier advances rise eastward gently and converge in elevation so that the moraines of the McConnell advance are so close to those of the earlier advances that they are virtually indistinguishable. On Talbot Plateau, southeast of Mayo, the McConnell and Reid Moraines reach altitudes of 4,000 and 4,400 feet respectively, and scattered erratics believed to be remnants of the Klaza Glaciation occur up to an elevation of 4,700 feet. Two small buttes rising above 4,700 feet elevation may have been nunataks. Other moraines are present in the area some of which may reflect oscillations of the ice fronts during the older glaciations.

In west–central Yukon, Vernon and Hughes (1965) found scattered evidence of one or more old glaciations beyond or above a clearly defined glaciated terrain of an intermediate glaciation which, in turn, lay beyond or above the deposits of the last major glaciation. Little is known of the oldest glaciations other than the occurrence of scattered erratics in otherwise apparently unglaciated areas. The trace of large transection glaciers, formed during the intermediate glaciation, have been delineated in the eastern part of the region and their gradients over long stretches have been determined as from 19 to 35 feet to the mile.

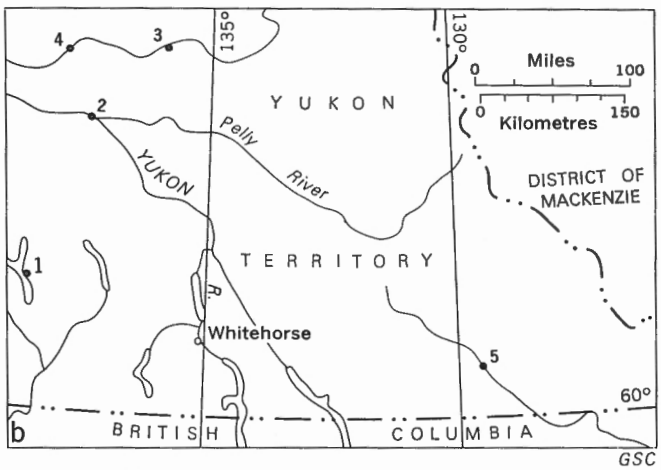
Buried Organic Deposits of the Main Glaciated Areas

Both interglacial and interstadial deposits are recognized in southern British Columbia and perhaps are the equivalents of the Sangamon and mid-Wisconsin intervals. In Yukon the extent of several Pleistocene ice sheets have been delimited but only in one area has it been possible, tentatively, to assign the buried organic deposits—and hence their associated glacial deposits—to any particular Pleistocene period. The location of buried organic deposits in southern British Columbia and the Yukon Territory are shown on Figures XII-14 (a) and (b) respectively.

Straits of Georgia and Juan de Fuca. In the Strait of Georgia region subtill organic-bearing deposits, known as the Quadra sediments, have been assigned to the Olympia Interglaciation which preceded the last major or Fraser



1. Vancouver Island and Fraser Lowland; Quadra sediments
2. Vancouver Island and Fraser Lowland; pre Quadra sediments
3-5. Thompson River valley and tributaries; younger and older intertill sediments
6. Kootenay Lake; palaeosol and intertill sediments



1. Kluane Lake; younger and older intertill sediments
2. Fort Selkirk area; sediments beneath striated lava
3. Mayo Landing area
4. McQuesten area
5. Liard River

FIGURE XII-14. Location of buried organic deposits in the Cordilleran Region: (a) Southern British Columbia; (b) Glaciated parts of Yukon Territory.

Glaciation (Armstrong, *et al.*, 1965). For comparative purposes, these may equate, on the basis of radiocarbon datings, with the mid-Wisconsin and Classical Wisconsin of the mid-continent and Great Lakes region. Recent investigations by J. G. Fyles and E. C. Halstead have shown the wide distribution of Quadra sediments on southeastern Vancouver Island. According to Fyles (1963b), the Quadra sediments comprise a lower unit of clayey silt with stones and marine shells, a middle unit of plant-bearing silt, gravel, and sand, and an upper unit of white sand with local beds of gravel and plant-bearing silt. Maximum thickness of these units is about 80, 40, and 250 feet.

The lower unit is considered to represent a transition from glaciomarine to marine conditions. The middle, plant-bearing unit appears to have originated in a swampy coastal lowland during, and also probably following, a regression of the sea in which the underlying marine clay had been deposited. The upper unit is a fluvial-plain deposit characterized by cut-and-fill structures; it contains beds of gravel and plant-bearing silt composed largely of debris from the Coast Mountains of the mainland. Fyles suggests that the present Strait of Georgia may have been completely filled with sediments prior to the Fraser Glaciation.

Pollen and marine molluscs in the Quadra deposits record a climate cooler than the present and somewhat comparable to the present climate of the Gulf of Alaska. The vegetation differs from that on Vancouver Island today in rarity of Douglas fir and abundance of spruce. Radiocarbon dates range from about 20,000 to perhaps 50,000 years B.P. Thus, either the lower limit of a very lengthy non-glacial interval remains to be established or two separate intervals are represented by deposits currently assigned to the Quadra on Vancouver Island.

Quadra sediments on Vancouver Island in places overlie a till and associated sediments, the Dashwood drift, which in turn rest on the non-glacial Mapleguard sediments. At Icarus Point, E. C. Halstead states that peat in the basal part of Quadra-type sediments overlies a till that rests on peat, silt, and silty sand. The lower peats were dated at >37,600 and >36,650 years (GSC-155, 191). These older sediments overlie some 10 feet of stony marine clay, which overlies a third till exposed near beach level. Pre-Quadra sediments containing organic matter also occur beneath till at Cordova Bay, Victoria. According to J. G. Fyles, these sediments contain a pollen assemblage that is believed to represent true interglacial conditions, perhaps the Sangamon Interval. Equivalent deposits that include peat and wood-bearing strata occur in coastal exposures west of Sooke in Juan de Fuca Strait and have been dated at >40,300 years (GSC-358).

Both Quadra and pre-Quadra organic-bearing sediments have been noted also by J. E. Armstrong near Lynn Creek in Fraser Lowland. Quadra sediments, dated 36,200±500 years (GSC-92-2), overlie a till that rests on an older sand, gravel, and peat unit from which wood has been dated >52,300 years (GSC-555). These older sediments overlie an older basal till.

The best-documented and most complete Pleistocene section in Fraser Lowland, if not on the whole west coast, is that at Point Grey, Vancouver (Table XII-2), where some 130 boreholes and a long tunnel have added immeasurably to data available from study of the seacliffs.

Quadra and pre-Quadra sediments occur respectively also in Coquitlam and Surrey municipalities east and southeast of Vancouver.

Southern interior, British Columbia. Fulton (1965) cites two distinct intertill sequences in the valley of Thompson River and its tributaries. The lower consists of oxidized sand, silt, and gravel containing volcanic ash, together with wood and freshwater shells. Shells from near Merritt gave a dating of >37,200 years (GSC-258). Wood and shells from Kamloops gave ages of >32,700 and >35,500 years (GSC-275, 413). The upper sequence, also sand, silt, and gravel but unoxidized, is correlated with deposits that at Salmon Arm, Shuswap Lake, contain wood dated at 20,230±120 years (GSC-194). An erosional unconformity separates the two sequences. Till and glaciolacustrine deposits, locally exposed beneath the older oxidized sediments, indicate that at least one glaciation preceded the older non-glacial interval. The intertill sediments were deposited prior to the Fraser Glaciation during a non-glacial interval which comprised at least two periods of aggradation separated by an interval of oxidation and soil formation.

Fulton (1968) indicated rather similar geological events in Purcell Trench, north of Kootenay Lake. He has demonstrated the occurrence of two glacial sequences separated by non-glacial sediments that contain organic matter and a paleosol. An unconformity separates the older till and varied associated deposits from the younger sediments. The paleosol, which includes A, B, and C horizons, is developed on some of the older materials. A finite age of 41,900±600 years (GSC-733) was obtained on roots embedded in the A horizon. The younger non-glacial sediments represent deposits of an aggrading flood plain. Two radiocarbon dates on materials successively nearer the top of the flood plain unit were 43,800±800 and 42,300±700 years (GSC-740, 720). An age of 41,800±600 years (GSC-716), furthermore, was obtained from a stump growing on a slope facies of the

TABLE XII-2 *Composite section of Pleistocene deposits, Fraser Lowland (by J. E. Armstrong)*

	(Max. known thickness, feet)
Stratified sediments (postglacial)—stream and marine	70
Till and associated sediments (Fraser Glaciation)—glacial, glaciofluvial, glaciolacustrine, glaciomarine	215
Non-glacial (Quadra) sediments (Olympia Interglaciation)—swamp, flood plain, channel and estuarine; peat in basal part dated >36,800 years (GSC-81)	200
Till and associated sediments (Semiamu Glaciation)—glacial, glaciomarine, marine	130
Non-glacial sediments (interglaciation?)—swamp, lacustrine, flood plain, channel, marine; includes some peat	155
Till and stratified deposits (glaciation)—glacial, glaciomarine; includes shells	15
Non-glacial sediments (intertill) channel and flood plain	15+
Till?	—
Bedrock (Eocene)	—

paleosol but embedded in the younger gravels. Silty sediments that intertongue with the gravel yielded wood that gave dates of 33,700±330 and 32,710±800 (GSC-542, 493) from successively higher positions. A coarse gravel overlying the silty sediments contained wood in its upper part that dated 25,840±320 (GSC-715). The gravel is capped by till. Fulton relates the upper till to the Fraser Glaciation which, judging by dates from Kamloops region, did not reach the Kootenay area until after about 20,000 years B.P. He relates the non-glacial sediments to the Olympia Interglaciation of the west coast. The paleosol was developed following the older glaciation and was successively buried by the interglacial sediments until perhaps 41,800 years ago.

Central and northern interior, British Columbia. In central and northern British Columbia interglacial sediments have been reported but their interpretation is doubtful. In places non-glacial intervals are indicated by buried river channels cut into older tills or bedrock and by the occurrence of buried placer deposits. Lignite or peat occurs in gravel beneath till along Stikine and Tuya Rivers, but the origin of the deposits is unknown. The indicated non-glacial intervals must predate the last major glaciation but whether earlier Wisconsin, pre-Wisconsin, or pre-Pleistocene has not been determined. In many parts of central British Columbia the Tertiary deposits are unconsolidated and not readily distinguishable from Pleistocene sediments. The occurrence of lava flows or sills in unconsolidated sediments is more characteristic of the Tertiary, but as lavas overlie till sheets in some places, volcanic activity must have continued into the Pleistocene. Lava flows associated with the tills or other Pleistocene sediments may serve as a means of dating some events and provide information on the older Pleistocene.

Interior Yukon Territory. Within the glaciated part of the Cordilleran Region in the Yukon few occurrences of buried organic materials have been reported, but undoubtedly many more will be found as the region is studied.

In Kluane Lake area on the northern side of St. Elias Mountains organic materials occur in drift that predates the last major glaciation, the Kluane, and, on the basis of geomorphological and stratigraphic evidence, indicate three separate non-glacial intervals (Denton and Stuiver, 1967). Organic debris in outwash related to the oldest recognized glaciation, the Shakwak, and organic debris from a silt bed beneath ice-contact stratified drift of the younger Icefield Glaciation, and peat from sinuous stringers contained in the Icefield till itself, proved too old for radiocarbon analyses (Y-1355, 1481, 1486). Organic debris in outwash overlying the Icefield till, however, yielded three finite dates ranging from 30,100 to 37,700 years (Y-1356, 1385, 1488). These deposits are overlain by the Kluane till.

The finite datings appear to represent a non-glacial interval, the Boutellier, of about 10,000 years duration which may be the equivalent of the mid-Wisconsin interval of the mid-continent. The Silver non-glacial interval between the Icefield and Shakwak Glaciations, from which the older age-datings were obtained, was much longer than the Boutellier interval judging by differences in the depth of oxidation of the underlying materials, but neither the duration nor the climate of the interval is known. It may be either early Wisconsin or Sangamon; if the latter, then the Shakwak Glaciation and its outwash deposits may be Illinoian.

Near Fort Selkirk on Yukon River a Pleistocene lava with a striated surface, overlies gravel, sand, and silt that in turn overlies glacial till. Charred wood found 4 feet from the top of the bedded unit was dated at >38,000 years B.P. (I-GSC-27). Another record of forest cover predating the last glaciation was found near Mayo on Stewart River. An abraded log was found in the base of a till lens occurring near the top of a 100-foot section of sand and gravel, the whole overlain by bedded silt. The till lens at this site is considered to be the same as the upper till at numerous sites along the river nearby, and to represent the glacier which apparently terminated some 8 miles below Mayo Landing. The wood was dated at >35,000 years (I-GSC-180). Nearby, wood from a silt, sand, and minor gravel unit overlain by about 10 feet of till and 30 to 50 feet of thinly bedded silt and sand, was dated at >46,580 years (GSC-331). Also on Stewart River, below McQuesten, old wood was found in an ash lens beneath 10 to 15 feet of organic silt (GSC-342).

Further interesting evidence of former forest cover is the report of wood from a 200-foot cut-bank on Liard River. A thin soil zone, including a volcanic ash horizon, and about 100 feet of brown till overlies about 100 feet of crossbedded sand. Wood from 30 feet above river level was dated >40,100 years (GSC-412).

CLASSICAL WISCONSIN AND POSTGLACIAL EVENTS

The last continental ice sheet had three main component parts, the Laurentide Ice Sheet, the Cordilleran Glacier Complex, and the Queen Elizabeth Islands Glacier Complex.[1] The Laurentide Ice Sheet is commonly regarded, on the basis of ice-flow patterns, as comprising the Labrador and Keewatin sectors which, although confluent at the Wisconsin glacial maximum, became discrete areas of outflow during deglaciation. The former retreated to one or more centres or ice divides in northern Quebec and Labrador, and the latter to an ice divide west of

[1] The term 'glacier complex' is used in the broad sense to include ice sheet, ice cap, piedmont glacier, and valley glacier.

Hudson Bay. These major areas of late ice flow themselves split up into smaller short-lived units of outward-flowing ice prior to the final dissipation of the main remnants. Deglaciation also resulted in other major components of Laurentide Ice Sheet. An ice sheet remained and was nurtured locally in Foxe Basin and, later, on Baffin Island. This ice sheet, the Foxe–Baffin Glacier Complex, was independent of the Keewatin and Labrador sectors at least in late Wisconsin time, and parts remain today. Also, early in the deglacial process, major independent ice caps formed in the Appalachian regions, for instance, in Newfoundland, and in part may have been independent of the Labrador sector throughout most of the Wisconsin. The above ice masses and stages in their deglaciation are shown in Figure 15.

The last continental ice sheet is described according to its major component parts regardless of the degree of interdependence experienced during their build-up, at their maximum, or during decay. These are: Appalachian Glacier Complex, Labrador sector of Laurentide Ice Sheet, Keewatin sector of Laurentide Ice Sheet, Foxe–Baffin Glacier Complex, Cordilleran Glacier Complex, and Queen Elizabeth Islands Glacier Complex. Due to the size of the areas within the zone of influence of the major glacier units chosen, it may be necessary to discuss some matters according to smaller areas or to particular glacial or deglacial events. As the ice sheets and glaciers waned and the land surface was uncovered, the forms implanted by the ice were left exposed or were covered by glacial debris washed out from the receding ice margins on land, in lakes, or in the sea. The complex record of glacial and postglacial features and deposits remaining today afford a means of tracing the paths of the receding ice margins and interpreting the deglacial and postglacial events.

Appalachian Glacier Complex

The Appalachian Glacier Complex occupied most if not all of the Appalachian Region in Canada. This includes the island of Newfoundland, the Maritime Provinces, and the Appalachian Mountains of southeastern Quebec. The build-up of glaciers in this region prior to the Wisconsin maximum is little known. The physiographic province is, in large part, a highland region of abundant snowfall, which under the Wisconsin climatic conditions may have developed a number of independent ice caps early in the glacierization process. Under the influence of prevailing northeasterly moving moisture-laden winds from the continental interior and from surrounding bodies of water, these ice caps may have spread radially by differential accretion and preferentially towards the southwest. There are many records of glacial striae and grooves, and of rock dispersal, that trend at right angles to what is commonly considered the ice-flow trend at the Wisconsin maximum and during later retreat. The Appalachian Region may well have been largely ice-covered before confluence with Labrador ice was established along the St. Lawrence River and Strait of Belle Isle. Early ice movements are thus postulated as local and independent of the outward growth of the Labrador sector of Laurentide Ice Sheet. It is probable that the Appalachian Glacier Complex was comprised of two main parts, one over Newfoundland and the other over the Maritime Provinces; both were confluent with the Labrador sector of Laurentide Ice Sheet for some time during the Wisconsin but confluence with each other along Laurentian Channel, through Cabot Strait, was short-lived at the Wisconsin glacial maximum.

The pattern of deglaciation as indicated by ice-flow features of all types (Map 1253A) gives little evidence of general ice-frontal retreat towards Quebec and Labrador. The influence of the maritime environment is obvious. Only on the mainland of Nova Scotia is there a regional southeasterly trend to suggest glacier movements from, or retreat towards, a Labrador centre, but this trend could equally well indicate a centre of outflow from New Brunswick or Maine. Though Labrador erratics have been reported on the Gaspé Highlands, a significant overriding of the Gaspé and New Brunswick Highlands by Labrador ice of Wisconsin age has not been proven.

Newfoundland

Glacial events. On Avalon Peninsula, Henderson (1960) found no evidence of a general ice invasion from the west and concluded that the peninsula had its own ice cap from which active glaciers flowed northward and southward down the channels now occupied by Trinity, Conception, St. Mary's, and Placentia Bays, and thence onto the Grand Banks. A few cobbles from the main part of the island occur along the east side of Trinity Bay at about 20-to-25-foot elevation and hence close to the marine limit; this suggests that the main island ice remained active after considerable recession of the Avalon ice cap and as the sea encroached on the present shores of Avalon Peninsula.

Henderson believes that the last active glacier on Avalon Peninsula lay in St. Mary's Bay and moved debris into the central part of the peninsula as it fanned out towards Trinity and Conception Bays; free flow southward along St. Mary's Bay was impeded by a 'baymouth' threshold. Transverse lineaments in the morainal materials of central Avalon are evident on airphotos, although such patterns are not invariably convex in the direction of flow. Henderson believes that the glacier down-wasted in St. Mary's Bay with marginal retreat taking place on all sides.

MacClintock and Twenhofel (1940, p. 1731) concluded that the whole island was glaciated during the Wisconsin and that " . . . ice spread as a complete cap from the Long Range Plateau, the Central Plateau and the Avalon Peninsula outward in all directions, to beyond the present shore lines of the island." Their studies gave no indication of pre-Wisconsin events as earlier postulated by Coleman, but rather served to establish a gross chronology for the Wisconsin, based mainly on observations in St. George's Bay area in southwestern Newfoundland.

The oldest drift—"St. George's River Drift"—includes till, ice-contact gravel, and marine silt, all of which display a topography suggestive of ice blocks surrounded or buried by marine beds. Following a significant ice-frontal retreat from the shores the "Bay St. George Delta" was built, complete with marine fossils like those of the present day. With re-advance the ice overrode the delta, deposited an upper till, and built an end moraine system near the coast; these deposits comprise the "Robinsons Head Drift." A major retreat of the ice towards the central plateau areas followed, but a still-stand produced the "Kittys Brook" moraine system in the valleys. Further recession resulted in the higher knobs of the plateau protruding above the ice sheet and becoming over-steepened by the outward-moving ice. Still later the ice occurred only as local valley glaciers on the steep sides of the plateaux. Some glaciers deposited small moraines and formed cirques, many of which still contain perennial snow and ice. The work of K. Widmer in the Hermitage Bay area, on south coast of Newfoundland, shows the same close association between 'late' ice and marine overlap.

On the main part of the island there was strong outward ice flow (Jenness, 1960). Fiords are particularly well developed along the south and northeast coasts. There are large areas of fluted and drumlinized terrain, and of ribbed moraine (Pl. XII-7). The lineations, together with recorded striae, grooves, and boulder train data, suggest a very complex ice-flow pattern with erratic, shifting centres of active flow during the waning of the island ice cap. The last active ice caps appear to have been on Newfoundland Highland.

The matter of Labrador ice occupying much of Newfoundland at the Wisconsin maximum, as postulated by Flint (1940) on the basis of increasing height of raised strandlines northward along the west coast, is open to question. The writer believes that Newfoundland maintained its own active ice cap throughout Classical Wisconsin time though it was confluent with Labrador ice in Gulf of St. Lawrence during most of this period. MacClintock and Twenhofel (1940) reported that overriding Labrador ice was possible but there was no real evidence, and concluded that the problem would remain unresolved until the transportation of drift boulders was studied in terms of regional geology of mainland and island alike. Cooper (1937), however, actually records mainland erratics dispersed over the highest part of the northern end of Great Northern Peninsula and associated with southeast-trending striae. Such glaciation by Labrador ice is to be expected in such close proximity to the mainland. On aerial photographs a system of De Geer moraines along Strait of Belle Isle is evident, which indicates late southwestward flow along the strait into Gulf of St. Lawrence,

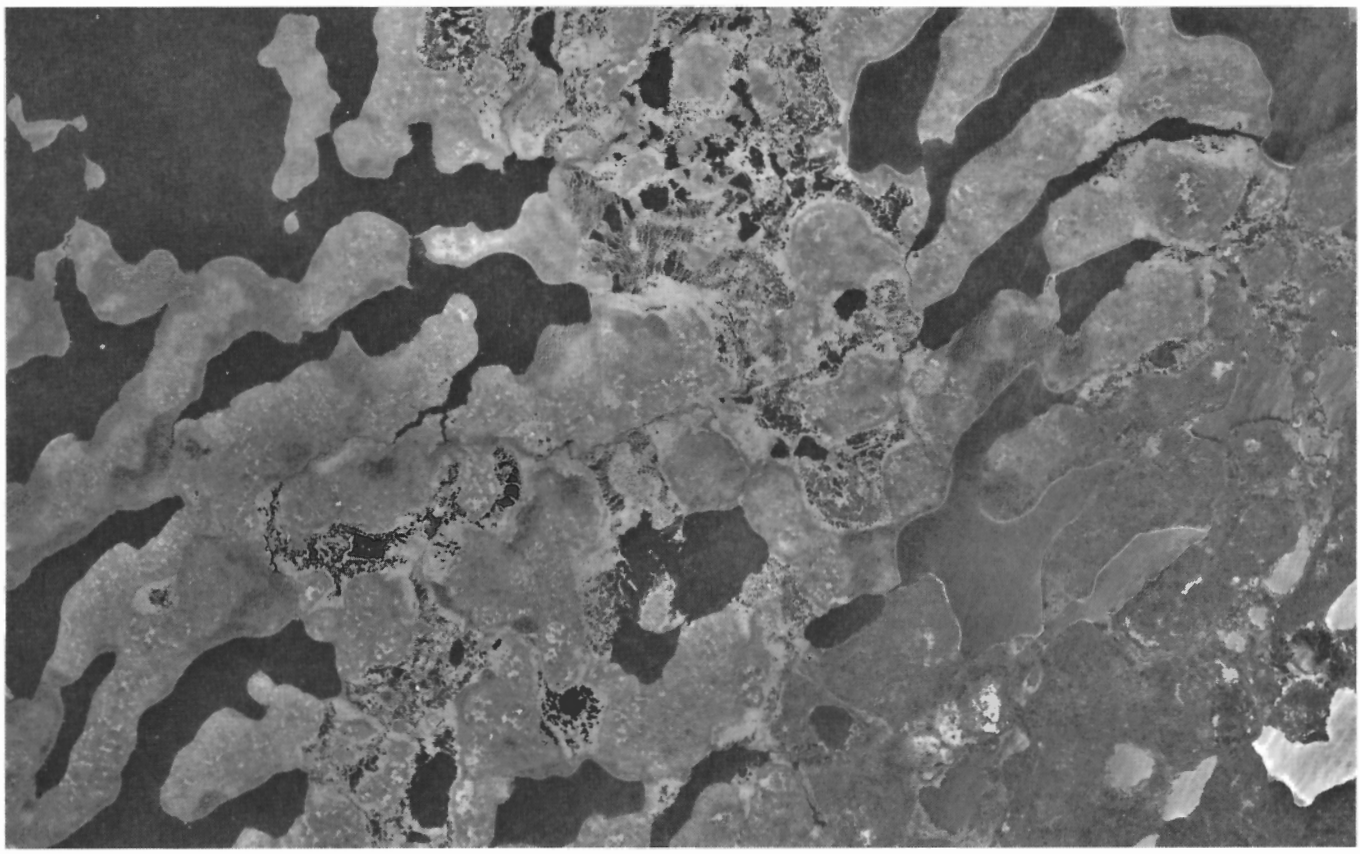


PLATE XII-7. Ribbed moraine near Meca Pond in eastern Newfoundland. Vertical airphoto. The rib ridges show no evidence of ice flow. Scale 1 inch to 1,320 feet.

probably of confluent Newfoundland and Labrador ice. As marine waters extended northeastward up the deep channel between Quebec and Newfoundland, and thence into the shallow Strait of Belle Isle, Newfoundland ice would be separated from Labrador ice. At about this stage of deglaciation, about 12,000 years ago, Mecatina Plateau northwest of Strait of Belle Isle was effectively diverting Labrador ice towards the northeast, away from the Strait of Belle Isle.

On northern Newfoundland variations in the trend of De Geer moraines suggest a short-lived active centre of outflow from the area of former Newfoundland and Labrador ice confluence at the extreme northern end of the peninsula. Late ice flow from the Newfoundland Highland changed from southwestward to northwestward as the centre of outflow shifted back to the Long Range highland and a system of lobate end moraines was formed in the vicinity of Ten Mile Lake.

The writer's observations in the central-west coast region support the concept of an active Newfoundland ice cap in late Wisconsin time. There was strong ice flow from the interior both along major valleys parallel the bedrock structures and also across rugged terrain towards the west coast.

Marine events. The interpretation of data on maximum marine overlap is complicated by the fact that 'late' ice has prevented the registration of the highest shorelines in some areas, as in southern, southwestern, and probably also northeastern coastal Newfoundland. This phenomenon implies that local as well as regional postglacial rebound is reflected in the resultant picture of marine overlap.

On Avalon Peninsula, Henderson (1960) found evidence of late-glacial marine overlap only along the east coast of Trinity Bay and on Avalon Isthmus. The isobase of zero uplift may trend northeastward along the east side of Placentia Bay, cross the isthmus at Norman's Cove, and thence pass just inland from the east coast of Trinity Bay. Henderson found that the elevation of raised marine features increased northwestward along the isthmus and also northward along the east shore of the island proper to Bonavista Bay and beyond. The 100-foot isobase passes along the southeast shore of Burin Peninsula and along the axis of Bonavista Bay. The isobase data may reflect an early period of marine overlap broadly concentric about the eastern side of Newfoundland.

The highest shoreline features decrease in elevation westward along the south coast. K. Widmer records wave-cut benches on islands at the mouth of Fortune Bay and on nearby mainland points at an elevation of 100 to 110 feet. The marine limit is about 70 feet at head of Fortune Bay and 30 feet near Burgeo. Between Burgeo and Port aux Basques raised shore features are absent. Similarly, along the southeastern side of St. George's Bay, on the west coast of Newfoundland, there is evidence of late ice advances into the sea that may have prevented the registration of high shoreline features (Flint, 1940; MacClintock and Twenhofel, 1940). Flint has reported, nevertheless, that the level of maximum marine overlap increases from zero at the southwest end of Newfoundland to 100 feet in head end of St. George's Bay, and to 200 feet in Bonne Bay, with isobases trending N70°E. At a later time, during a pause in the uplift, the sea carved benches in bedrock along the west coast and on the north coast, termed the "Bay of Islands surface." The trend of the isobases on this surface is N80°E, the zero isobase lying along the north shore of St. George's Bay, the 100-foot isobase passing through Bonne Bay, and the 250-foot isobase passing through Hare Bay. Study of aerial photographs of this area reveals another prominent bench near sea level, possibly that recorded by Cooper (1937) as a wave-cut bench at 30 feet elevation.

On the north coast raised marine features exhibit three still-stands, two of which may correlate with those recorded on the west coast. The third set may reflect the early period of marine overlap recorded along the eastern shores of the island. Marine shells from Baie Verte, considered to be related to a former sea level at 180 feet elevation, gave dates ranging from 11,520 to 11,950 years B.P. (GSC-55, 75, 87). As there is evidence of marine features and deposits well over 200 feet and perhaps up to 250 feet, it is evident that deglaciation of the north coast had begun well before 12,000 years ago. Marine shells from the west coast of Newfoundland have also been radiocarbon dated; these indicate marine invasion into St. George's Bay prior to $13,420 \pm 190$ (GSC-598), and into Bay of Islands before $12,600 \pm 170$ years B.P. (GSC-868).

Cape Breton Island, Nova Scotia

It has been presumed that the Laurentide ice invaded the Gulf of St. Lawrence and passed southeastward across Cape Breton (Goldthwait, J. W., 1924). Glacier ice certainly flowed southeastward through Cabot Strait, scouring St. Paul Island and the northern tip of Cape Breton and also passed through the Strait of Canso, but elsewhere the regional ice-flow trends are poorly preserved and little understood. The writer believes that Cape Breton Highlands were not glaciated by Labrador ice but rather by a local ice cap, and that outward ice flow was not extensive.

There is evidence, however, of a glacier in Northumberland Strait on the west side of Cape Breton Island. Valleys between 200 and 250 feet elevation appear to have been filled, as if by glacial imponding, and later terraced by stream action. E. H. Muller records a river section a mile north of Southwest Margaree where 8 to 10 feet of gravel overlies about 20 feet of red lake clays, 10 feet reddish gravel, and 5 feet of blue cohesive till. It may well be that the glacier that imponded lakes in valleys along the west coast of Cape Breton was the one in Northumberland Strait, which may represent confluent ice from the highlands of Nova Scotia and New Brunswick from the south and west.

Goldthwait (1924) reported northeast-trending striae over eastern Cape Breton and refuted an older hypothesis that this ice flow was from Newfoundland. He noted ice-flow features and drift 'tails' that unmistakably proved a northeast to northward ice flow but he was concerned as to the source area. Though the more northward-trending striae on the east coast require an ice flow from the continental shelf towards the present island he suggested a possible source area south of Strait of Canso. Lacking information on such a centre, however, and with knowledge of east and east-northeast striae along Northumberland Strait he favoured a flow from New Brunswick with, presumably, a radial expansion as the ice emerged from Strait of Canso and Chedabucto Bay. This event preceded his Acadian Bay ice lobe which he considered the main Wisconsin Glaciation. The late H. L. Cameron, however, indicated south-southwest-trending striae east of Bras d'Or Lake and an end moraine system thought to be formed by this glaciation. Some indications of former southwestward ice flow are also evident on aerial photographs (Map 1253A). The anomalous southwest ice-flow trend of eastern Cape Breton appears to be developed on a broad morainal tract. D. R. Grant (GSC) considers that this trend represents a younger ice flow that oriented the drift and lightly striated the bedrock. On the southeast side of many shore outcrops he found evidence of intense scouring action by glaciers directed towards the northeast. The older striae, grooves, and *roches moutonnées* and the trough-like topography of Chedabucto Bay suggest a funnelling of ice down the trough with northward splaying or lobing along eastern Cape Breton. The younger glacier movements appear to have been southeastward through Strait of Canso and northeastward along Bras d'Or Lake basin, and also southward and southeastward out of the lake basin towards the sea. Neither the ice-flow trends in eastern Cape Breton, their indicated source areas, nor the indicated order of glacial events are in accord with the concept of the Labrador ice overriding the island and extending onto the continental shelf.

Information on marine overlap in Cape Breton is limited. A prominent subhorizontal feature at about 200 to 250 feet is present over long stretches of both eastern and western shores. H. L. Cameron reported limited marine overlap in Aspy Bay on northern Cape Breton, and along the southeast coast, but also proglacial lakes in several valleys in-filled to about 250 feet. In view of the semi-coincidence in levels some of the raised coastal features may be ice-contact phenomena, or else both valley-fills and coastal terraces are graded to a common sea-level stand, possibly of pre-Wisconsin age. The writer has not seen any evidence of marine overlap on Cape Breton Island.

Mainland Nova Scotia

Glacial events. The pattern of ice-flow features in Nova Scotia south of the Cobequid Mountains is orderly; at the Wisconsin maximum the ice moved southeasterly across the region. This ice-flow direction is exhibited in many places by drumlins composed of fine-textured till. In areas underlain by granite or quartzite the till is silty to sandy and light coloured. In a zone along LaHave River the till is olive-grey and derived from local slates. Elsewhere to the northeast the drumlins are composed of an anomalous red till that overlies locally derived tills, and is believed by Grant to have been derived largely from Pleistocene sediments scoured from Bay of Fundy, Minas Basin, and Northumberland Strait. L. H. King (The Bedford Institute) located off-shore moraines that may represent the Wisconsin maximum or a somewhat later stage of retreat. He believes that the off-shore moraines probably formed at a time of low sea level when relatively thin ice calved into marine embayments.

As the Wisconsin glacier waned the ice front along the Atlantic coast retreated inland to the northwest. Absence of marine overlap along this coast suggests a relatively thin ice sheet near its terminus. As deglaciation proceeded marine waters invaded the Bay of Fundy giving rise to a concentric pattern of retreat towards the Nova Scotia Uplands. Late stage upland centres of outflow were active. Drumlins have been re-oriented in some areas and eskers appear at variance with earlier ice-flow features. Late ice in the central upland region remained active or was reactivated so that it moved granitic drift northward across Cornwallis–Annapolis Valley and deposited it on the Triassic trap rocks of North Mountain, and along the Fundy shore (MacNeill, 1953; Hickox, 1962). That the central Nova Scotia Uplands should retain an ice cap late in the period of deglaciation is not surprising in view of the maritime setting; it is at present an area of heavy snowfall. The precise age of this upland ice cap is not known and there are currently no dates on relevant materials, but it is known that the sea invaded the western side of Bay of Fundy prior to $13{,}325 \pm 500$ years B.P. (I-GSC-7).

The pattern of ice-flow features in northern Nova Scotia is highly irregular. The Cobequid Mountains and Antigonish Highlands appear to restrict the general southeast flow so typical of the region to the south and east. There was some glacier flow northward from these highlands towards Northumberland Strait. Glacial striae and boulder trains indicate northward movement in the vicinity of Pictou and New Glasgow. Cobequid-type igneous stones occur sporadically to the north of the mountains in the red sandy tills of the lowlands; they are especially noticeable near Pictou and on Pictou Island. Goldthwait (1924) believed the former were derived from the north flank of the mountains by eastward-moving glaciers from New Brunswick in advance of the Acadian lobe incursion from the north. Others contend that the last ice to invade the lowlands was from New Brunswick rather than from the north. The occurrence of Cobequid igneous stones in the red till of the Carboniferous Lowlands required some northward transport; this might have resulted from stream deposition prior to emplacement of the last red till sheet

but later northward glacier movements are more probable as such stream transport is not effective with the present relief. Late active glaciers on the eastern Cobequid Mountains and Antigonish Highlands are indicated also by the occurrences of eskers and associated outwash, with north-dipping beds, as along River John near Scotsburn and Maryvale. In Barney's River area, at the east end of Merigomish Island, the outwash is graded to, or below, present sea level. Furthermore along the coast near Malignant Cove kame terraces, with northeast-dipping crossbeds, occur up to elevations of about 100 feet; these formed while the same late glacier occupied Northumberland Strait.

The western end of Cobequid Mountains does not appear to have harboured a late ice cap. In this region glacier retreat was northward and meltwater poured through gaps in the western Cobequid Mountains to deposit extensive valley trains with kettled deltas in Minas Basin. Small elongate moraine ridges on the north-central part of the Cobequids may be end moraines. The northeasterly trend of Joggins moraine (Wickenden, 1941), south and east of Amherst, has been assumed to indicate glacial retreat towards the northwest. However, both northwest and southeast of this moraine, fluting and striae trend south-southwest and the ice flow was undoubtedly southward; this ice-flow direction is the reverse of that farther east and was most likely formed at a somewhat earlier date as a calving bay developed in Bay of Fundy.

Marine events. Postglacial changes of level in Nova Scotia are complicated. The Fundy embayment is the only area affording evidence of former emergence but the whole province is now involved in submergence. In southwest Nova Scotia the isobases of differential uplift trend northeastward, roughly parallel to the Fundy shore, with the zero isobase at Yarmouth; uplift is 120 feet at Digby and 150 feet on Long Island, southwest of Digby. The isobase trend changes abruptly in Minas Basin. This may result from changes in amount and rate of uplift due to early opening of Bay of Fundy as a calving bay in the ice front, to presence of late ice north of Minas Basin, and possibly also to the structural influence of Cobequid fault. For these reasons, projection of the Fundy isobase data northward from Minas Basin may not be valid. Borns and Swift (1966) report glaciofluvial deposits overlying glaciomarine deltaic deposits; the top surface of the former slopes from approximate 140 feet elevation at Advocate Harbour to 60 feet near Truro whereas the surface of the lower unit slopes from elevation 130 feet (110 feet above high tide level) in the west to mean high tide at Five Islands, about halfway between these places. The writer, however, believes the upper unit is basically glaciomarine and that the relict marine surface slopes from 120 feet west of Parrsboro to 20 feet at east end of Cobequid Bay. He believes that 'late' ice prevented marine waters from entering the Truro lowland, and also prevented development of higher-level strandlines near Advocate Harbour. The tidal range in Minas Basin is today about 40 to 55 feet but it was undoubtedly different in the past; for this reason the marine limits are referred here to high tide.

There is no evidence of marine overlap along most of the Northumberland shore. A sea bench on the northeast-trending coast east of Arisaig records a former sea-level stand some 5 to 10 feet above the present, and comparable sea-level stands have been recorded elsewhere along this and the George Bay coast; all these are believed to be pre-Wisconsin. In the extreme northwestern part of the province, however, the zero isobase of postglacial uplift is believed to pass through Northport, with maximum marine overlap increasing westward to perhaps 50 feet at the provincial boundary. The evidence is, however, rather inconclusive.

Prince Edward Island

Evidence of glaciation in the western part of Prince Edward Island has long been recognized (Chalmers, 1895) by an abundance of igneous and other stones in the drift. These were derived from westerly or northwesterly sources, whereas the bedrock is red sandstone and shale, mainly of Permian age. The general absence of foreign stones in the drift of the central and eastern parts of the island led early workers to regard the drift mantle as a regolith; but undoubted glacial till is present in all parts of the island (Prest, 1962; Frankel, 1966). It has also been assumed that the last major glacier to override the whole island was from the north—the Acadian Bay lobe of Goldthwait (1924). The distribution of erratics along the north shore lends some support to this concept, but south-trending striae are generally lacking whereas east-trending striae occur along or near the coast. A few north-south trending striae have been recorded from central parts of the south coast but the sense of movement is unknown. These striae predate the last major glacier movements along Northumberland Strait. The general southeasterly trend of ice-flow features across most of the island, coupled with the prevalence of foreign stones in the western end, suggests glaciation from New Brunswick. The island was deglaciated, however, over a period of time during which significant lobing of the ice fronts occurred. On the basis of miniature crag-and-tail features on the south shore west of Borden it is known that there was late westward flow along Northumberland Strait. But numerous glacial striae along the south shore between Borden and Hillsborough Bay, however, trend northeast-southwest rather than along shore, and similar trends occur in the interior on the southwest side of the central higher parts of the island. This suggests that late ice flowed southwestward from the interior towards the Straits and thence westward beyond Borden. Also eastern and north–central parts of the island are characterized by an abundance of glaciofluvial deposits and an anastomosing system of eskers, whereas elsewhere these features are uncommon; this situation is thought to reflect the final decay of remnant ice on central and eastern Prince Edward Island.

Marine overlap in western Prince Edward Island reaches a maximum of 75 or 80 feet along the west coast. The marine limit in the northwestern part of Malpeque Bay is about 30 feet and in the southeastern corner about 10 feet. The zero isobase appears to be near Borden and probably trends southward across Northumberland Strait to Cape Tormentine, New Brunswick. There is no evidence of marine overlap in central or eastern Prince Edward Island. Shells from northwestern Prince Edward Island, believed related to a sea-level stand about 50 feet above the present, were dated at 12,410±170 years and 12,670±340 years (GSC-101, 160).

Magdalen Islands, Quebec

The Magdalen Islands in the central part of the Gulf of St. Lawrence are commonly regarded as lying squarely in the path of Laurentide Ice Sheet as it spread southward or southeastward towards the Atlantic. There is, however, no evidence of Wisconsin Glaciation above the limit of postglacial marine overlap, and below this limit there is only a till-like material that was deposited under water.

At Amherst Harbour, Goldthwait (1915) observed a 15-foot sandy deposit, resembling glacial till and containing only local rock materials overlying deeply decayed buff weathering, grey sandstone. As some of the stones had striations parallel to their long axes he concluded that a glacial origin was more valid than a marine drift origin. Coleman (1920) also noted the complete absence of glacial features on Amherst Island aside from the sandy drift, but he concluded that the thin margin of the continental ice sheet was afloat at a time of higher sea level.

Alcock (1941) reported boulder clay on Amherst and Entry Islands, ground moraine on Grindstone Island, an end moraine on Coffin Island, and large erratics at elevations over 200 feet on Grosse Isle. He saw no evidence of marine uplift, thus disagreeing with Chalmer's interpretation of beaches and terraces to about 115 feet and Coleman's marine overlap to about 200 feet. Prest (1957) considered the Coffin Island end moraine to be a kettled deposit of ice-contact stratified drift, or kame moraine, with the beds dipping inland to the west, and that its bouldery surface was the result of ice rafting during a period of marine overlap. The boulders are mainly foreign to the islands and are believed derived from a northern source. He confirmed the presence of marine gravels on Grindstone Island to a maximum elevation of 120 feet. Alcock considered the 'till' on Amherst Island to be a deposit let down gently from floating ice, an opinion shared by the writer in view of the contact relations between drift and bedrock.

The exact reason for the presence of striated local stones in the marine drift mantle of Amherst Island remains obscure; these stones were probably derived by glacier action some distance beyond the present shores and later ice rafted into their present position. Marine or lacustrine submergence of at least 120 feet elevation is indicative of ice near at hand; this is also indicated by the ice-contact stratified drift on Coffin Island. Glacier ice evidently reached the islands from the north, but only shelf ice reached the southern shores.

New Brunswick

Glacial events. The pattern of ice-flow features in New Brunswick clearly reveals a south to southeast trend in the western and southern parts, and an east to northeast (and/or southwest) trend in eastern parts. Granitic rocks from the Precambrian Shield occur in the Saint John River valley, on the adjoining Chaleur Uplands (Lee, 1955), and in an end moraine along the northern flank of the Chaleur Uplands. It is thus clear that Laurentide ice did reach the upland as well as flow strongly down the Saint John River valley, and probably also down the Matapedia Valley in Gaspé into Chaleur Bay, but elsewhere glaciation was relatively light. Alcock (1948) found evidence of strong glacial flow from the northern end of New Brunswick Highland northeast towards Chaleur Bay and east towards Gulf of St. Lawrence, and concluded that such movements must have preceded as well as followed the arrival of Laurentide ice in New Brunswick. He assumed that Laurentide ice over-topped the highlands on the basis of south-trending striae in southern New Brunswick and particularly in the Moncton area, but striae trends in this area are now known to be diverse. It is more likely, judging by ice-flow patterns and an end moraine (Map 1253A), that ice flowed mainly around the highlands rather than over them. Flint (1951) suggested that the ice responsible for the northeast- and east-trending striae was of local derivation and subsequent to the main Laurentide glaciation. South of Bathurst, the ice-flow features visible on airphotos, suggest that some late ice flow was out of Chaleur Bay and towards the south-southwest but this has not been confirmed. Farther south, between Newcastle and Moncton, ice-flow features and an east-west end moraine also suggest a southerly ice movement as if from Miramichi Bay. On the other hand, but a short distance southeast of the ice-flow features and closer to Moncton, miniature crag-and-tail features observed on a pebbly sandstone indicate ice flow to the northeast parallel to the regional drainage.

On both the mainland and islands in the extreme southwest corner of New Brunswick, Alcock noted *roches moutonnées* and striations trending southeast to east, in contrast to the general south-southeast trend typical of the western part of the province. He concluded that there had been an ice movement from the mountains of Maine, presumably prior to the main Laurentide glaciation.

Marine events. Chalmers (1890) noted marine terraces along the Fundy shore both southwest and northeast of Saint John. In the former area the deposits had a maximum altitude of 225 feet and contained shells; in the latter area the highest terrace deposits were only 125

feet and no shells were seen. In Saint John River valley at Fredericton, Lee (1959) found estuarine deposits at 125 feet. It is possible that the lower part of Saint John River valley and the New Brunswick side of Bay of Fundy were covered by late ice while the high-level terraces were being formed southwest of Saint John. Shells from the seacliff 5 miles west of Saint John are dated at 13,325 ± 500 years (I-GSC-7).

Along the east coast of New Brunswick large tracts of low-lying ground formerly covered by the sea show little or no evidence of marine action; there were apparently no significant halts as the land rose from the sea and, in general, marine sediments appear to have been removed by erosion or incorporated into the soil profile. Local areas of marine sand, gravel, and poorly washed sediments, and the disposition of ice-rafted boulders do provide some evidence of former marine action. The marine limit varies from zero at the east end of Cape Tormentine to 100 feet at Moncton. It is 150 feet about 15 miles west of Richibucto and 225 feet at Newcastle and Bathurst. Farther west along Chaleur Bay the marine limit is not evident, probably as a result of late ice in Chaleur Bay during the period of maximum overlap along the east coast. The isobases of differential uplift trend northeast along the Bay of Fundy shore, northward through Moncton area, and northeastward again along the east shore of the province; this gives them an open S-shape. The effects of late ice and the possible influence of Cobequid fault in the amount and rate of uplift on either side of Bay of Fundy, together with the lack of age determinations, preclude correlation across the bay. The uplift in western Prince Edward Island, however, appears in accord with data from eastern New Brunswick.

Gaspé Peninsula, Quebec

Glacial events. Much has been written concerning glaciation of Gaspé Peninsula; this is succinctly summarized by McGerrigle (1952). Cirque and local ice-cap development preceded Laurentide glaciation of Gaspé Peninsula. Shield erratics have been found on the highest parts of the peninsula. There is no proof, however, that the erratics were emplaced during the last major glaciation. McGerrigle notes the general sparseness of erratics in eastern Gaspé, and the seeming absence of anorthosite in a 50-mile-wide belt along the central part of the peninsula. It is noteworthy that neither McGerrigle nor Brummer (1958) reports Canadian Shield stones in the Béland–Upper York River highland area though they occur in lower areas farther east. It is perhaps noteworthy that granite gneiss erratics in southeastern Gaspé are only recorded along or close to the shore highway and their mode of emplacement is uncertain. The Matapedia Valley was the main ice-flow route into Chaleur Bay.

A highland ice-cap phase followed the maximum phase of Laurentide glaciation and, as the ice receded from the north side of Gaspé, this ice cap remained active and transported boulders northward as far as St. Lawrence River (McGerrigle, 1952). On the eastern end of the Shickshock Mountains, erratics derived from lower areas some miles south of the mountains indicate that the highland ice cap once had a centre of outflow lying south of the highlands proper. As the ice cap waned it broke up into minor ice caps that gave rise to radial flow as from the Tabletops and Béland–Upper York River highlands. The local ice-cap phase of glaciation finally gave way to cirque and valley glaciation.

Marine events. The extent of marine overlap along the southern and eastern coasts of Gaspé is not precisely known. Marine sediments are reported at 180 feet on the north side of Chaleur Bay and at 224 feet on the eastern end of the peninsula. The writer observed gravel deposits up to a maximum of 180 feet near Grand Cascapedia, but regarded these as outwash deposits graded to a wave-base perhaps as low as 120 feet above the present sea level. The steep outcrops of the eastern end of the peninsula do not lend themselves to the preservation of marine strandlines. A marine terrace at Prével has an elevation of only 90 feet and at Cap-des-Rosiers Est the upper limit of near-shore sediments is at 75 feet. At present there is evidence of coastal 'drowning' at the head end of Gaspé Bay. Along the western part of the north shore marine deposits have been recorded up to 300 feet, but 4 miles east of Mont Louis the writer did not see any evidence of marine overlap above the surface of a small delta at an altitude of 95 feet.

Appalachian Mountains of Quebec

Flint (1951) has summarized the voluminous literature pertaining to an Appalachian centre of glacial outflow (that either preceded or followed glaciation) from the highlands of Maine, New Hampshire, Vermont, and adjoining Quebec. In the last area the problem concerns the evidence of northward flow towards the St. Lawrence from Notre Dame and Green Mountains. Flint concluded that ice has flowed northward, at least locally. Recent studies by Gadd (1966), however, do not support the concept of glacial transport to the north. It may be that glaciation of the western part of the Appalachian Highlands, south of Quebec City, was unlike that in the east where late northward flow is evident. Lee (1962) reported striae and crag-and-tail indicative of northward ice flow in the Rivière du Loup–Trois Pistoles area. These were formed when a calving basin developed in St. Lawrence Valley while ice remained over the drainage divide in Notre Dame Mountains.

Labrador Sector, Laurentide Ice Sheet

Little is known about the initial development of the Labrador sector of the Laurentide Ice Sheet or the course of glacier movements that led to the ice cover of the entire region between the Labrador coast and western

Manitoba, and possibly beyond. It is probable that in response to a change in climate, the Lake Plateau of central Quebec–Labrador developed an ice cap by "instantaneous glacierization" (Ives, 1957), followed by radial outflow. The early influence of an open Hudson Bay in the development and differential expansion of the ice cap towards the west was favoured by Hare (1951) and Derbyshire (1962) but not, on climatological grounds, by Barry (1960). The prevailing southwesterly winds from the mid-continent no doubt contributed to southwestward expansion of the ice sheet towards Ontario and the Great Lakes region. There were probably local centres of excessive snow accumulation in the marginal zone of the expanding ice sheet; these may account for glacier lobations and resulting striae which in many areas diverge markedly from the general, deglacial, trend.

Early in Classical Wisconsin time the glaciers flowed off the Laurentian Highlands into St. Lawrence Valley and, upon filling the valley, sought escape northeastward towards the Gulf of St. Lawrence and southwestward towards the Ontario basin. Gadd (1966) reports striae indicative of southwesterly ice movements from the north flank of the western end of Notre Dame Mountains south of Quebec City, which he believes were formed during an overriding phase of glaciation. Strong eastward ice flow was long-continued down Saguenay River and thence northeast down St. Lawrence River as indicated by ice-flow features on the land and a very deep channel in St. Lawrence River beginning at the mouth of the Saguenay. The channel extends east-southeastward across Gulf of St. Lawrence through Cabot Strait and thence to the edge of the continental shelf, and is known as Laurentian Channel. Ice-flow features give some indication that the ice-flow divide in St. Lawrence Valley may have migrated westward from well east of Saguenay River to somewhat west of the Saguenay during Wisconsin time. Gadd suggests also that during the Wisconsin there was a shift of the centre of outflow in the Laurentian Highlands towards the west and that by the time the glacier was waning in St. Lawrence Valley the ice-flow direction was southeastward across the valley. He reports evidence of strong ice flow south–southeastward up the Chaudière River valley at this later stage.

The Laurentide ice upon filling St. Lawrence Valley and flowing along it sought escape southward through cols in the eastern part of Notre Dame Mountains and hence combined with elements of the Appalachian Glacier Complex; ultimately, as earlier mentioned, it may have overtopped the highest mountains in Gaspé, but its southward extent was perhaps more limited than has been generally recognized.

Study of the mineralogy of tills of southwestern Ontario led Dreimanis, *et al.* (1957) to conclude that glacier movements into Ontario and Erie basins were first from the north-northeast and only later from the east along the axis of the lake basin. The seeming reversal of the order of major periods of ice flow between Erie basin and Trois-Rivières region may be a result of their wide geographic separation and also to the different parts of Wisconsin time being considered. The occurrence of cobbles and boulders of Gowganda conglomerate in southwestern Ohio also indicates strong southerly movements from the Huron basin prior to the incursion of the Erie ice lobe. Also, in northwestern Ontario the writer found pebbles and cobbles of oölitic jasper, derived from the Sutton Mountains–Nowashe Lake belt west of James Bay, or from the Belcher Islands on the east side of Hudson Bay, and of a characteristic greywacke reportedly derived from Cape Jones at northeast end of James Bay, perhaps as much as 600 miles from their outcrop area. These indicate an early southwesterly ice-flow direction that is somewhat divergent from the deglacial flow pattern.

Information on the retreat of the Labrador sector of Laurentide Ice Sheet is naturally far more abundant than on the advance or build-up stages. The present surface bears a record of the last glacial events, modified only in part by postglacial changes, whether erosional or depositional. After the Wisconsin glacial maximum it is probable that there was thinning of the ice sheet over a broad marginal zone as the ice front retreated. The following discussion attempts to follow the major changes and events as the ice front receded from various parts of the country.

St. Lawrence Lowlands

In southern Quebec the margin of active ice receded northward from Notre Dame Mountains into St. Lawrence Valley, but flow probably continued both up and down the valley from near the mouth of Saguenay River for some time before giving place to a general southward flow across the valley. During this latter stage the ice advance was restricted by the north side of Notre Dame Mountains and a system of end moraines was deposited, the lowest of which is known as the Highland Front moraine (Gadd, 1964, 1966). The whole system represents a lowering of the surface of about 1,000 feet. The St. Antonin Moraine (Lee, 1962) is the easternmost end moraine of the Highland Front system, and near Rivière du Loup the glacier that produced it was moving down valley and calving into the sea which occupied the lower part of the valley.

Farther west, in the wide part of St. Lawrence Valley, meltwater was ponded between the Appalachian Highlands, including Adirondack Mountains, and the ice front to form glacial Lake Vermont, with discharge south down Hudson Valley. The Drummondville Moraine (Gadd, 1960) was probably built during the Fort Ann phase of this lake. As the ice front receded northward the lake expanded northeastward along the south side of St. Lawrence Valley until it was able to discharge into the sea near Quebec City. The last short-lived lake is presumed herein to represent the Trenton phase of the post-Iroquois lakes as recognized in the Ontario basin by E. Mirynech, and is considered to be the last phase of con-

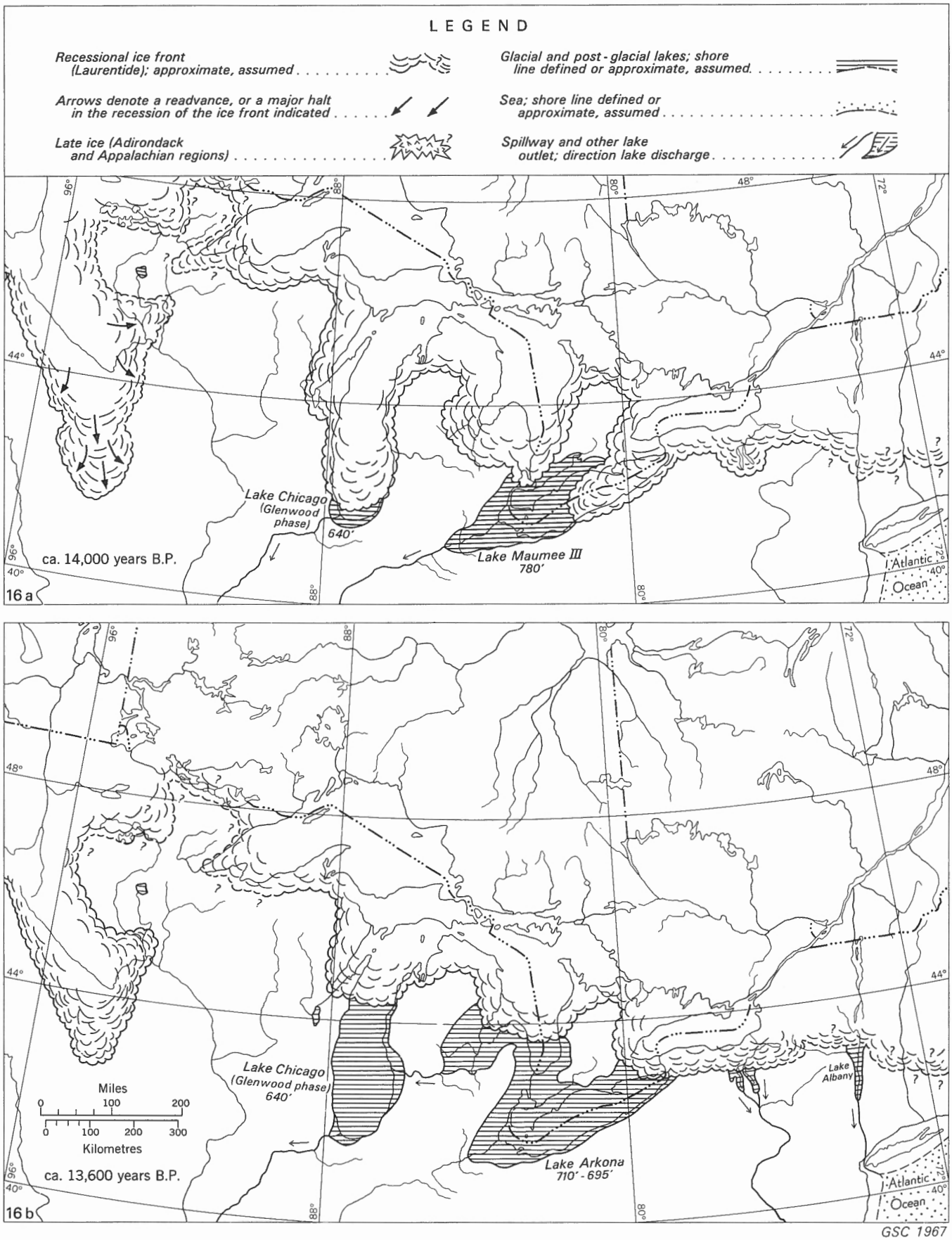


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada.

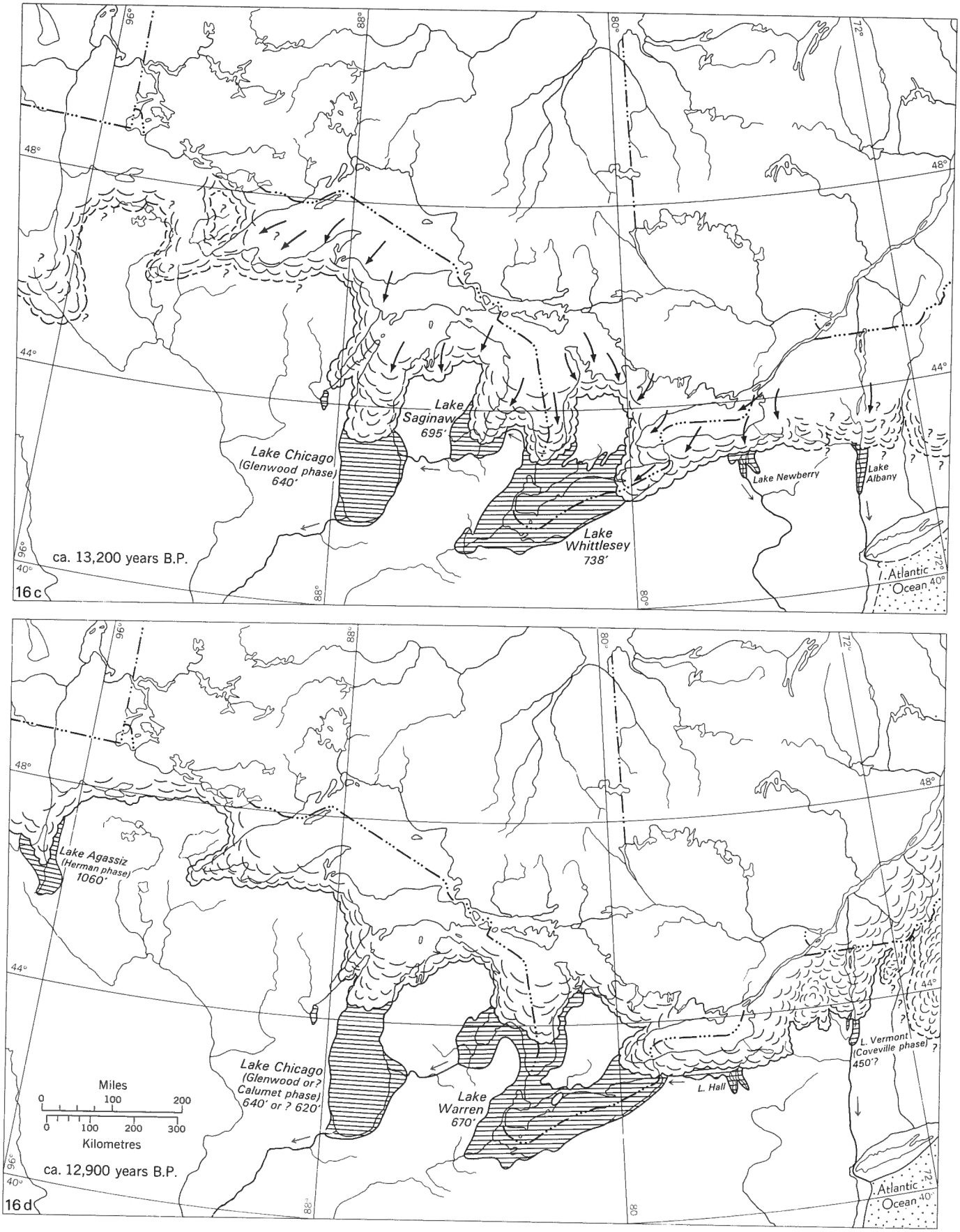


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)

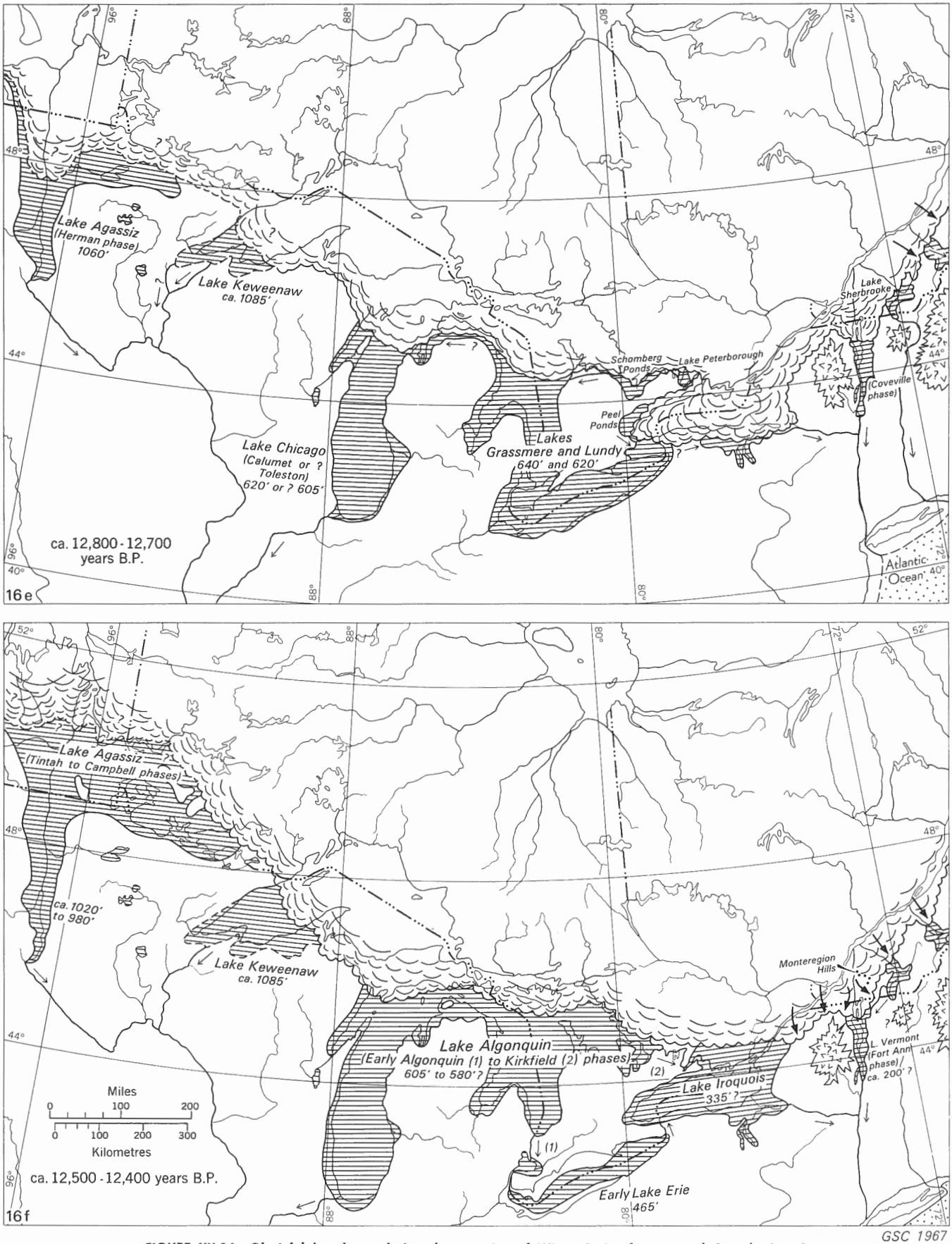


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)

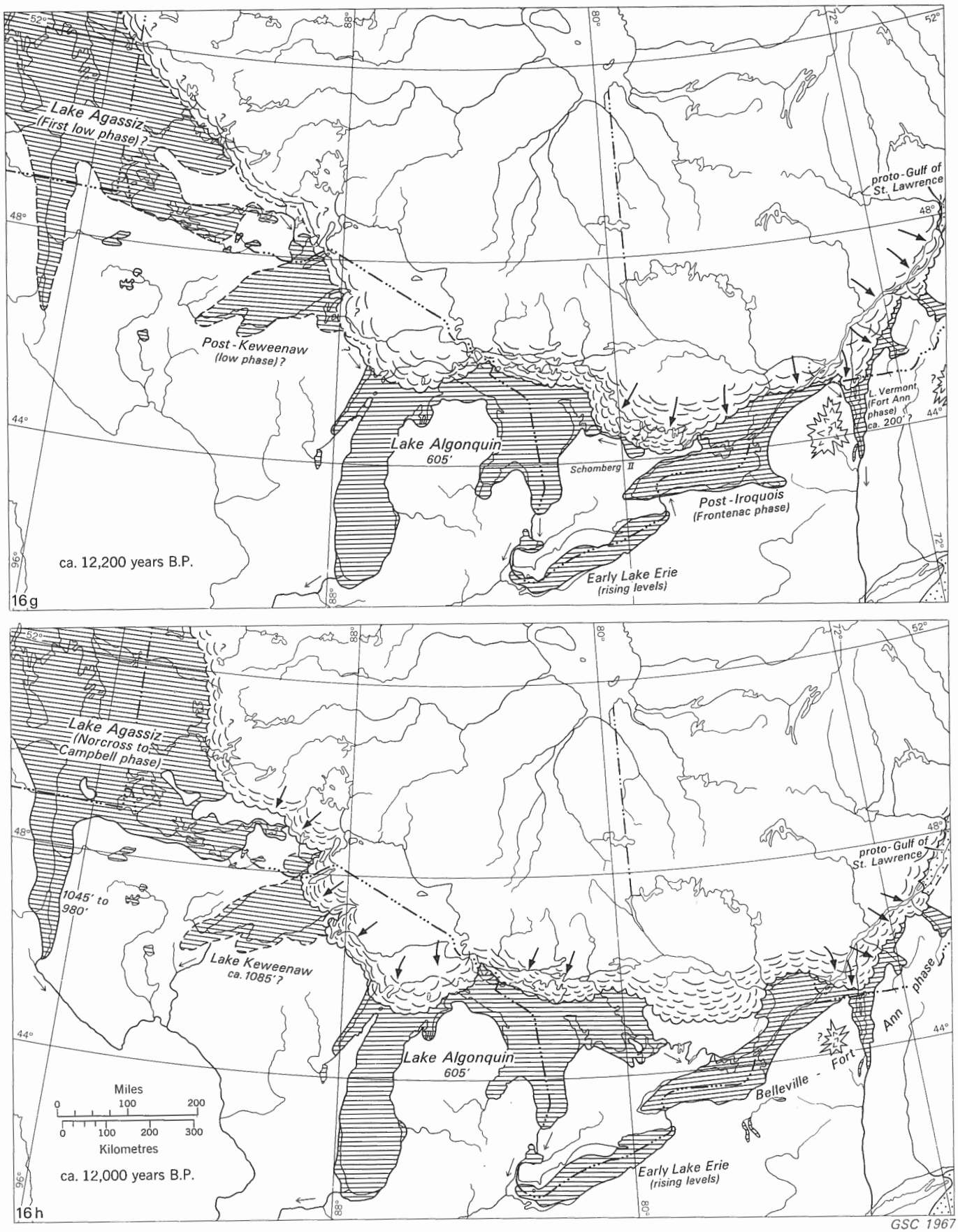


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)

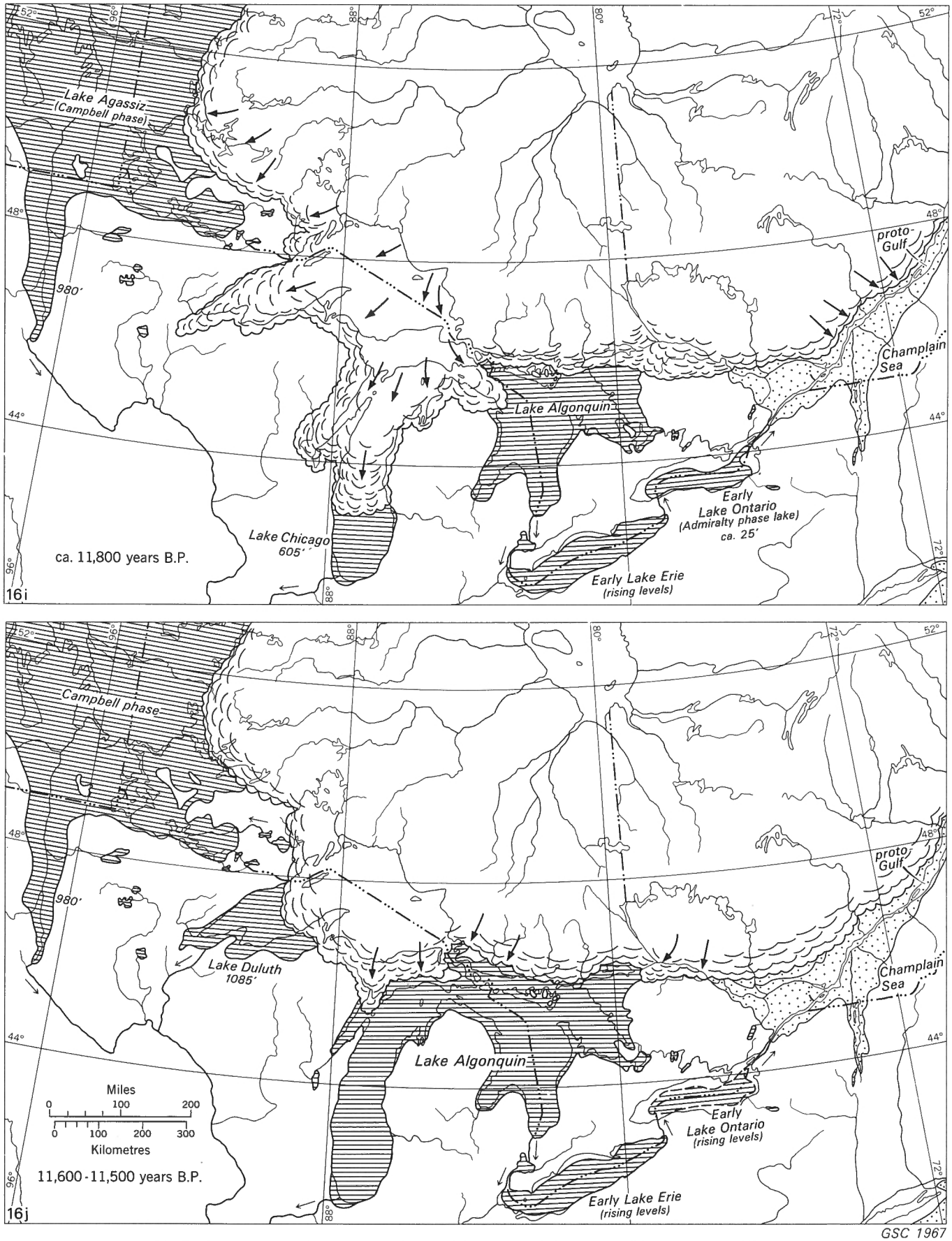


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)

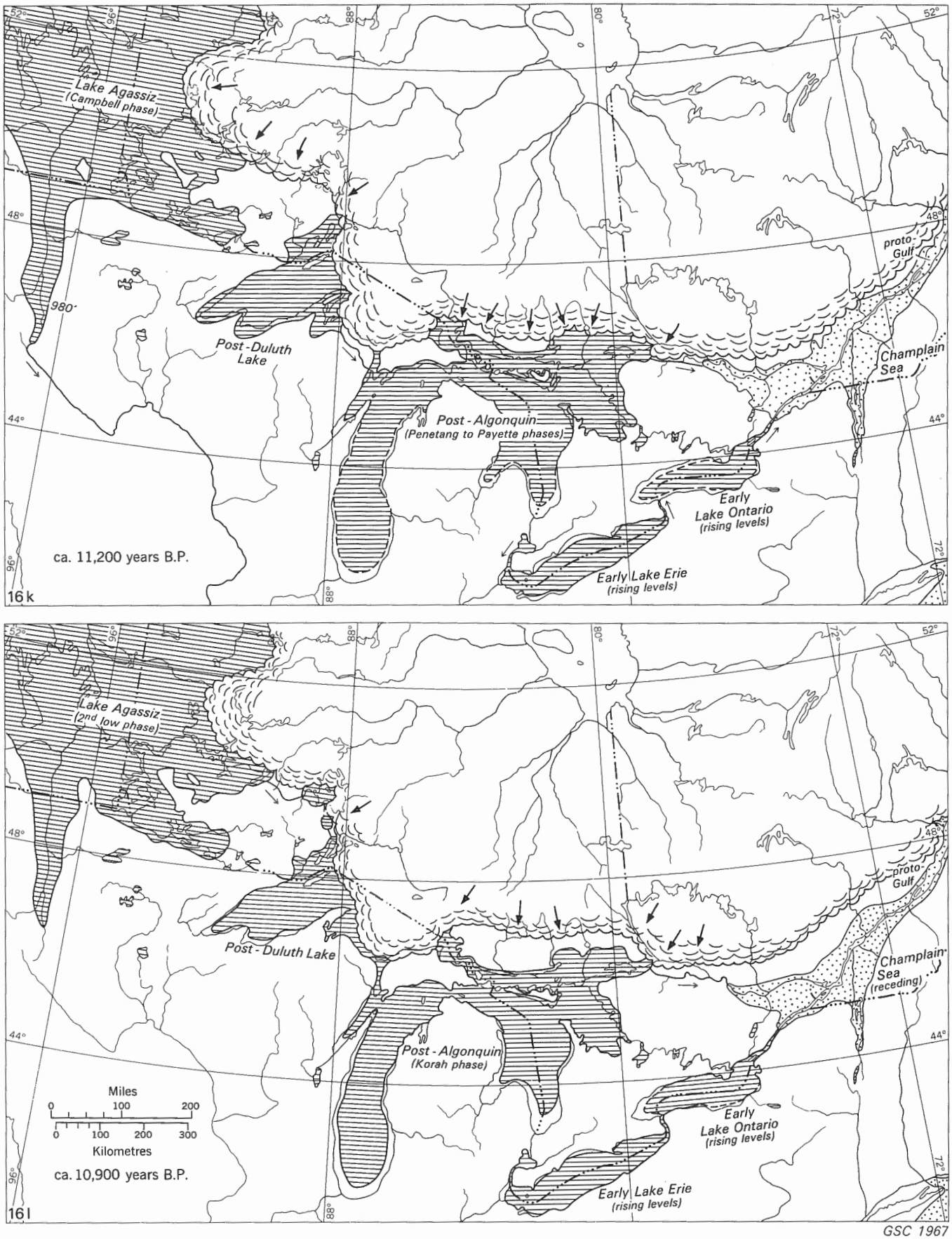


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)

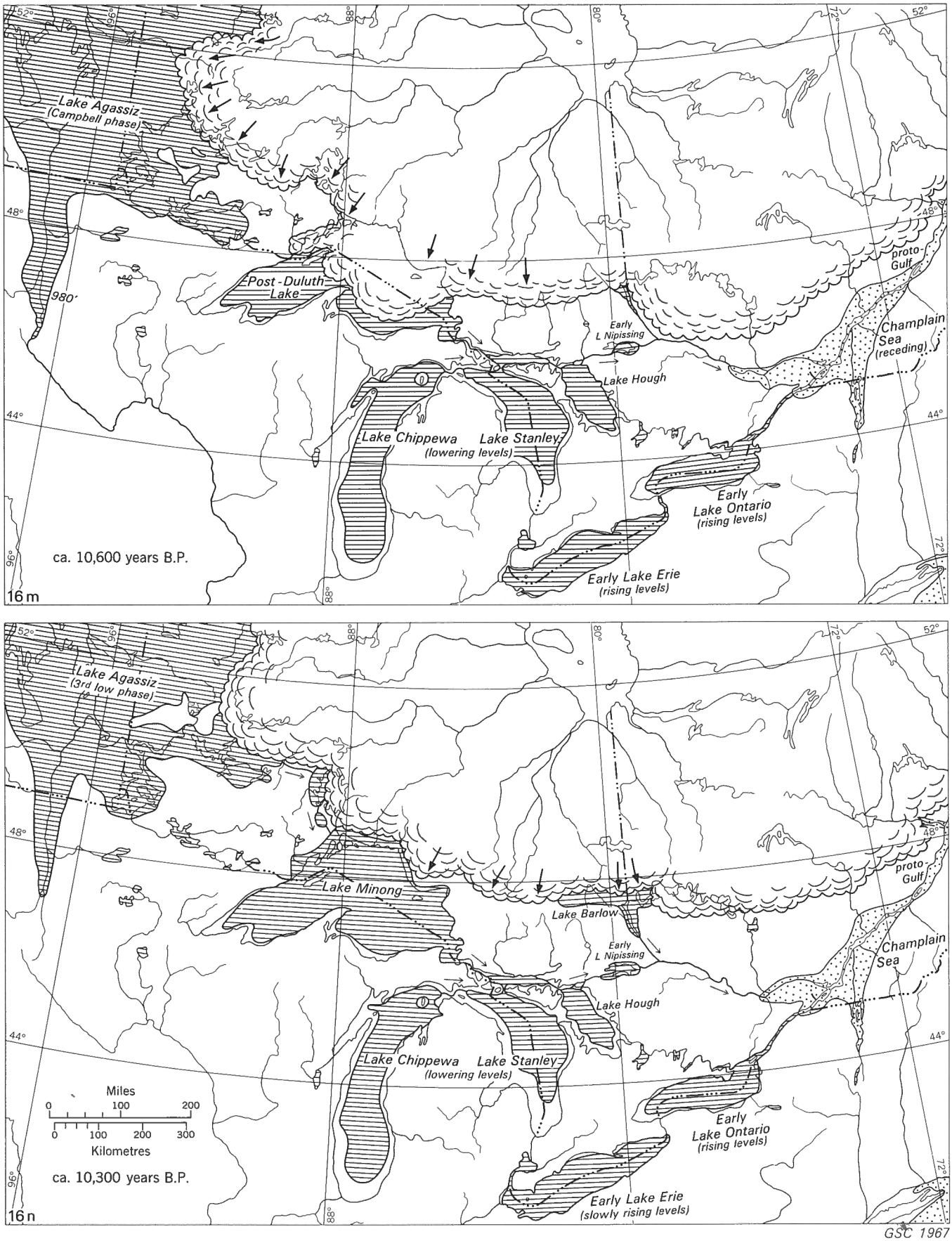


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)

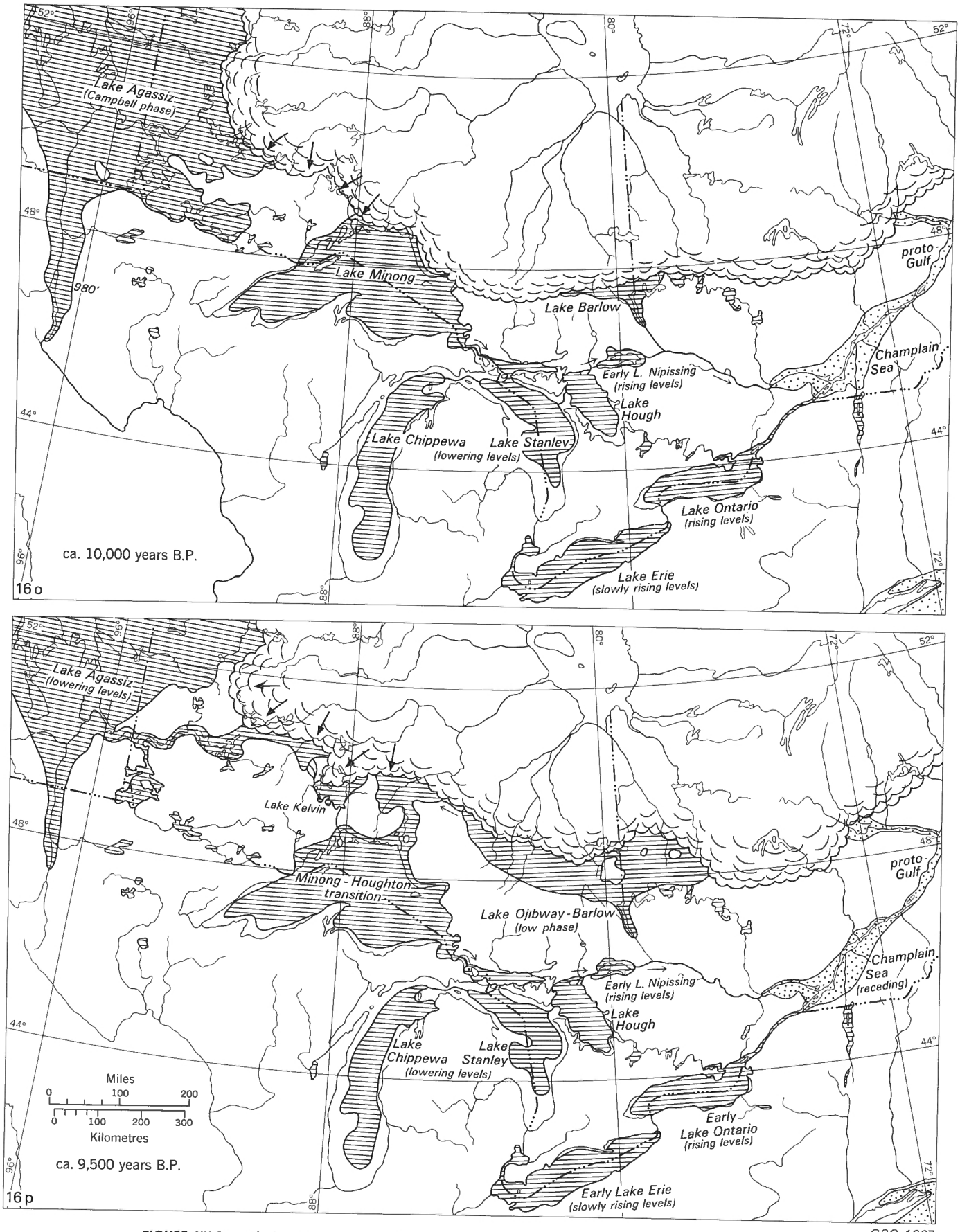


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)

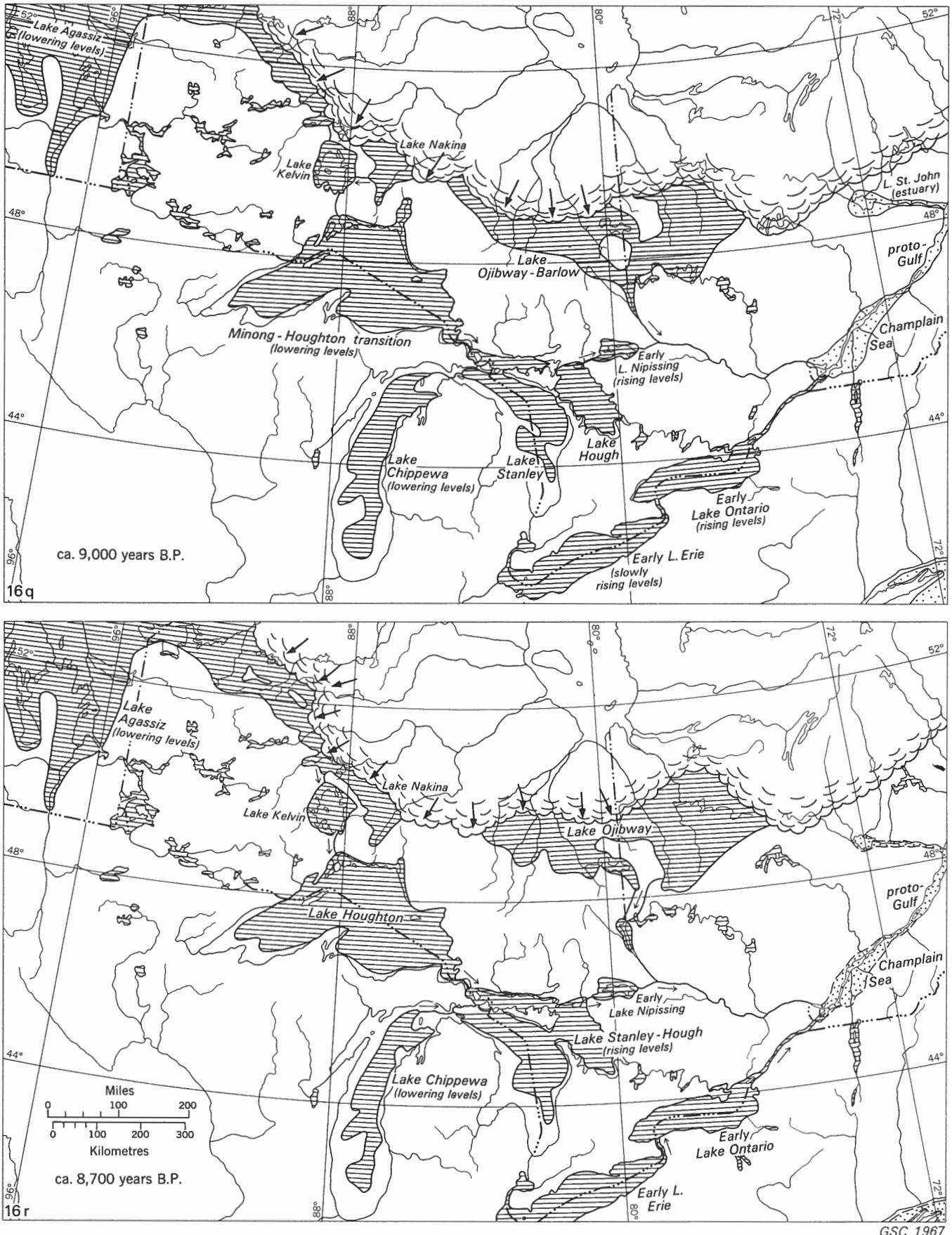


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)

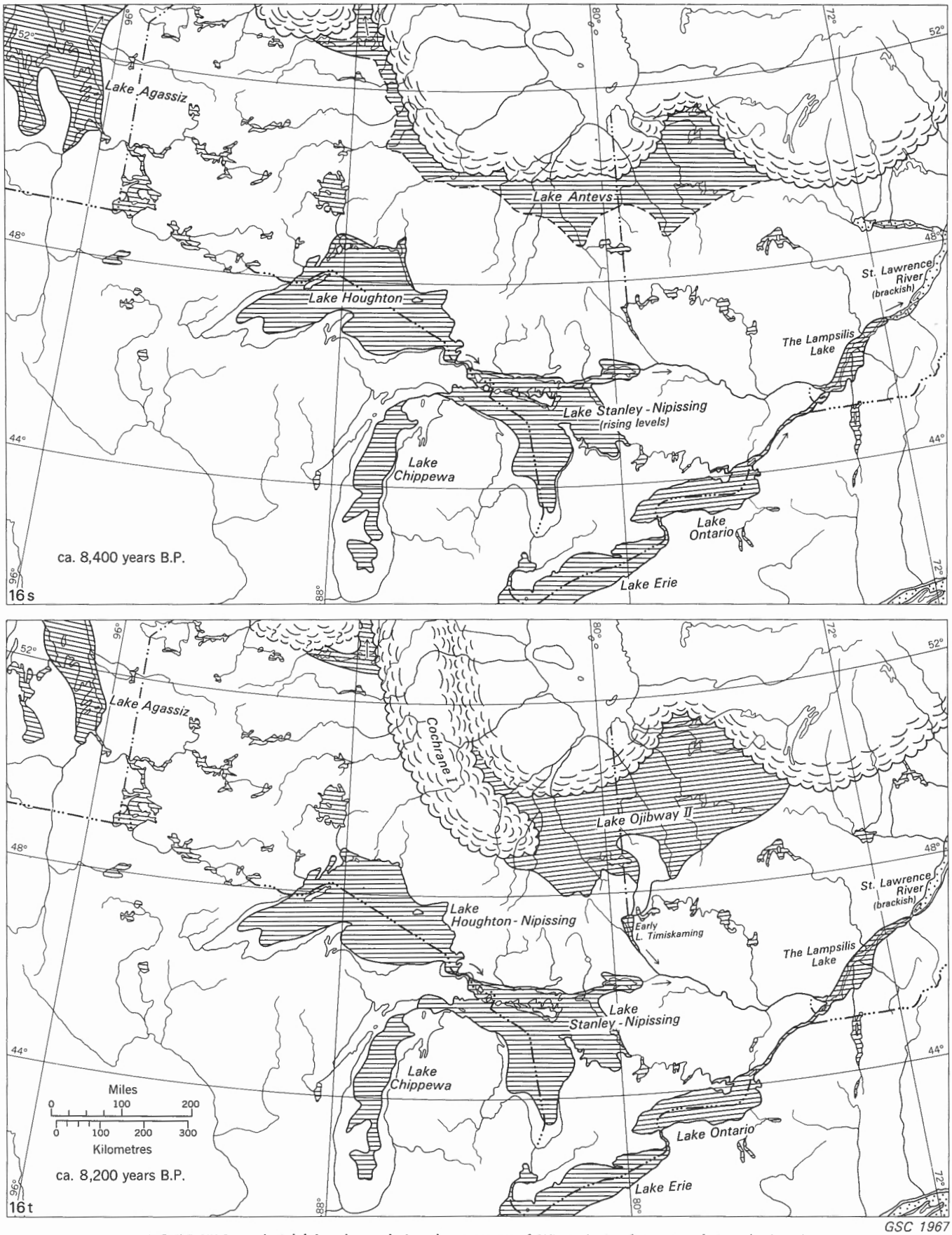


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)

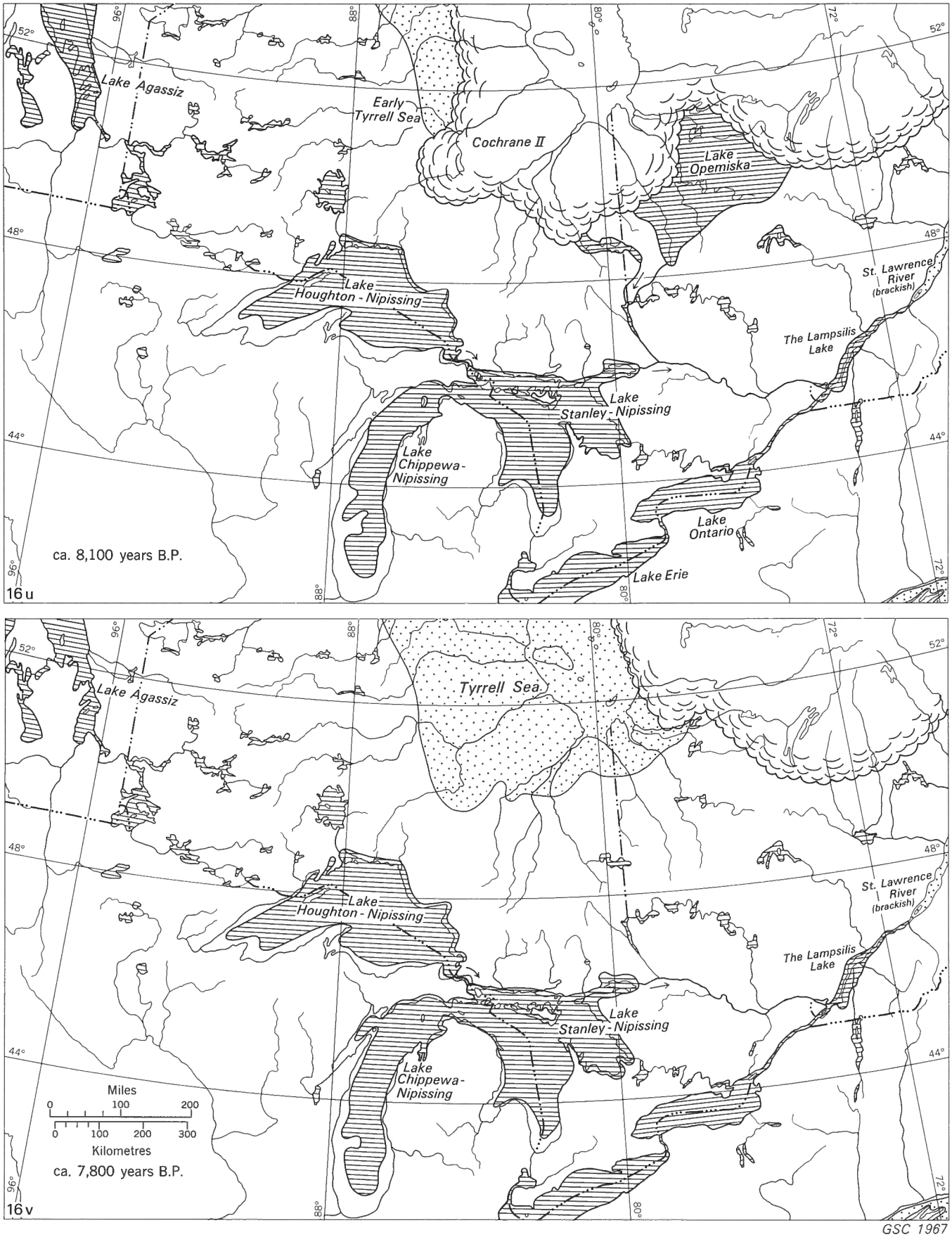


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)

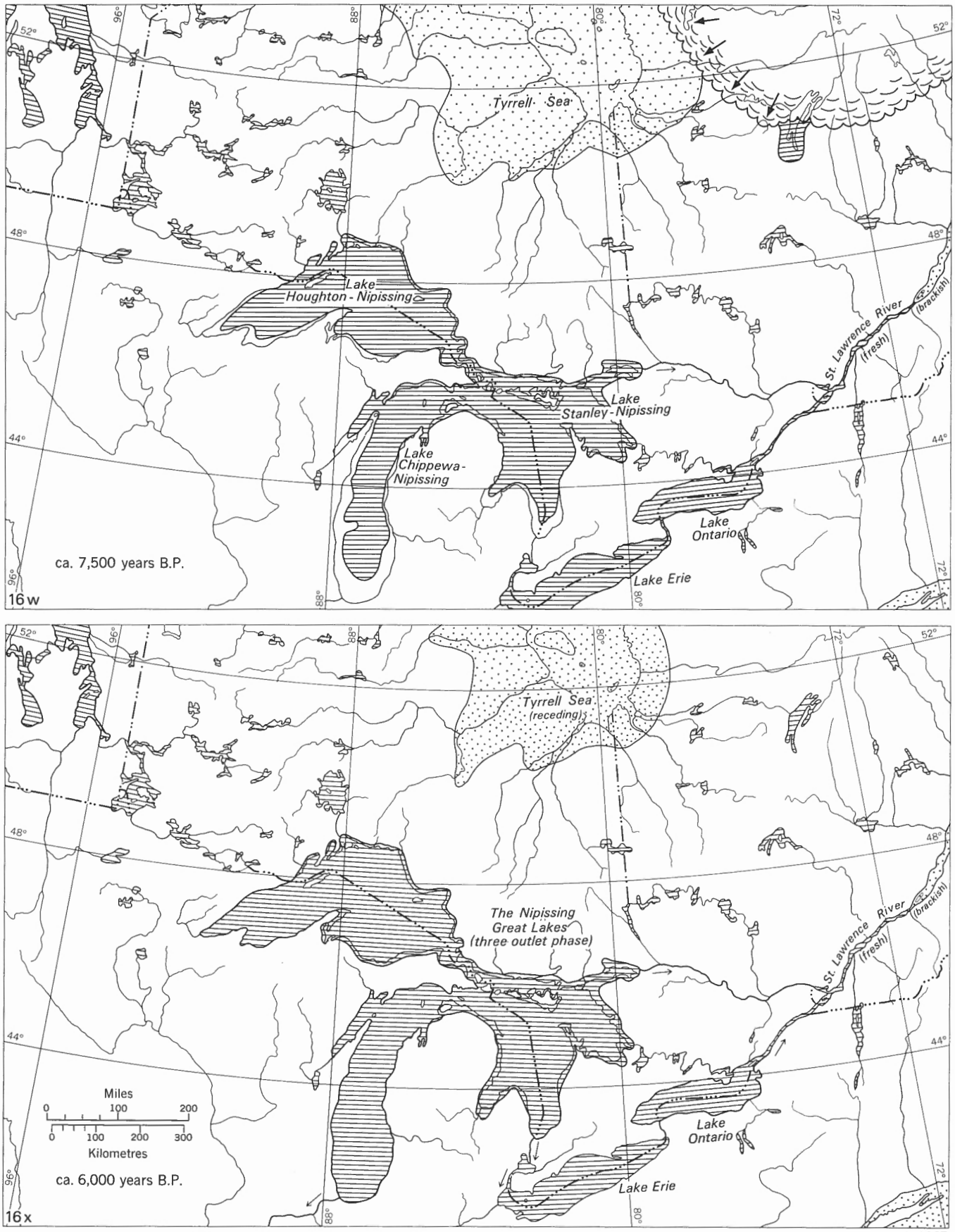


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (conc.)

fluent Iroquois–Vermont waters. According to N. R. Gadd an ice lobe occupied Chaudière River valley and probably formed the last barrier between the glacial lake and the sea (Fig. XII-16h). Recession of the ice front from Chaudière Valley allowed the lake to drain to sea level whereupon the sea encroached the upper St. Lawrence and Ottawa River valleys. This inland sea in St. Lawrence Valley above Quebec City constitutes the Champlain Sea (Gadd, 1964). Labrador ice remained active and built an end moraine, presumably in the sea, parallel to the valley in the vicinity of St. Narcisse (Karrow, 1959). This moraine may be traced westward some 100 miles to near Lachute, Quebec (Parry and MacPherson, 1964). As the ice front receded from St. Lawrence Valley the Champlain Sea attained its maximum northward extent. Strandlines occur up to 750 feet above sea level north of Trois-Rivières and Montreal. Thereafter differential uplift brought about regression of Champlain Sea and development of the modern river system (Figs. XII-16k-w).

In the Great Lakes region, thinning of the Labrador sector of Laurentide Ice Sheet and retreat of the ice margin from the farthest point of advance, some 150 miles south of Lake Michigan, began about 17,000 years ago. The ice front retreated into the Lake Erie basin about 14,500 years B.P. Meltwaters were ponded between the basin rim and the receding ice front to form the first of the glacial Great Lakes, Lake Maumee, with discharge southward into Mississippi River (Hough, 1958, 1963, 1966). Thereafter, fluctuations of the ice front and differential uplift of the land, consequent upon removal of the ice load, combined to produce a complex system of lakes and spillways throughout the Great Lakes region and extending into the upper St. Lawrence Valley.

The present indicated history of the Great Lakes is the result of studies by geologists, geographers, and others that date back well before the turn of the century. The early work established the fundamental concepts pertaining to glacial lakes and the framework upon which later work was based. Studies made during the last decade, supported by radiocarbon datings, have made necessary some major changes in the basic chronology and sequence of events. The most recent attempts to correlate events between the various lake basins and discharge routes are by Hough (1963, 1966), Wayne and Zumberge (1965), and Chapman (1966). Two contrasting correlation charts are, in fact, introduced by Wayne and Zumberge to indicate differing concepts held by various authors. On the Canadian side of the Great Lakes available data and radiocarbon dates of lake deposits remain in conflict with the generally accepted correlations.

The writer has attempted to harmonize viewpoints, or otherwise point out discrepancies, concerning deglacial events in the well-documented history of the central Great Lakes and upper St. Lawrence regions with events in southern Manitoba and in James Bay region. The account invokes non-uniform fluctuations of several major ice lobations—as a result of variably delayed responses to climatic changes—to overcome some problems raised by radiocarbon dating of lake and sea deposits in widely separated regions (Figs. XII-16 a-x). It is based also in part on regional ice-flow trendlines and morainal positions developed during deglaciation (Map 1253A).

The southern glacial Great Lakes. The first of the glacial Great Lakes formed in the western end of Erie basin and is known as glacial Lake Maumee (Fig. XII-16a). It expanded into southern Huron basin as the Huron ice lobe receded. Successive water levels, resulting from use of two different outlets as the ice margin fluctuated, were at 800, 760, and 780 feet a.s.l.[1] During the high phases discharge was to the south via Wabash River, and at the low phase to the west, via Grand River, into glacial Lake Chicago in the Michigan basin and thence south, in both phases, to the Mississippi River system. Further ice retreat resulted in final abandonment of the Wabash River spillway and reopening and modification of the Grand River spillway to form glacial Lake Arkona, with its unwarped strandlines developed at 710, 700, and 695 feet a.s.l. (Fig. XII-16b). Continuing retreat of the ice front resulted in low-level lakes that are recorded by deposits in the United States and locally in Canada. A major glacial advance that built the Port Huron moraine system and a correlative moraine that crosses the base of Long Point in Erie basin gave rise to glacial Lake Whittlesey in western Erie basin, with prominent beach at 738 feet a.s.l. (Fig. XII-16c). Discharge was to the northwest via Ubly River channel into Saginaw basin and thence to Lake Chicago and Mississippi River.

Fluctuations of the ice front resulted in expansion of the lake both in the north and east, although at lower levels because of down-cutting of the outlet across Michigan Peninsula. These lake phases are known as Warren and Wayne and, in the main, they occupied the whole of Erie basin and southern part of Huron basin (Fig. XII-16d). Hough related them to fluctuations of the Port Huron (Mankato) ice. The successive lake levels are reported as 680, 670, 660 (Wayne), and 685 feet a.s.l. Continued recession of both Huron and Ontario ice lobes resulted in still lower lake levels known as Grassmere at 640 feet and Lundy at 620 feet a.s.l. (Fig. XII-16e). Discharge was believed to have been eastward, for the first time, along south side of Ontario basin into the Mohawk and Hudson River systems. Hough has favoured an outlet around the northern end of Michigan Peninsula and thence via glacial Lake Chicago and Illinois River to the Mississippi, pointing to the similarity of elevations of major lake phases in the lake basins concerned. Leverett and Taylor (1915), however, thought that by Lundy time the Huron basin waters drained southward to Erie basin. Further retreat of the Ontario basin ice lobe facilitated discharge eastward; early Lake Erie was established in

[1] a.s.l.—above Atlantic mean sea level. The given elevation of a glacial lake is the present elevation of the unwarped part of the inferred lake basin.

Erie basin and early Lake Algonquin in Huron basin with the controlling sill at Port Huron (Sarnia) at 605 feet a.s.l. (Fig. XII-16f). Early Lake Erie occupied only a small part of the present lake basin due to depression of the outlet at Buffalo (Fort Erie), but differential uplift later raised the outlet area with consequent filling of the Erie basin (Lewis, 1966).

When the ice sheet receded from the southern part of Lake Ontario basin, a lower outlet was uncovered at Rome, New York, and the Lundy Lake was lowered rapidly. Glacial Lake Iroquois was established in the Ontario basin at about 335 feet a.s.l. (Fig. XII-16f). The spillway led eastward through Mohawk River valley to Hudson River and thence south to the Atlantic Ocean. Due to prolonged use of Rome outlet, many excellent shoreline features were developed. The writer believes, however, that the shoreline features at 1,100 feet southwest of Covey Hill and ascribed to main Lake Iroquois are the work of a younger and lower lake phase.

The main Iroquois shoreline in the Ontario basin is now warped upward along a line trending N20°E with the hingeline situated south of the lake basin. The shoreline is about 360 feet at Hamilton, 460 feet at Rome, and 700 feet at Watertown, New York (Coleman, 1937). The indicated uplift gradient between Rome and Watertown is about 6 feet per mile. If this gradient and trend of uplift is projected beyond Watertown, the isobase of differential uplift through Cherabusco, New York, would be about 1,270 feet. An increase in the rate of tilt northward, as probably occurred, would give a higher figure but this would be offset by any swing towards the north in the line of maximum uplift, as does occur in the Adirondack region. Short spillways, at elevations of 1,305 and 1,290 feet, have been reported south of Cherabusco (MacClintock and Terasmae, 1960; MacClintock and Stewart, 1965) and were believed related to local lakes or pondings in front of Fort Covington ice with discharge westward into Lake Iroquois. Though the Iroquois shoreline is reported at only 1,100 feet southwest of Covey Hill there is evidence of shoreline features to about 1,250 feet, 1¼ miles south of Cherabusco. This is considered to be an Iroquois shore rather than that of a local lake. A small embayment of Lake Iroquois extended eastward from Cherabusco as a re-entrant between the receding northern and eastern ice fronts around Adirondack Mountains. This ice, probably Fort Covington, built an end moraine and ribbed moraine complex south and southwest of Covey Hill. C. S. Denny (U.S.G.S.) believes that an outlet opened southeast of Ellenburg, New York, which allowed main Lake Iroquois to breach the end moraine and discharge into glacial Lake Vermont farther south in Lake Champlain valley. In the northeast, strandlines indicate a lowering of about 150 feet during operation of this outlet; the real lake level, however, was lowered perhaps only 75 feet. The new lake level, formerly considered main Lake Iroquois, is herein referred to as the Ellenburg phase of the post-Iroquois lakes.

As glacier recession was resumed, Covey Hill outlet (sill elevation 1,010 feet) was uncovered and the post-Iroquois lake level was lowered a further 75 feet. This lake phase was named glacial Lake Frontenac and is herein considered the Frontenac phase of the post-Iroquois lakes (Fig. XII-16g). The Covey Hill outlet today consists of a 60-foot dry waterfall or cliff, a 75-foot deep lake with an unknown thickness of sediment at its base, and a mile-long gorge about 130 feet deep and 300 to 600 feet wide. When the ice withdrew from the northern and eastern flanks of Covey Hill there was a major drop in lake levels of some 125 feet. The short stand at this level, probably occasioned by an ice-marginal fluctuation on the northern sides of Covey Hill, is termed the Sydney phase lake by E. Mirynech.

In the Trenton embayment in eastern Ontario the lake was lowered a further 30 to 75 feet according to Mirynech, prior to a significant halt responsible for development of the Belleville beach. Isobases drawn on the Belleville beach would place the strandline on Covey Hill at about 750 feet. This is the same as that of the Fort Ann phase of glacial Lake Vermont which expanded northward as the eastern side of Covey Hill was uncovered by the ice (Chapman, 1937). The now confluent water bodies discharged southward along the earlier established route to Hudson River and the sea (Fig. XII-16h).

Rapid ice-marginal retreat in St. Lawrence Valley allowed the Belleville–Fort Ann phase lake to expand northward and also to extend a long arm northeastward between the ice front and the eastern Quebec Uplands which opened an escape route eastward to the sea at Quebec City. Water levels first dropped about 40 or 50 feet and, according to Mirynech, the Trenton shoreline was formed in the eastern part of Lake Ontario basin. This temporary halt in lowering lake levels was probably occasioned by an ice-marginal fluctuation in conjunction with a topographic barrier near Quebec City. As Labrador ice receded from Chaudière River valley the Trenton phase lake drained to sea level; marine waters then interchanged with the remaining lake waters in upper St. Lawrence Valley to form Champlain Sea (Fig. XII-16i). The sea extended up St. Lawrence River to Brockville and, as the ice front receded, up the Ottawa River to Petawawa. The Lake Ontario basin waters were drained to the Admiralty low water phase, probably less than 25 feet above Champlain Sea. These last events occurred shortly after 12,000 years B.P. as Champlain Sea shells have been dated 11,880 ± 180 years (GSC-505).

The indicated maximum drop in lake levels from main Iroquois (presumably 335 feet a.s.l.) to the level of Champlain Sea is more than 335 feet. This is probably due to differential uplift during the later phases of lake history and to the difference in real sea level during early Iroquois and Champlain Sea time as compared to the present. Also, the main Iroquois data may well be in some error.

The Algonquin Lakes. As ice recession continued in the Huron basin, early Lake Algonquin occupied the southern part of the basin and, also, Georgian Bay and Lake Simcoe lowland. Numerous strandlines have been named in various parts of the Huron and Michigan basins applicable to a long 'Algonquin' interval of ice retreat and differential uplift, but it is evident that the lake phases and their implications are not yet adequately known.

Early Lake Algonquin at 605 feet a.s.l. was drained southward, possibly first via Chicago and later via Port Huron, into Lake Erie until such time as Lake Simcoe lowland became free of ice and discharge was possible via the Kirkfield-Fenelon Falls and Trent Valley river system into Lake Iroquois (Fig. XII-16f). Continued use of this outlet system, by the Kirkfield phase lake (at about 580 feet a.s.l.), was blocked by a readvance of the Simcoe ice lobe (Deane, 1950) and discharge returned to the Port Huron outlet (Fig. XII-16g). Retreat of the ice front gave access to the Kirkfield outlet system again, but in the meantime differential uplift is considered to have raised the drift-filled outlet; the water plane may thus have remained at about 605 feet a.s.l. for a lengthy period (Fig. XII-16 h,i,j).

The main Algonquin strandline in the Huron basin is a prominent feature, in many places marked by a strong shore bluff. According to L. J. Chapman, it is not found south of Point Clark, midway up the eastern shore, having been undercut by the present lake. It extends northward into Georgian Bay and eastward around Lake Simcoe to the Kirkfield outlet where, due to differential uplift, it has a present elevation of 870 feet (Deane, 1950). The strong Algonquin strandline may be a result of operation of both the Kirkfield–Fenelon Falls and Port Huron outlets during a halt in isostatic uplift of the region. The writer considers it likely that down-cutting at Fenelon Falls greatly reduced the discharge at Port Huron over a lengthy period. Deane believed the Fenelon Falls channel was about 30 feet deep. Extensive gorges were cut in limestone along Trent River valley above the Sydney strandline of the post-Iroquois lakes; thereafter discharge was appreciably lessened.

The gradient of the warped main Algonquin strandline in Lake Simcoe and Lake Couchiching areas increases northward from 2.8 to 4.0 feet per mile along a line N21°E. Farther north, the highest strandline recorded by Chapman (1954, 1966) of 1,070 feet elevation at Huntsville and 1,245 feet elevation at Sundridge shows a further increase in gradient to 5.0 and 6.3 feet per mile. This high rate of tilt militates against rapid differential uplift during the time represented by the highest shorelines and, accordingly, main Lake Algonquin must have extended northward as far as Sundridge without any appreciable change of level while the ice front retreated in the Huron–Georgian Bay region. Between Sundridge and Trout Creek, however, Chapman (1966) records that the highest strandlines were lowered northward by about 50 feet, and he relates this lowering to differential uplift. If the main Algonquin water plane is projected northward to Trout Creek, however, the amount of lowering would be about 125 feet. Such sudden uplift during the existence of the lake in this area seems unlikely. It is more likely that a period of ice-marginal and subglacial discharge occurred at Fossmill, the waters flowing eastward through Petawawa and Barron Rivers into the Champlain Sea in Ottawa River valley. If the main Algonquin strandline was projected northward to the Fossmill outlet sill at Kilrush (elevation 1,145 feet) it would be at about 1,415 feet elevation. Thus flow at Fossmill outlet over the Kilrush sill might have lowered Lake Algonquin by as much as 270 feet had this not been offset to some degree by differential uplift over a fairly lengthy period and provided the area remained above sea level.

This two-fold process, ice-controlled discharge and differential uplift, may account in part for the long-standing controversy as regards parallelism or convergence of some of the post-Algonquin beaches, namely Ardtrea, Upper Orillia, Lower Orillia, Wyebridge, Penetang, Cedar Point, and Payette. According to Deane (1950) the Fenelon Falls outlet remained in use until after the Ardtrea phase. There had been about 50 feet of differential uplift by Upper Orillia time, and Lake Simcoe was separated from 'Algonquin' waters following the Penetang phase (Fig. XII-16k.)

Prior to this event, during the life of Lake Algonquin, there was a major re-advance of the ice sheet in the Superior and Michigan basins, but apparently it was of little consequence in the Huron basin. This ice advance, the Valders, is interpreted as a glacier surge from northeast of Lake Superior (Fig. XII-16i). The highland east of Lake Superior may well have restricted the flow of ice into the Huron basin while it lobed far southward into Michigan basin. This lobation persisted as the Valders ice receded (Fig. XII-16j) and hence glacial Lake Algonquin was able to invade the Sault Ste. Marie area while being restricted at the north end of Michigan basin and virtually excluded from Superior basin (Hough, 1958).

Glacial Lake Algonquin did not enter the Sudbury basin; strandlines occur at various elevations but all are well below a projected main Algonquin water plane. Farther west, however, higher strandlines are evident and 3 miles northwest of Sault Ste. Marie they reach an elevation of 1,025 feet. This strandline is clearly related, on the basis of isobase trend and rate of uplift, to the highest strandline recorded on Manitoulin Island. Three miles south of Little Current this is given as 1,013 feet, and is reported to be main Algonquin. Projection of the isobase eastward, however, shows it to be about 25 feet lower than the main Algonquin shoreline east of Huntsville. This suggests that Manitoulin Island was ice covered until after some 25 feet of uplift had occurred, probably during transition from the Ardtrea to Upper Orillia phase.

In the northern part of Lake Simcoe area the difference in elevation between Upper Orillia and Payette strandlines is about 140 feet (Deane, 1950, Fig. 7). Thus

operation of Fossmill outlet system over an extended period, with a maximum draw-down at Fossmill of about 220 feet, readily accommodates all the strandlines recognized by Deane. Had there been no uplift of the Fossmill area during the life-span of these post-Algonquin lakes, the Sheguiandah and Korah lake phases, which represent a lowering of 70 feet below the Payette strandline at Sault Ste. Marie (Hough, 1958), might also correlate with discharge at Fossmill over the Kilrush sill. It is probable, however, that as differential uplift was already in progress an additional 50 to 100 feet of uplift may have occurred in the Fossmill area over a period of several hundred years. Accordingly the Sheguiandah and Korah lake phases may be correlated with use of subsidiary routes into the Petawawa River valley. One of these subsidiary routes is through Sobie and Guilmette Lakes, the controlling sill position being east of the latter lake at about 1,125 feet elevation. This site, some 3 miles north of the Kilrush sill in the direction of maximum uplift, permits a further draw-down of 40 feet, and is considered by the writer to correspond with the Sheguiandah phase lake, 35 feet below the Payette level at Sault Ste. Marie. A second outlet route became available when the ice front receded into the Mattawa River valley and exposed part of the Amable du Fond River valley; this led southward to Mink Lake and the Petawawa River system. The sill at Mink Lake is at 1,075 feet elevation and lies on the same isobase as the Guilmette sill (elev. 1,125 feet). Thus a maximum draw-down of 50 feet is possible but there may well have been some 15 feet of uplift as the ice withdrew into the Mattawa valley. The writer therefore correlates the Korah lake phase, about 35 feet below Sheguiandah stage at Sault Ste. Marie, with use of the Amable du Fond outlet over the Mink Lake sill (Fig. XII-16l).

Discharge of the post-Algonquin lakes eastward at Fossmill, and by subsidiary channels, to Petawawa and Barron Rivers resulted in deposition of a large delta, composed largely of sand and some gravel, in the Champlain Sea. Judging by the size of the delta, the spillway must have lain along an active ice front. Gadd (1963), in fact, located a small end moraine in Ottawa River valley near the mouth of Petawawa River. Boulders in the Petawawa–Barron Rivers spillway system are unusually large, in places averaging 2 to 3 feet in diameter. They form ridges and hummocks up to 20 feet high that appear to be bars built by torrential streams.

While the discharge of the post-Algonquin lakes was controlled by Mink Lake sill, immediately prior to eastward discharge down Mattawa River valley, the water depth over North Bay sill was about 400 feet—a calculation based on an uplift rate of 6.5 feet per mile in a direction N20°E and substantiated by strandline positions north of North Bay. As Hough (1958) considers the Payette phase lake had a surface elevation of 465 feet, the Korah phase was about 400 feet. Hough has given the elevation of Lake Stanley, the lowest level lake in Huron basin, as about 190 feet though it may have been appreciably higher according to C. F. M. Lewis. It is thus evident that following perhaps 200 feet of draw-down, as the post-Algonquin Lake discharged eastward along Mattawa River valley, the Huron basin part must have become separated from the lake remaining in the Nipissing Lake basin, which was then lowered a further 200 feet to expose the North Bay sill (Fig. XII-16m). There was also a separate low-level lake in Georgian Bay basin; W. M. Tovell proposes to name it Lake Hough in recognition of J. L. Hough's work on the documentation of the low-level lakes in Michigan and Huron basins.

During the period of ice retreat eastward along Mattawa River valley, the lowering of lake levels must have been accomplished mainly by subglacial discharge. The writer has found no evidence of a surface outlet in the valley between Amable du Fond and a point 5 miles southeast of Mattawa, Ontario. By the time surface (ice-marginal) discharge to the Ottawa River valley began, an over-all lowering from the Korah lake level in Nipissing–Mattawa Valley of about 370 feet had already taken place. Surface discharge then lowered the lake a further 30 feet at which time North Bay sill emerged. Withdrawal of the ice from Ottawa Valley between points 7 and 20 miles east of Mattawa only served to steepen the spillway gradient east of North Bay.

Superior Lake Basin

The Duluth–Minong–Houghton Lakes. Glacier retreat in Superior basin was in a northeasterly direction, resulting in ponding of meltwater at the western end. The history of Superior basin lakes and ice retreat is dealt with by Farrand (1961) and Zoltai (1965). The early phases of Superior basin lakes, as recognized by Farrand, did not affect Canadian shores. Glacial Lake Duluth may have formed a beach on Isle Royale just east of the International Boundary at 1,060 or 1,075 feet elevation, and Zoltai shows Lake Duluth extending along the boundary west of Lake Superior. Lake Duluth discharged southwestward to the Mississippi Valley (Fig. XII-16j). Retreat of the glacier subsequently opened a lower route into Michigan basin that resulted in a series of short-lived lakes. These were formerly considered equivalent to the Algonquin lake phases of Huron basin, but are now thought to be independent and are referred to by Farrand as the post-Duluth lakes (Figs. XII-16j–m). Beaches referable to some of these lakes occur on Isle Royale and west of Fort William. The latter, however, partly represent a separate and earlier lake in the Kaministikwia River basin (Zoltai, 1965).

When glacier retreat from the eastern end of Superior basin allowed the post-Duluth lakes free access to Whitefish Bay they may have become briefly confluent with post-Algonquin lakes in Huron basin. Somewhat later the glacier receded entirely from Superior basin. Farrand refers to the resulting lake as Lake Minong; it was responsible for the highest shoreline features in the northeastern

end of the lake basin. The upwarped beaches occur between elevation 950 and 1,000 feet on the north shore. Farrand relates Lake Minong to Sheguiandah lake phase in Huron basin but the writer considers it to be somewhat younger and with its outlet at Sault Ste. Marie (Figs. XII-16n, o).

Below the Minong strandline on the north shore of Lake Superior there are a succession of beach ridges or small wave-cut bluffs, depending on the character of the shore, that Farrand relates to post-Minong lake phases. They have a vertical range of over 50 feet. The lowest and best formed shoreline at 750 to 765 feet near Dorion is known as the Dorion beach and appears to have resulted from outlet adjustments rather than differential uplift (Figs. XII-16p, q). Another adjustment was responsible for a further lowering of several tens of feet, and development of a shoreline that Farrand considers correlative with the low-level lake phases in Huron basin. He named the low-level lake in Superior basin as Lake Houghton and reports that it discharged to the Huron basin by the proto-St. Marys River at Sault Ste. Marie. He considers the elevation of the lake as about 360 feet a.s.l. Its shorelines have been destroyed along the east side of Lake Superior by a younger lake but are preserved on the north shore as a result of differential uplift (Figs. XII-16r, s).

Upper Great Lakes Region

The low-level, Nipissing, and post-Nipissing lakes. As mentioned earlier, glacier retreat from the eastern end of Mattawa River valley served to drain the lakes in the upper Great Lakes basins to very low levels. The indicated amount of draw-down at North Bay also necessitates establishment of separate lakes in the Lake Nipissing, Georgian Bay, and main Huron basins (Fig. XII-16m). Erosion of outlets from each of these basins allowed for continuing draw-down of the upper lakes over a very long interval during which differential uplift was raising the northeastern end of the drainage system. Thus the lowest levels of Stanley, Chippewa, and Houghton were not established simultaneously. These lakes remained until their outlets were drowned as water encroached from the east consequent upon the continuing uplift of the North Bay region. A speculative configuration of the lakes is shown in (Figs. XII-16m–w). Dependant upon the assumptions made as regards both the ages of the several post-Algonquin lake phases and the amount and rate of differential uplift during their life-span and during the subsequent low-level lake phases, various ages may be deduced for these low-level lakes. Lake Stanley may have merged with Lake Chippewa as early as 10,400 years B.P. rather than as late as 8,100 years B.P. as shown. Perhaps an age of 9,500 years B.P. is more likely than either extreme but it is necessary to allow sufficient time for the re-excavation of Mackinac channel between the Michigan and Huron lake basins.

When North Bay was raised to 605 feet a.s.l. discharge began again at Port Huron and at Chicago. Continuing differential uplift finally raised North Bay outlet above the lake surface and drainage was entirely by the southern outlets. Long use of the southern outlets resulted in a rather stable lake level and a consequent well-marked shoreline, that of the Nipissing phase of the Great Lakes (Fig. XII-16x). The confluent water bodies in Huron, Michigan, and Superior basins with discharge at Port Huron and Chicago are referred to as the Nipissing Great Lakes. The Nipissing terrace is well displayed a short distance west of North Bay at 700 feet elevation whereas the present shore of Lake Nipissing is at 648 feet. The highway to Sault Ste. Marie crosses the Nipissing terrace many times.

The Nipissing phase of the Great Lakes is usually dated at about 4,200 years B.P. Many organic deposits from positions below the Nipissing strandline, in various parts of the Great Lakes area, have been given ages ranging back to more than 7,000 or 8,000 years B.P.; these are commonly referred to the interval of pre-Nipissing rising water levels. However, recent dates from Little Pic River, north shore Lake Superior, indicate the age of a Nipissing phase lake to be about 6,000 years B.P. Wood from silty clay beds beneath a 15-foot capping of sand that forms a terrace at the highway bridge over this river was dated at $5{,}920 \pm 120$ and $5{,}960 \pm 120$ (GSC-83, 103). The terrace has an elevation of about 700 feet. Farrand (1961) considered the deposit Nipissing/Algoma transition as he determined the elevation at 692 feet and found a cobble ridge at 718 feet which he considered to be the Nipissing beach. The writer and S. C. Zoltai noted a transition from plant-bearing silty clay, with shells in the upper few feet, upward into shell-bearing (mainly *Sphaerium sulcatum*) sand. Tiny plant tissues from the basal few inches of the sand were dated at $6{,}100 \pm 160$ (GSC-285). The writer therefore considers the plant-bearing silty clay, the shell-bearing sand, and the boulder beach to represent the same lake phase, probably the Nipissing. On the eastern side of Lake Huron, 4 miles northeast of Owen Sound, wood has been found in the basal part of an extensive gravel ridge that L. J. Chapman considers an undoubted Nipissing beach. The gravels are 18 feet thick, pass down-slope into sand, and rest on a silty clay. The wood from the sand-clay contact was dated at $5{,}770 \pm 130$ (GSC-347). Fourteen miles to the east at Meaford, wood was found in clay beneath 8 feet of sand with surface elevation of 605 feet that is also considered related to a Nipissing phase lake. The wood was dated $6{,}300 \pm 150$ years (L-312).

Radiocarbon dates on basal organic materials from small lakes at the Nipissing level on Manitoulin Island (Lewis, 1968) prove that a Nipissing phase was near maximum about 5,500 years ago. Similar evidence from North Bay reported by Lewis indicates that the spillway ceased to function before 5,000 years B.P. It is thus evident that the Nipissing Great Lakes (605 feet a.s.l.) formerly considered to be in the order of 4,200 years B.P. are in fact a combination of lake phases reflecting

first the use of three outlets and later the use of only the two southern outlets, the whole ranging from about 6,000 to 4,200 years B.P. The three-outlet phase may be termed Nipissing Great Lakes I (Fig. XII-16x) and the two-outlet phase Nipissing Great Lakes II.

Erosion of the southern outlets of Nipissing Great Lakes II lowered the water level until a segment of Port Huron spillway system became stabilized by a combination of bedrock and concentration of boulders at the same time that Chicago outlet was rock-controlled; the resulting halt in the lowering of the outlets initiated the Algoma phase lake at about 596 feet elevation (Hough, 1958). This event may have occurred about 4,000 years ago (GSC-301). As the northern parts of the lake basins were slowly uplifted, a lateral shift in the Port Huron channel into more easily eroded materials brought about a resumption of down-cutting. Uplift of the sill at Sault Ste. Marie gave rise to a separate lake in Superior basin known as the Sault stage (Farrand, 1961) possibly about 2,000 to 3,000 years ago. This was followed by a sub-Sault stage and finally Lake Superior at 602 feet elevation. In the meantime down-cutting at Port Huron lowered the Huron and Michigan basin waters to their present 580 feet elevation.

Northwestern Ontario

Retreat of the ice in northern Ontario beyond the Superior and Huron basins is dealt with by Zoltai (1961, 1963, 1965) and Boissonneau (1966, 1968). Prior to its retreat from Superior basin, the Laurentide Ice Sheet west of Superior had advanced southwest into Minnesota. As it receded northward into western Ontario it remained in contact with glacial Lake Agassiz over a wide front; the history of this lake is discussed separately. Fluctuations in the rate of ice flow resulted in construction of a number of noteworthy end moraines. The most southerly end moraine system in western Ontario trends southeastward from Lake of the Woods to Rainy Lake (Map 1253A). Following an ice-frontal retreat of about 75 miles to the northeast the Steeprock Moraine was constructed; this extends from near Steeprock Lake southeastward to International Boundary about 25 miles west of Lake Superior. Shortly thereafter the much more extensive Eagle-Finlayson Moraine was constructed; this may be traced for about 175 miles southeastward from 30 miles northwest of Kenora to within 30 miles of Thunder Bay, where it is truncated by an end moraine left by a Superior basin ice lobe. Following a major retreat in western Ontario, a re-advance of the ice front was responsible for the Hartman Moraine which Zoltai (1965) considers to be of Valders age. At its northwestern end it curves sharply to the north and then northeast until truncated by the younger Lac Seul Moraine. When the ice front was at the Hartman Moraine, major ice lobes were also active in Nipigon and Superior basins and built the correlative and adjoining Dog Lake and Marks Moraines respectively. The latter has been traced to a high bedrock area southwest of Thunder Bay and hence the 'Valders' ice border may cross the International Boundary only a few miles west of the Superior shore.

Some 10 to 20 miles northeast of Hartman Moraine, the ice sheet built the extensive Lac Seul Moraine, which curves around the western side of Lac Seul. To the southwest this moraine merges with the Kaiashk Interlobate Moraine (Zoltai, 1965), which was built in part between the ice lobes that produced the Hartman and Dog Lake Moraines, but Zoltai has suggested that the northeastern end of the Kaiashk Moraine is a reconstructed end moraine of the Lac Seul lobe. A retreat followed by a re-advance of the ice front, of at least 20 miles, is indicated by overridden varved clays in Lac Seul area (Zoltai, 1965). Lac Seul Moraine extends northwestward into Red Lake–Lansdowne House area (Prest, 1963). On the southwestern side of Trout Lake it rises sharply some 270 feet but on the distal side it slopes more gently to the southwest and displays many excellent beaches and terraces of glacial Lake Agassiz. The moraine was produced beneath this lake and only three small parts remained above lake level. To the northwest the moraine is lower and less distinct but a series of large, wave-modified ridges of sand and gravel, unlike adjacent reworked eskers, may be morainal and indicate the former position of the ice lobe. The end moraine appears to loop northward and then eastward towards Windigo Lakes where it is truncated by end moraines of the minor Windigo ice lobe. Ice-flow features and eskers also indicate the convex form of the Lac Seul ice front. The westward lobation of the ice sheet was probably a result of active calving into the deep water of a Lake Agassiz embayment in the ice front (Fig. XII-16m).

Following a short period of recession in western Ontario, a halt in retreat or minor re-advance of the ice sheet produced the Sioux Lookout Moraine on relatively high ground west of Lake Nipigon. To the northwest are sporadic De Geer moraines which delineate the ice-frontal positions during calving into Lake Agassiz. Following a retreat of at least 80 miles the northern part of the ice sheet halted to form the Whitewater Moraine, and an eastern lobe of the ice re-advanced to form the Nipigon Moraine along west side of Nipigon basin (Zoltai, 1965). The writer believes that only the northern part of this end moraine system was built at this time (Fig. XII-16p), the southern part being older.

Greater activity of a northern component of the receding ice sheet relative to an eastern component gave rise to a series of recessional end moraines in Windigo Lakes area in northwestern Ontario and to an interlobate moraine south of these lakes.

The next major moraine system left by the fluctuating though generally receding ice sheet in northwestern Ontario was the extensive and generally southeast-trending Agutua Moraine (Tyrrell, 1913; Prest, 1963). A 25-mile re-advance is indicated by subtill lake clays along Otoskwin River. The highest part of the morainal system is at a

major bend in Albany River west of Miminiska Lakes where it rises 500 feet above the river. The Crescent Moraine system (Zoltai, 1965) north of Lake Nipigon is probably a correlative of Agutua Moraine.

After construction of Agutua Moraine the northern ice sheet receded some 20 miles to the northeast before halting long enough to leave several short segments of end moraine on-line with belts of De Geer moraines. The eastern ice sheet at this time appears to have thrust actively forward—the Miminiska ice advance—and overrode parts of the Agutua Moraine in the vicinity of Albany River, and left a complex of end moraine segments, interlobate moraine, eskers, and ice-flow features that serve to delineate its form. To the southwest, in the vicinity of Ogoki River, this same ice advance appears to have constructed the Nakina Moraines (Zoltai, 1965).

The northern lobe, following a further retreat of about 30 miles, halted to construct a moraine near Big Beaverhouse on Winisk River. This moraine trends northwestward from south of Wunnummin Lake to Little Sachigo Lake. The eastern lobe left little evidence of its ice-frontal position; it may have been responsible for short segments of moraine west of Lansdowne House, and may have extended south-southeastward through Ogoki Lake. A short distance eastward from this lake the ice-flow features are masked or destroyed by the younger Cochrane ice advance from the north (*see under* Northern Ontario and Western Quebec).

Glacial Lake Agassiz. Glacial Lake Agassiz occupied large areas in Minnesota, North Dakota, Saskatchewan, Manitoba, and Ontario as Laurentide Ice Sheet receded north of the Mississippi drainage divide. As the lake expanded northward its southern limit contracted and, although the total area covered by Lake Agassiz was more than 200,000 square miles, it was probably no more than about 80,000 miles in extent at any time (Elson, 1967).

The history of glacial Lake Agassiz is, in part, intricately related to the period of ice-sheet recession from northwestern Ontario previously described. During this recession the outlets were controlled by position of the ice front and by topography in northwestern Ontario. Many of the strandlines were formed as a result of the operation of eastern outlets that discharged into Nipigon basin and thence southward into Superior basin.

Though the existence of a former huge lake in the Red River basin in the United States was recognized as early as 1822 and named in honour of Louis Agassiz in 1879, the first comprehensive work was that of Upham (1895) and later that of Leverett and Sardeson (Leverett, 1932). Many have also contributed new information on the Canadian part of the lake basin, particularly Johnston (1946) and Elson (1957, 1967). Information on eastern outlets discharging into glacial Lake Kelvin in Nipigon basin and the extent of Lake Agassiz in northwestern Ontario has resulted from field studies by Zoltai (1961, 1963, 1965) and Prest (1963). Elson has established a multifold history of Lake Agassiz. He recognizes a series of progressively lower lake levels related to successive operation of southern, northwestern, eastern, and northern outlets, with some rises in lake level due to ice-marginal fluctuations in the spillway areas. He has related strandlines in the Red River basin to these various outlets. The writer has also attempted to restore the configuration of some Lake Agassiz phases relative to the eastern outlets using the rates of tilt as determined by Johnston, the spillway positions and elevations given by Zoltai, his own data on lake levels and ice-margin positions in northwestern Ontario, and recently available contoured topographic maps.

Lake Agassiz began to form as Laurentide ice receded northward into the headwaters of Red River basin along the Minnesota–Dakota boundary, and it expanded northward into Manitoba and northwestern Ontario bordering on both Keewatin and Labrador sectors of the ice sheet. Discharge was through Lake Traverse at the south end of the Red River system into Minnesota River of the Mississippi River system. Lake Agassiz was lowered some 80 to 90 feet as the outlet was eroded to a bedrock sill. The lowering was irregular, perhaps due to formation of boulder armaments along the spillway that were periodically removed by increase in rate of discharge. Four lake phases, the Herman, Norcross, Tintah, and Campbell, and numerous minor intermediate strandlines are recognized in southern Manitoba and northwestern Ontario (Figs. XII-16d–f).

Elson (1966, p. 92) has suggested that a lowering from and return to the Norcross level was related to a short period of eastward discharge into Lake Superior via a Dog River spillway route in northwestern Ontario (Figs. XII-16g, h). The writer believes that this early phase of discharge may account for the first low water phase of Lake Agassiz perhaps about 12,000 years ago (Elson, 1966, Fig. 6). Sandy peat from alluvial fill at about the Campbell level in Assiniboine River valley and associated with a deepening lake was dated at 12,400±420 years (Y-165). When an ice advance plugged the Dog River outlet, discharge was returned to the south. Lake levels rose to the Norcross level due to in-filling of the Lake Traverse outlet route in the interim. As this sediment was removed by erosion Lake Agassiz was lowered again to the Campbell level, controlled by the bedrock sill (Figs. XII-16i, j). The ice front at this time was at the Hartman and Dog Lake Moraines which are considered to be of Valders age.

When the ice-sheet margin again receded, the Dog Lake spillway was not re-opened; eastward discharge was not possible until the ice front receded northeastward beyond Sioux Lookout, whereupon the lake began to discharge into Nipigon basin (Fig. XII-16k). This event occurred about 11,000 years ago. A re-advance of the ice front closed the eastern outlets, returned the lake level to the Campbell strandline, and built the Lac Seul Moraine (Figs. XII-16l, m). Retreat of the ice again uncovered

outlets into Nipigon basin with resulting lowered lake levels (Fig. XII-16n). A re-advance of the ice to build the Sioux Lookout Moraine about 10,000 years ago again closed the eastern outlets and raised Lake Agassiz to the Campbell level for the last time (Fig. XII-16o). Thereafter glacier retreat resulted in operation of ever-lower eastern outlets; a succession of prominent beaches was built in parts of the Red River basin. As the lake surface was lowered, more and more ground was exposed west of Nipigon basin and spillways lengthened westward (Fig. XII-16p).

During operation of the lower eastern outlets the ice receded from the region northwest of Nipigon basin (Prest, 1963) and exposed a lower outlet route that led southeastward along the ice margin to the northern end of Nipigon basin. As there was a major retreat of the northern ice front prior to the Agutua advance, low ground may have allowed discharge from Lake Agassiz southeastward into glacial Lake Barlow–Ojibway; if so, the Agutua re-advance has obliterated all trace of these outlets. Some readjustments in the drainage routes no doubt took place as the ice front receded from Agutua Moraine, and the routes were deranged again as the ice advanced to the position of the Nakina Moraines. At this time Lake Agassiz was confluent with glacial Lake Nakina (Zoltai, 1965) and the outlet was into the northern end of the Nipigon basin (Fig. XII-16r).

The retreating ice sheet remained in contact with Lake Agassiz until it receded from the position of the Big Beaverhouse Moraine. Discharge may have remained to the southeast, possibly into Barlow–Ojibway, as the moraine was built, but then changed to northward, probably down Echoing River valley into Hudson Bay Lowland. The drop in lake level occasioned by this event severed Lake Agassiz from the ice-marginal lakes remaining along the ice front on the Ontario part of the Hudson Bay watershed (Fig. XII-16s). At a still later date Lake Agassiz discharge was diverted to the Nelson River system and as the last confining ice melted the glacial lake gave place to early phases of Lakes Winnipeg, Winnipegosis, and Manitoba.

Northern Ontario and Western Quebec

Glacial Lake Barlow–Ojibway. In northern Ontario, east and northeast of Lake Superior, glacier retreat was northward towards James Bay, and in the adjacent parts of western Quebec it was mainly northeastward towards central Quebec. Lake waters were ponded between the ice front and the height-of-land (Figs. XII-16m–q). The manner by which this great lake remained at high levels astride the Hudson Bay–St. Lawrence drainage divide area and discharged southward along the valley of Lake Timiskaming has long been a subject of controversy. The outlet was plugged by a moraine some 10 miles north of Temiscaming, Quebec, and it is likely that lake levels were stabilized as a result of boulder concentrations in the spillways. New gradients and lake levels probably resulted from periodic isostatic adjustment. Progressive shallowing of the lake is indicated by sand horizons in the deposits of varved clay and by a greater rate of tilt on the highest strandlines in the southern part of the basin relative to those in the northern part. The writer determined a rate of tilt from the outlet to Larder Lake of about 4 feet per mile, and from Larder Lake to Plamondin Hill of 2.1 feet per mile. Hughes (1965) states that the maximum tilt of short segments of lower strandlines varies from 2.1 to 3.8 feet per mile and that the maximum rate of tilt must be in excess of 3.8 feet. The location of the hinge-line has not been established but it is probably to the south of the outlet. It is possible that differential uplift of the northern part of the basin took place along a second hinge-line in the Larder Lake–Noranda area.

A plot of the maximum waterplane from Plamondin Hill to Larder Lake indicates a minimum depth of water, over the present drainage divide near Noranda, of about 300 feet. It is thus evident that a combination of uplift and erosion of the outlet near Temiscaming must have lowered the lake by this amount before glacial Lake Ojibway was confined to the north; it then discharged from a point 15 miles west of Noranda (Fig. XII-16r) southward into ancestral Lake Timiskaming. The drift plug at Temiscaming has been channelled to a depth of over 350 feet; a small part of this no doubt occurred in post-Barlow–Ojibway times.

Limited information on maximum strandlines east of Malartic, Quebec, indicates that lake levels had lowered appreciably before the Quebec highlands glacier receded from that part of the Lake Barlow–Ojibway basin. A puzzling matter concerning Lake Barlow–Ojibway is the sudden deepening of the lake, indicated by increase in varve thickness and decrease in secondary sand, in varve year 1528 when the ice front was in the vicinity of Cochrane (Hughes, 1965). Perhaps discharge had been westward into Superior basin for some time prior to this varve year, whereupon a re-advance of the ice front closed this outlet, deepened the lake, and returned discharged to Timiskaming channel. This re-advance may correlate with formation of either the Crescent or the Nakina Moraines.

During retreat of the ice front northward from the Hearst–Cochrane region a prolonged halt occurred in the vicinity of Fraserdale, 60 miles north of Cochrane. A major east-west kame moraine composed of silt, sand, and some gravel was deposited in contact with the ponded waters on its south side (Boissonneau, 1966). This moraine may correlate with interlobate moraine west of Hearst and north of Hornepayne. These deposits are mainly sand and gravel with an appreciable local content of limestone. All these moraines may correlate with the Nakina Moraines (Zoltai, 1965) farther northwest; indeed there are no other comparable pre-Cochrane moraines in northwestern Ontario.

The lake in front of the moraine near Fraserdale is believed to have discharged southward, via an outlet near

Noranda, into Lake Timiskaming basin. This drainage route is in accord with the Barlow–Ojibway isobase trends and an indicated lowering of lake levels by about 300 feet in Lake Abitibi region. When the ice retreated north of the kame moraine, however, discharge and complete drainage must have been towards Hudson Bay, either laterally or subglacially. Antevs (1925, p. 77), with regard to the immediate pre-Cochrane stage of Barlow–Ojibway, reported that discharge into Hudson Bay took place across thin remnant ice in James Bay. Later (1931) he thought that drainage around the ice front to the sea occurred in the vicinity of the mouths of Hayes and Nelson Rivers, Manitoba. The writer considers that the lake, here named glacial Lake Antevs (Fig. XII-16s), may have drained northwestward along a line of weakness between a large mass of semi-stagnant ice in northwestern Ontario and a major glacier in James Bay and southern Hudson Bay, and discharged into the sea in the easternmost part of Manitoba. It is assumed that the sea had entered Hudson Bay via Hudson Strait and then along the line of a southwest-trending bottom channel in Hudson Bay, and effectively severed the Keewatin and Labrador sectors of the Laurentide Ice Sheet at this time.

Cochrane phase of recession. The next major event in the recessional history of the Laurentide Ice Sheet is the emplacement of the Cochrane till in Hudson Bay Lowland. Ice from Hudson Bay evidently moved southward more or less along the flank of the Shield (Cochrane I); somewhat later there was a southwestward and southward thrust from James Bay region (Cochrane II). The Cochrane ice front is seldom marked by end moraine. As pointed out by Hughes (1956) "The term 'Cochrane moraine' . . . is misleading, and should be dropped in favour of 'Cochrane till,' for the Cochrane till is a stratigraphic unit recognizable over a considerable area." The Cochrane limit is merely a line beyond which Cochrane till has not been found. Beyond the till limit the Cochrane event is represented by a series of varved sediments (Connaught sequence) containing pebbles lithologically similar to those of the till, but contrasting sharply with pebbles in underlying varved sediments (Frederickhouse sequence) that comprise a large part of Barlow–Ojibway deposits. Hughes reports counting about 60 varves of Connaught sequence in individual exposures. Maximum varve thickness is attained at varve 25, and this is considered to correlate with emplacement of the Cochrane till farther north. The Connaught sequence also includes varved sediments deposited in a shallow remnant of glacial Lake Antevs during recession of the Cochrane ice.

The typical Cochrane till is a stone-poor, blue-grey clay till that rests on varved sediments. It takes on a very pale pinkish tint when weathered, and is a yellow-brown when oxidized. Being calcareous, it contrasts markedly with the older, sandy, and non-calcareous tills of the Timmins–Cochrane region (Hughes, 1965). The clayey character of the till is believed to be due mainly to the incorporation of lake clays as the Cochrane ice advanced southward.

Boissonneau (1966) refers to an ice-front retreat of several miles to the north of the kame moraine near Fraserdale on Abitibi River, prior to the advance that deposited a thin mantle of clayey till over parts of it. The writer has observed the Cochrane clay till overlying varved sediments near Coral Rapids (364 feet elevation). The Cochrane till has been recognized as far north as the mouth of Little Abitibi River at 200 feet elevation but in this area its clayey character is produced from older, dark grey clay and light buff limestones. The ice front, prior to the Cochrane advance, probably receded to the vicinity of Coral Rapids where it impinged on a land surface at a present elevation of under 400 feet a.s.l.

Recession of the ice sheet down the Hudson Bay watershed from Timmins at 1,029 feet elevation to Coral Rapids at 364 feet elevation, a distance of 110 miles, would seem to indicate a lowering of the ice-sheet surface by about 600 feet. As the most southerly point where Cochrane till has been recognized is 12 miles north of Timmins at an elevation of about 960 feet, this would seem to require a thickening of the ice sheet of some 500 to 600 feet during the Cochrane advance. Such a thickening of the ice sheet is improbable at this late stage of deglaciation especially when marine water of the Tyrrell Sea (Lee, 1960) flooded much of Hudson Bay Lowland shortly thereafter (I-GSC-14, 7875 ± 200 years). It is more probable that the emplacement of Cochrane till is somehow related to the incursion of the sea into Hudson Bay. The two main surges of ice, envisaged by the writer as responsible for the Cochrane ice-flow features and emplacement of the clayey till evident on Map 1253A, are shown in Figures XII-16t, u.

Northern Quebec and Labrador

The retreat of the last ice sheet in this region was mainly towards the higher terrain of the interior, and the pattern of ice-flow features and eskers is generally radial about an interior U-shaped area south and southwest of Ungava Bay within which the last remnants of ice melted. Glacial lakes were not common but a few were extensive and long-lived.

Recession from Labrador and Ungava coasts. At the last glacial maximum the ice sheet in northern Labrador extended major tongues through cols in the Torngat Mountains, spread laterally in the coastal areas, and calved into Labrador Sea a short distance off the present coast. The upper limit reached by these valley glaciers has been referred to as the Saglek level by J. T. Andrews and the event itself to the Saglek Glaciation (Andrews, 1963; Løken, 1962). The elevation of the trimline, which becomes lower to both the north and east, indicates the surface slope of the ice sheet which lay west of the Torngat Mountains. Farther south, in the coastal Kaumajet and Kiglapait Mountains which at one time were believed

to have remained unglaciated, erratics on the higher peaks indicate ice flow from the west at some earlier glacial maximum. This was considered Wisconsin (Tomlinson, 1959) but, in the light of J. T. Andrews work, is not likely Classical Wisconsin. Various levels of ice movements have been noted in coastal Labrador by many workers but it is not always clear which observations are related to the last regional glaciation (Daly, 1902; Wheeler, 1958; Ives, 1958a, b, 1960a; Tomlinson, 1959; Andrews, 1963). All agree that cirque glaciation in the coastal mountains was minor, thus supporting the concepts of Tanner (1944). Cirque development followed the retreat of the ice sheet and probably also took place in the Little Ice Age. According to Tomlinson, many nunataks and a large strip of continental shelf appeared as the ice sheet retreated from the coastal area, prior to transgression by the sea.

Shells from Eclipse Channel near the northern end of the Torngat Mountains indicate that the sea encroached on the northern Labrador coast prior to 9,000 ± 200 years ago (L-642). The shells were at an elevation of 95 feet in a region where Løken has placed the marine limit at 185 feet. On the basis of pollen analysis, Wenner (1947) placed deglaciation of the central Labrador coast prior to 10,000 years B.P.

Torngat Mountains were mainly nunataks at the maximum of the last major glaciation (Ives, 1958a; 1960a, Fig. 3), though overridden during an earlier glaciation formerly considered maximum Wisconsin (Ives, 1957). As the Labrador ice thinned and the supply of inland ice was cut off at the cols, the trunk glaciers in the through-going mountain valleys stagnated. The mountains themselves supplied little ice; some cirque glaciers formed, but seldom cut through the lateral moraines of the trunk glaciers. Løken (1960) has reported some late, mountain ice caps in the Abloviak River drainage basin east of Ungava Bay and, on the nearby Labrador coast, has noted features around Ryans Bay indicative of a re-advance of valley glaciers late in the deglacial process. In Okak Bay area between the coastal Kaumajet and Kiglapait Mountains, Tomlinson (1959) also found evidence of a late stage re-advance that formed push moraines prior to the final waning of the ice sheet. Both there and farther south Wheeler (1958) and J. T. Andrews noted high-level lateral moraines formed when the ice front lay east of the present coast, and also prominent younger moraines believed to represent a late glacial re-advance. These eastward re-advances across the low, southern Torngat Mountains may indicate that in the region south of the Torngat the Labrador ice was less restricted than to the north and flowed readily towards the coast. During retreat it maintained an active ice front but, as a result of a rapid thinning, did not form moraines.

Glacial lakes. As the ice sheet thinned west of Torngat Mountains, the eastern outlet glaciers retreated and left numerous lateral and end moraines along the Labrador coast. Later the ice margin receded west of the drainage divide and meltwaters were ponded in two of the major stream valleys east of Ungava Bay (Ives, 1958b, 1960a; Løken, 1962). At the same time, farther south the ice sheet receded across the plateau between Torngat Mountains and George River basin. Discordance between the drumlinoid and esker trends may indicate that recession of the ice sheet was not along the same path as the flow during the earlier advances (Matthew, 1961).

In the George River drainage basin the meltwaters were ponded to form numerous high-level, ephemeral lakes, followed by the long-continued glacial Lake Naskaupi (Ives, 1960b). Several major stands of this lake are recognized. The remarkably well developed shorelines of glacial Lake Naskaupi II are related to discharge down Koraluk (formerly Kogaluk) River to Labrador Sea. Other discharge routes are via the Fraser, Harp, and Kanairiktok Rivers and via the George River–Lake Michikamau col but their respective lake phases have not been determined. At its maximum development, glacial Lake Naskaupi was about 200 miles long and 30 miles wide, occupying the entire George River basin except for the ice-dammed northern end. The correlative glacial Lake McLean (Ives, 1960b) formed in the upper part of Whale River basin immediately west of George River, and discharged at its southeastern end into Naskaupi II. It was about 70 miles long and 30 miles wide, and was confined on both northwestern and southwestern sides by the ice sheet. The strandlines of these lakes are tilted up towards the southwest and the Labradorean centre (Barnett and Peterson, 1964).

The development of the Naskaupi and McLean strandlines, which are cut in bedrock in places, required a long stand of each of the several lake phases. This requires rock-controlled sills at the outlets, and an active confining ice barrier in Ungava Bay (Ives, 1960b). The esker and ice-flow pattern south of Ungava Bay appear to be in accord with the concept of an Ungava Bay ice dome, but most evidence as to the sense of glacier movement south of the bay indicates northward rather than southward flow with recession of the ice sheet towards the interior (Matthew, 1961). Northward flow would appear to be a natural consequence of incursion of the sea into Ungava Bay. Within George River basin itself Matthew noted that certain ice-flow features indicate westward flow where only eastward flow might be expected because of the ice-dammed lakes. Thus there has been considerable controversy over the position of the ice divide. The writer considers that three periods of ice-divide or ice-dome migration are needed to satisfy the field data and the following glacial history is suggested: (i) an early north-trending ice divide must have been present within George River basin (Map 1253A) to account for westward ice-flow features within the basin; (ii) as Ungava Bay and the lowland to the south filled with ice the ice divide shifted westward; (iii) eastward ice flow was maintained in George River basin during

the early stages of deglaciation but did not remove all evidence of the earlier westward movement; (iv) with the incursion of the sea into Ungava Bay a major rearrangement of the ice flowage resulted; (v) a short-lived ice divide probably then lay a short distance west of glacial Lakes Naskaupi and McLean, trending slightly east of north; (vi) ice flow was everywhere towards Ungava Bay, rapid southward recession of the ice divide followed, and the glacial lakes drained northward; (vii) the ice divide then lay south of Ungava Bay and no longer had a north-trending extension; subsequent changes in position took place in the interior as recession continued.

As the ice sheet thinned due to climatic amelioration, glacier retreat from northern New Quebec, west of Ungava Bay, followed as a natural consequence of the opening of Ungava Bay and Hudson Strait. Ice dispersion was outward towards the coast. As the ice sheet thinned and receded over the height-of-land near the coast, meltwater was ponded along the ice front southwest of Hudson Strait. The highest lake at about 1,800 feet elevation was ponded in the headwaters of Povungnituk River and drained eastward into Joy Bay (Ives, 1960a). As the ice margin receded southwestward down Povungnituk River valley it became concave to the northeast, and successive lower lake phases discharged northward to Hudson Strait and eastward to Ungava Bay (Map 1253A). Maximum lake levels were at about 1,500, 1,200, and 1,000 feet a.s.l. and several systems of deep spillway channels occur. Data available do not warrant conclusions regarding the direction or amount of postglacial uplift. Matthews (1962) outlined small lakes along the ice front south of Cape Wolstenholme. These presumably drained westward or southwestward and were somewhat younger than at least part of the above-mentioned larger lake system.

Recession from Hudson Bay Coast. As Labrador ice receded inland from northern Hudson Bay it calved actively into the sea, and between Cape Wolstenholme and Port Harrison left a magnificent series of De Geer moraines; these preserve the lobate form of the ice front in most valleys along the coast. These coastal moraines may correlate with a similar belt of De Geer moraines well inland from east coast of James Bay. The lack of De Geer moraines in the coastal belt east of James Bay may indicate that the ice was too thick for their formation at the time the sea encroached on the region; east of Richmond Gulf, the near absence of De Geer ridges is more likely due to thin ice relative to the depth of the sea.

Recession to the interior. Inland from the area of marine overlap, ice recession left the usual pattern of eskers and

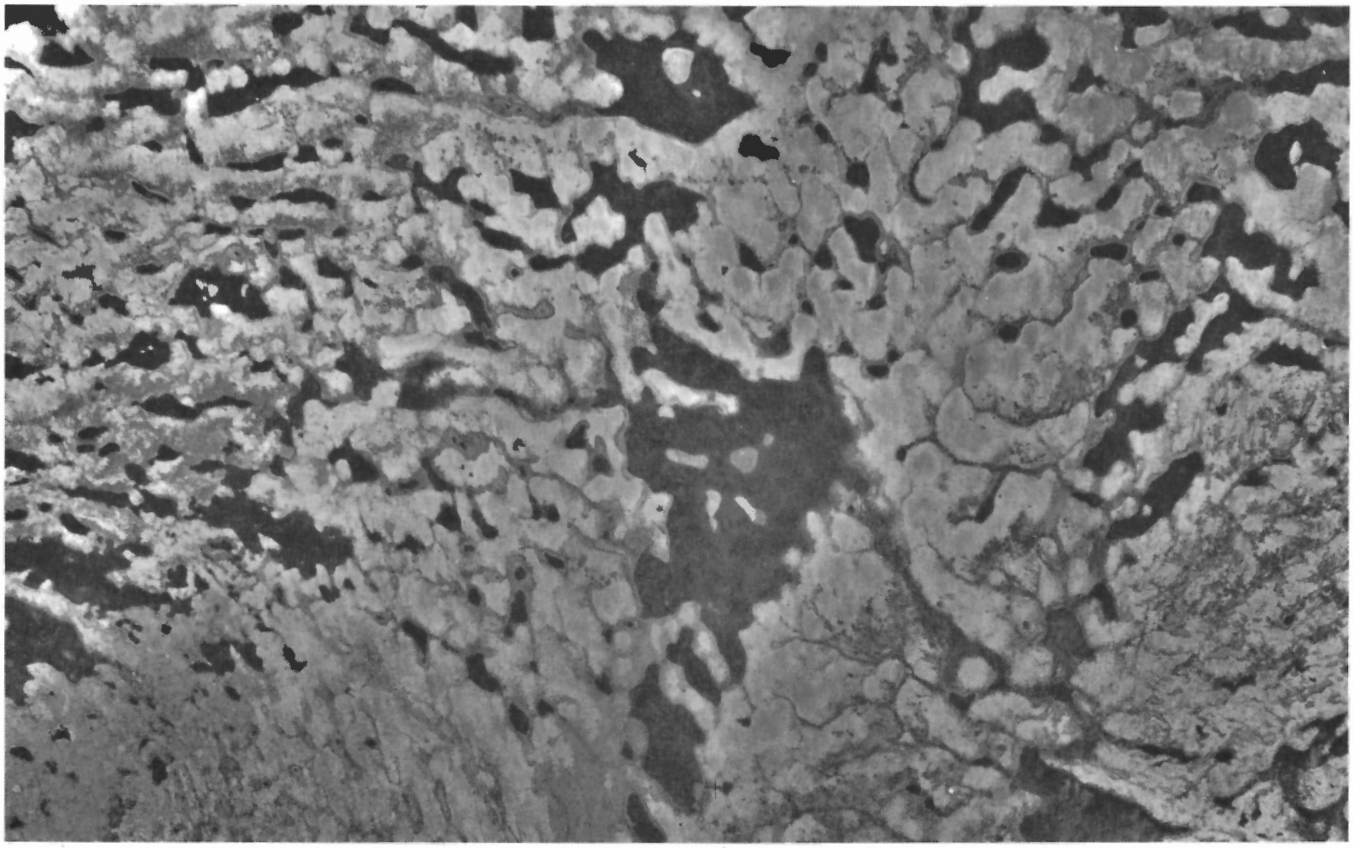


PLATE XII-8. Fluted ribbed moraine south of Ungava Bay, Quebec. Ice flowed northward. Photo suggests close temporal association between ribbing and fluting. Vertical airphoto; scale 1 inch to 1 mile.

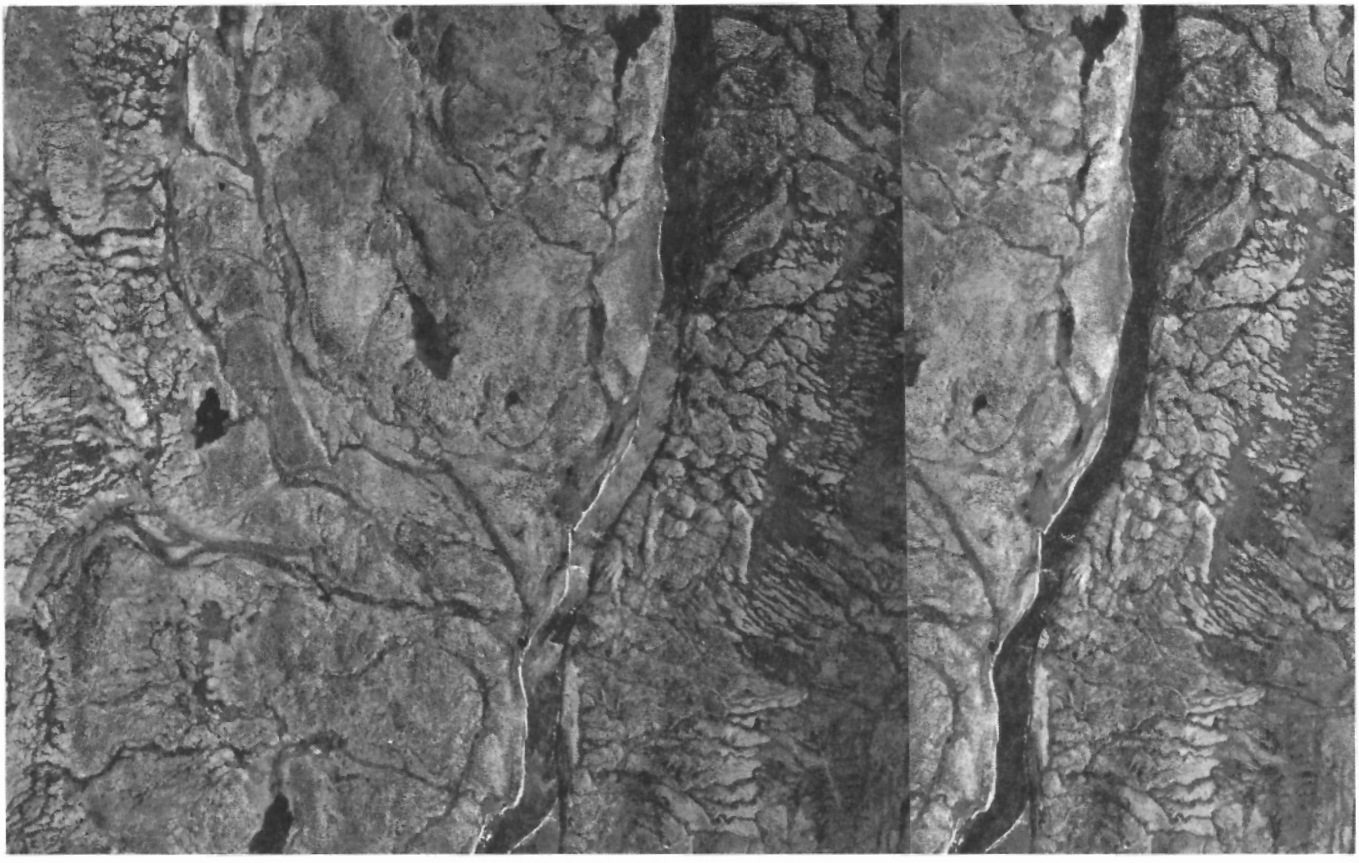


PLATE XII-9. Ice-marginal and subglacial meltwater channels, Mushalagan River, Central Quebec. The ice receded northward. Series of small meltwater channels formed subglacially or in much crevassed ice; the larger channels are considered ice-marginal channels. Stereoscopic pair; scale 1 inch to 3,330 feet.

ice-flow features characteristic of the Canadian Shield. Ribbed moraine is prevalent in many places associated with other ice-flow features and with eskers (Pl. XII-8). Final recession was probably to a number of scattered centres or ice divides in the interior of northern Quebec peripheral to Ungava Bay (Henderson, 1959b; Ives, 1960c; Derbyshire, 1962). One of the last remnant bodies of ice was in Howells valley west of Schefferville where there was much shifting of the late centres of outflow, with topography in control of the final ice movements. Late ice remnants were also present in Knob Lake depression and in the Lac le Fer–Swampy Lake depression. Subglacial and submarginal meltwater channels (Pl. XII-9) were the most effective forms of drainage in the final dispersal areas (Ives, 1959). Derbyshire believed that subglacial channels were first controlled by ice thickness and differential hydrostatic pressure, but as the remnant glaciers thinned they became much crevassed and waterlogged and the topography directly controlled the subglacial drainage.

From the pattern of ice-flow features and eskers (Map 1253A) it is clear that there was no major final dispersal centre, but rather a complex shifting of centres in response to changing climate. The Schefferville region, however, was a major centre of dispersal. This is indicated by the presence of various sets of striae in the region and by the directions of differential isostatic uplift determined from glacial lakes to the south, east, and northeast of this area (Barnett and Peterson, 1964).

A radiocarbon date on gyttja beneath peat in Chibougamau district, Quebec, indicates that pond deposits were formed as early as 6,960 $\pm$ 90 years B.P. Pollen extends through the uppermost 3 feet of silty clay of the glacial lake deposit. It is therefore likely that the ice front had receded from this area prior to about 7,500 years ago. Bog bottom dates from Grand Falls, Hamilton River (Morrison, 1963) and from near north end of Ashuanipi Lake, 100 miles to the west-southwest (Grayson, 1956) range from 5,250 $\pm$ 800 to 5,575 $\pm$ 250 years B.P. In all these sequences palynological evidence indicates a time gap between the base of the peat and the underlying glacial deposits. The dates indicate the initial development of widespread peat deposits and not the time of deglaciation of the area. This region was probably deglaciated prior to 7,000 years ago. A radiocarbon dating on a composite peat sample from near the base of a peat bog near Marymac Lake, 170 miles north-northwest of Schefferville, indicates that lake deposits had given place to peat prior to 6,400 $\pm$ 900 years ago. Grayson estimated the start of bog formation as closer

to 8,000 than 7,000 years ago. Deglaciation of this area between 8,000 and 7,500 years ago is in keeping with incursion of the sea into Ungava Bay about 9,000 years B.P. and into Hudson Bay by 8,000 years B.P.

Keewatin Sector, Laurentide Ice Sheet

The Keewatin sector of the Laurentide Ice Sheet, or simply Keewatin ice sheet, is that part of the continental ice sheet which left a pattern of transverse morainal elements and longitudinal ice-flow features and eskers as it receded towards the Keewatin Ice Divide, an elongate area northwest of Hudson Bay. It covered the whole of western Canada east of the Cordillera, and an adjoining part of the north-central United States, except for a few small areas near its terminus near the International Boundary where the ice sheet was relatively thin. It also covered some of the arctic coastal islands and the southern side of Melville Island. At its maximum it was in contact with Cordilleran ice on the west, and Arctic, Baffin, and Labrador ice on the north and east. Tyrrell (1897) states that the name Keewatin is derived from Cree Indian meaning north or north wind and he used the term as applied to the ice sheet which he believed had its "gathering ground" northwest of Hudson Bay, in District of Keewatin, Northwest Territories. He was aware of the difficulty of establishing a glacier in this inland region of low altitude but nevertheless envisaged an active and shifting centre of outflow operating over a long period of time. The term Keewatin Ice Divide is now applied to a linear zone, almost 500 miles long, lying northwest of Hudson Bay, around which eskers and ice-flow features are arranged in a roughly radial pattern. It represents "the zone occupied by the last glacial remnants rather than a centre of ice dispersal" (Lee, *et al.*, 1957).

Keewatin Ice Divide is an area of few eskers and flow features as compared to adjacent areas (Lee, 1959). Those present indicate that the area of late ice dispersal migrated westward during the last stages of activity, the result of marine transgression into Hudson Bay. Thus, in many places, ice-flow features that were formed by ice moving in opposing directions merge against or into one another. The origin and initial development of Keewatin ice, however, remains conjectural.

Southern Interior Plains

The thickness of glacial drift on the Prairies varies greatly, probably averaging around 200 feet and locally is known to be more than 1,000 feet. Its general character is different from that of glacial drift in eastern Canada. The Prairies include great tracts of hummocky drift (Pls. XII-10, 11) that in places show transverse ridging (Prest, 1968). These features appear to reflect the clayey nature of the Prairie tills which have been derived from dominantly shale strata, and some of which contain much montmorillonite. Major end moraines per se are relatively uncommon and largely a matter of interpretation. Many of the great drift ridges formerly regarded as end moraines of southwestward-moving glaciers are now considered to be drift-mantled bedrock ridges, and for the most part mark marginal rather than end positions of lobate ice fronts. Furthermore, there has been much thrusting and folding of bedrock by glacier pressures, including induced pore-pressure (Slater, 1927; Byers, 1960; Kupsch, 1962) which have made it generally difficult and commonly impossible to differentiate between recessional drift ridges and ice-thrust ridges mantled with drift (Prest, 1968). Both types are included under the transverse ridge symbol on the glacial map (Map 1253A). Great blocks of bedrock, some perhaps measurable in miles, are underlain by till along Oldman River in southern Alberta (Stalker, 1963). L. A. Bayrock reports similar masses east of Edmonton. Sheets of bedrock resting on and capped by till have been encountered in drilling in Saskatchewan, according to E. A. Christiansen. Whitaker (1965) reports similar occurrences in southern Saskatchewan and also refers to very large masses of stratified drift that occur as



PLATE XII-10
Hummocky disintegration moraine near Cypress Hills, Saskatchewan.

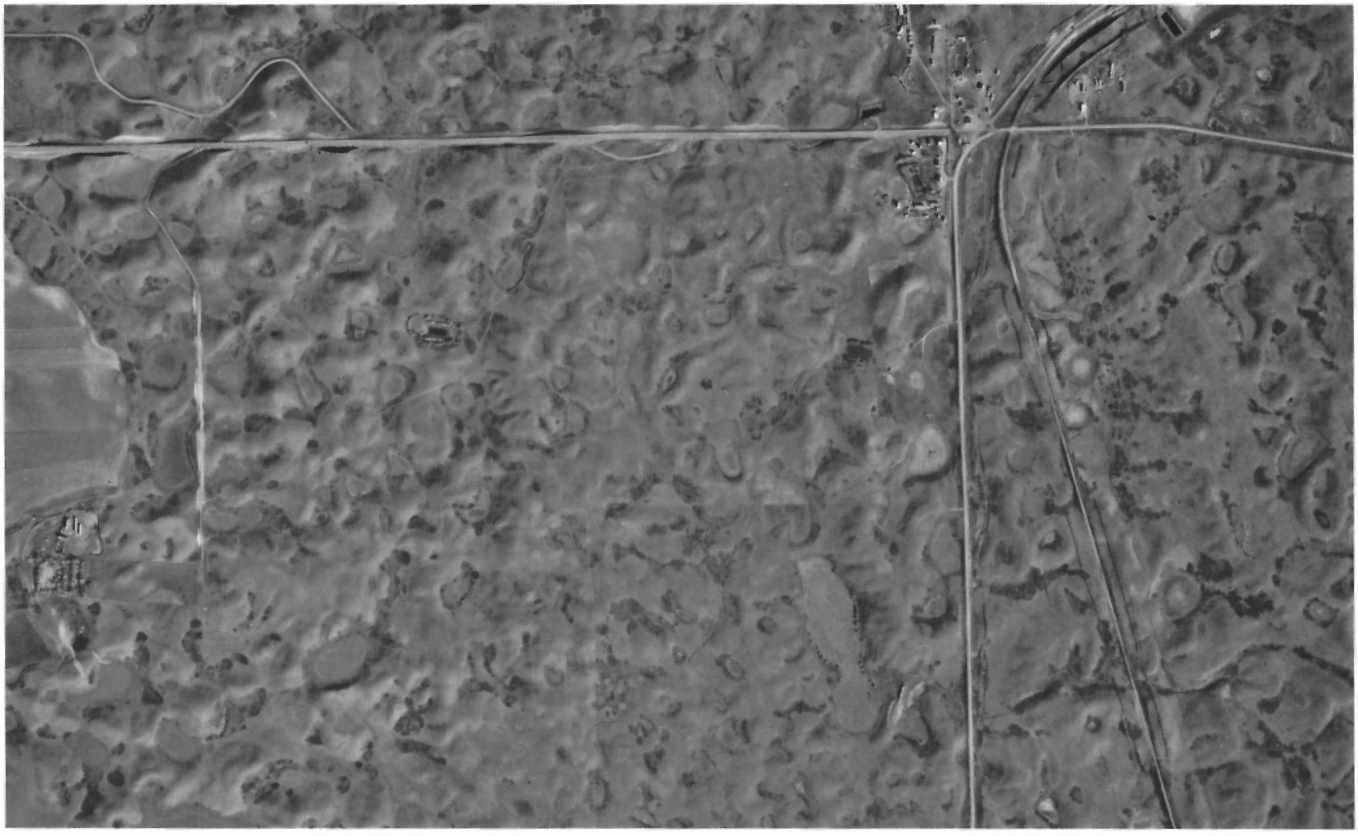


PLATE XII-11. Hummocky disintegration moraine east of Lake Johnstone, southern Saskatchewan. Occurs on western side of ice-thrust moraine of the Coteau moraine complex; local relief is about 80 feet. Town of Crestwynde in upper right corner. Vertical airphoto; scale 1 inch to 1,320 feet.

erratics and presumably were frozen when moved. The obvious thrusting action involved in the emplacement of ice-thrust ridges and discrete sheet-like erratics between tills, coupled with the sliding of debris in hummocky disintegration moraine and consequent inversions of topography, militate against easy development of a late glacial chronology for the Interior Plains. At present it is difficult to document an orderly general pattern of ice retreat despite the large amount of work done.

Ice-flow directions. On the southern Interior Plains the ice-flow and recessional moraine patterns indicate former glacial movements towards the southwest, south, and southeast. Local topographic features were instrumental in directing the late phases of glacier flow. It is also evident that the regional northwest trend of bedrock and the regional slope to the northeast have played important roles in deflecting the southwest flow from the Shield, towards the south and southeast. Southwest-trending ice-flow features in some relatively high areas on the Plains may be related to movements of a southwestward expanding ice sheet, but it is more likely that these features are related to south and southeastward moving glaciers which locally overtopped the valleys containing them. Shield stones indicative of a distant source are present along the length of the Rocky Mountain Foothills, but their provenance has not been determined. At the Wisconsin glacial maximum Keewatin ice was in contact with Cordilleran ice, or overlapped ground vacated by this ice, along the entire length of the Foothills in Alberta, except for a small area in the Porcupine Hills of southwestern Alberta (Douglas, 1950).

In Manitoba the ice-flow pattern indicates that regional glacier movements were to the southeast. Cretaceous rocks from the west side of Red River valley, near the International Boundary, were transported more than 400 miles across Paleozoic and Precambrian terrain and incorporated in till of the Grantsburg sublobe east of Minneapolis, according to H. E. Wright, Jr. (Univ. Minnesota). This indicates long-continued ice flow parallel to the presumed retreat flow pattern. Glacier retreat was mainly towards the northwest, but near Lake Winnipeg basin the influence of the Labrador ice has complicated the ice flow and the age relationships between various patterns have not been established. Along the Manitoba escarpment, the prominent hills and valleys brought about lobations of the late ice fronts and hence locally the ice receded towards the east and northeast.

In southern and central Saskatchewan the ice-flow patterns are exceptionally intricate and the order of late-glacial movements varies greatly from one area to another

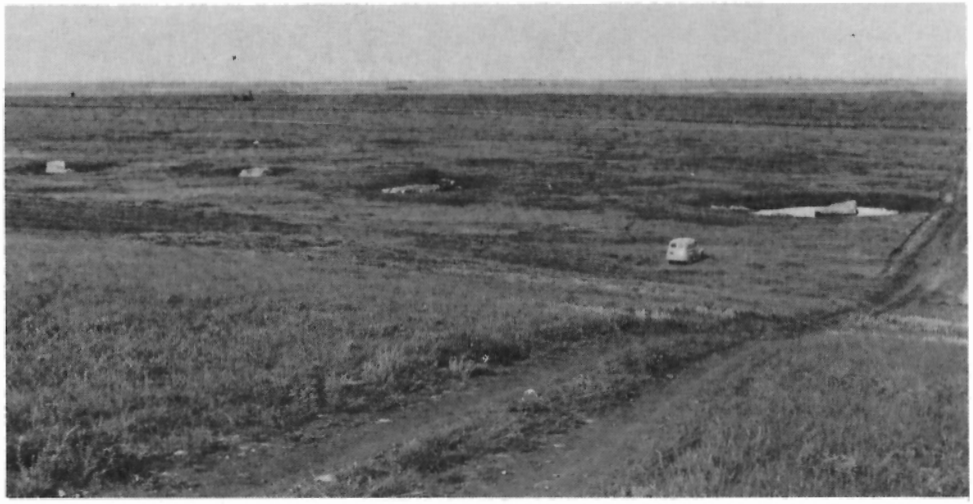


PLATE XII-12
Blocks of pebbly quartzite of erratics train from Athabasca Valley glacier, Claresholm, Alberta. The depressions around the boulders are mainly due to buffalo wallowing. The largest boulder, now broken in three pieces, measured about 50x13x4 feet.

(Parizek, 1964; Christiansen, 1965). There appears to be a regional recession of the ice sheet towards the north-northeast. The major ice-frontal positions, however, may only reflect thinning of the ice sheet and a lowering down the regional northeast slope while the ice flow remained essentially south and southeast along the slope. The longitudinal and transverse glacial lineaments appear to indicate this general trend but with many variations due to local topographic effects.

In southern Alberta the glacial features clearly indicate a predominant south to southeast trend. Keewatin ice must have flowed southward for an extended period for Cordilleran ice from Athabasca River valley merged with it and emplanted immense blocks of pebbly quartzite upon it; these were transported southward as far as Montana (Stalker, 1956; Mountjoy, 1958). These erratics were distributed along a path generally only a few miles wide and more than 400 miles long. They comprise the Foothills erratics train (Pl. XII-12). A. MacS. Stalker believes that this train of erratics represents the limit or near-limit of the Classical Wisconsin ice sheet. The group of quartzite blocks at Okotoks, Alberta, are considered to be parts of a former single block that weighed about 18,000 tons.

Regional ice recession was to the northwest parallel to the regional structure, with minor recessional shifts to the north and northeast occasioned by local high ground and by northeast-trending valleys. In northern Alberta the direction of ice recession gradually shifted to the northeast towards the Shield and hence the ice-marginal positions and moraines trend northwest. There were, however, many lobations and variations in glacier movements during the period of recession that are now revealed by a complex pattern of ice-flow features.

Glacial lakes. Meltwater was commonly ponded along the receding ice margins, but other than glacial Lake Agassiz, already discussed, the lakes were generally ephemeral with poorly marked shorelines. The bottom deposits are generally thin but in some areas are thick enough to completely mask hummocky terrain. Where thick lake sediments were deposited prior to the melting of buried ice blocks a hummocky terrain formed which closely resembles that of disintegration or dead-ice moraine.

In southern Saskatchewan, the ice front along the upper part of Qu'Appelle River ponded meltwater on its south side to form glacial Lake Regina (Johnston and Wickenden, 1930; Christiansen, 1961). This lake at the 1,900 to 1,950 feet level discharged southeastward via the Souris River system into Mississippi River. As ice recession continued into north-central Saskatchewan, the Qu'Appelle River valley served as a spillway for a series of lakes ponded in the Saskatchewan River valleys, discharging into glacial Lake Agassiz. These lakes received meltwater from the ice front and from glacial lakes as far northwest as the Rocky Mountains of east-central Alberta. As ice recession continued, the upper reaches of Assiniboine River in eastern Saskatchewan then served as a spillway into the Agassiz basin. The precise relationship of the various lake stages in Saskatchewan River valley with the glacial lakes in Alberta or with glacial Lake Agassiz have not been established.

In southern Alberta glacial lakes were numerous but small; meltwaters readily escaped to the southeast. Glacial lakes were more extensive in northern Alberta where the ice front plugged the major river valleys that trend northeast, and lateral discharge was difficult. Glacial Lake Edmonton occupied part of the North Saskatchewan River valley near Edmonton and at one stage extended into the Athabasca River valley (Gravenor and Bayrock, 1956; and Taylor, 1958). Deposits in the western end of the basin occur at altitudes up to 3,400 feet, but southeast of Edmonton lie below 2,500 feet. Discharge was to the southeast, at first into Red River and later into Battle River and the Qu'Appelle Valley to Lake Agassiz. A succession of lakes in Smoky River and Little Smoky River valleys west of Athabasca River are described by Henderson (1959a) and St. Onge (1966). The highest lake

phase at 2,800 feet elevation is known as glacial Lake Rycroft. It extended west and northwest along the convex ice front into Peace River valley. Discharge was via the Pass Creek spillway from the southeastern end of the lake across a low divide at the head of Iosegun River into the Athabasca and presumably into a low stage of Lake Edmonton. St. Onge (1966) reports a radiocarbon date on shells pertaining to glacial Lake Rycroft of 12,190 ± 350 years (GSC-508).

By the time the ice front had receded so that discharge could take place into Lesser Slave Lake basin, the ice-marginal lake in the Smoky River valleys had lowered some 800 feet but remained in the Peace River valley and occupied a somewhat larger area than did Lake Rycroft. This phase is known as glacial Lake Fahler. It discharged eastward over a sill 10 miles west of High Prairie at an altitude well below 2,000 feet (Henderson, 1959a) into a lake in the Lesser Slave basin that was confluent with water ponded in the Athabasca River valley. These waters discharged across a sill east of Lac La Biche, at an altitude of 1,850 feet, and thence flowed via Saskatchewan River to glacial Lake Agassiz. The succession of lakes in Peace River valley, including Rycroft, Fahler, and lower levels were termed glacial Lake Peace by Taylor (1958) and Mathews (1963).

The Lake Fahler phase came to an end when the ice front had retreated about 150 miles down Peace River valley and thereby opened lower discharge outlets that led to the Arctic Ocean. According to Henderson (1959a), the Lake Fahler basin above the town of Peace River was drained prior to a re-advance of the ice sheet that returned the lake to its former level with discharge eastward to Lesser Slave Lake basin. Retreat of the ice sheet again resulted in a drainage reversal. Numerous spillways led westward from Peace River valley to Fort Nelson River in British Columbia, permitting northward discharge via Liard and Mackenzie Rivers. Glacial Lake Peace was lowered by a succession of outlets at about 1,800, 1,500, and 1,200 feet. Further ice retreat allowed discharge northward along Hay River into the Mackenzie. This lake phase would appear to correlate with a high level phase of glacial Lake Tyrrell (Taylor, 1958). As the ice sheet receded northeastward and eastward in northern Alberta, this lake expanded into the Lake Athabasca basin and the lower reaches of Athabasca River. When the Slave River valley became ice-free, glacial Lake Tyrrell merged with glacial Lake McConnell which discharged to Mackenzie River from the western end of Great Slave Lake (Craig, 1965a).

Northern Interior Plains and Mainland Arctic Plain

The character and thickness of drift in the northern Interior Plains vary greatly. Craig reports that the till, in the region west of Great Slave Lake, is silty to clayey and stony. Drumlinoid forms are locally very abundant; some of those west of Great Slave Lake are up to 100 feet high and display surface furrowing with a relief of 10 to 15 feet. Hummocky and kame moraines are limited but north of Great Bear Lake hummocky moraine is widespread and there are many end moraines. Mackay (1958) states that the till in Darnley Bay area is rich in silt and clay due to prevalence of carbonate rocks. Eskers, though not abundant, are dispersed widely over the northern plains.

Craig reports drift depths up to 380 feet in the area west of Great Slave Lake and up to 150 feet in river bank sections along Mackenzie River, parallel to the mountain front, but the bedrock is not exposed. Drillholes east of Inuvik on lower Mackenzie River indicate a maximum drift thickness of 230 feet. The drift is probably very thick in the moraines north of Great Bear Lake. Ice-thrust features are common along the Arctic coast (Mackay, 1963).

At the climax of the last or Classical Wisconsin Glaciation, the Keewatin ice west of Great Slave Lake does not appear to have made contact with the Cordilleran Glacier Complex although it reached the mountains west of lower Liard River. There was, rather, a mainly unglaciated zone separating them along the mountain front that was 10 to 30 miles wide and almost 200 miles long extending north from the British Columbia boundary; this zone did, however, contain some valley glaciers. Another zone, slightly larger, occurs west of Great Bear Lake where the Keewatin ice flowed northwestward along the mountain front. These mainly unglaciated areas have been little studied but there is geomorphological evidence along the eastern mountain front that the last ice sheet did not reach as high as an earlier ice sheet. Keewatin ice extended as elongate lobes into some of the valleys along the eastern sides of the unglaciated areas and reached altitudes of above 4,500 feet. These two relatively unglaciated areas may be joined by a similar narrow strip or corridor to the large unglaciated region in northwestern Yukon and Alaska.

Ice-flow directions. West of Great Slave Lake the last ice sheet appears to have met the mountains at almost right angles, but farther north it was deflected parallel to the mountain front and Mackenzie River valley. It also moved parallel to the mountain front over most of the Plains east of Mackenzie River. As the ice sheet thinned and the ice front receded into the Great Slave Lake basin active flow there remained westward with the result that late ice-flow features cut across the trend of some earlier formed features parallel to the regional trend. This late ice was forced to flow around Horn Plateau west of Great Slave Lake and the ice front was, therefore, deeply indented. Elsewhere between Great Slave and Great Bear Lakes a complex pattern of ice-flow features discernible on airphotos suggests some differences in regional flow coupled with local lobations due to topography as deglaciation took place. With continued eastward recession towards the Shield the ice front straightened along the length of the Plains escarpment, some 20 to 30 miles west of the

Canadian Shield edge. Segments of end moraine and some minor moraine occur near Lac la Martre (Craig, 1965a). Craig suggests that this ice front extended southward to Great Slave Lake and beyond. Morainal features, discernible on airphotos of the region north of Lac la Martre, are suggestive of a continuation of the same 'straightened' ice front.

Keewatin ice did not extend beyond the present Mackenzie Delta except west of the river mouth. West of Peel River, the last ice sheet was not so extensive as an earlier ice sheet (Hughes, 1965), but it did extend west along Arctic Coastal Plain as far as Kay Point. Much information on the Pleistocene deposits and history of Mackenzie Delta area is given in a comprehensive report by Mackay (1963).

Between Mackenzie and Anderson Rivers ice-flow features trend north along the strike of the Paleozoic rocks but south of Eskimo Lakes trend northeastward; the reason for this deflection is not known. Mackay (1963) considers the morainic topography on the north side of Eskimo Lakes, in a belt 5 to 10 miles wide, as the terminal zone of a major ice sheet. He also noted smaller end moraines in the Campbell–Sitidgi Lakes area east of Inuvik, and also ice-flow features that trend north-northeast and northwest. He also suggested that the northeastern end of the Tuktoyaktuk Peninsula, which separates the Beaufort Sea from Eskimo Lakes and Liverpool Bay, might be unglaciated and that the limit of the last major ice sheet lay more or less along the middle of the peninsula. Fyles (1967a), however, suggests that the limit of the last ice sheet may have lain along the length of the south side of Tuktoyaktuk Peninsula, and that it may not have reached the northern part of Cape Bathurst.

In general, ice retreat north of Great Bear Lake was towards the southwest, parallel the strike of both Proterozoic and Paleozoic rocks. The complex pattern of ice-flow features, however, indicates that during the late stages of deglaciation the ice margin was highly lobate. Near the Arctic coast the late ice-frontal lobations were further complicated by Amundsen Gulf lobe (Mackay, 1958). The interplay of this ice lobe with that north of Great Bear Lake, and of local draw-down effects in vicinity of Great Bear Lake basin are described by Craig (1960).

Glacial lakes. As Keewatin ice receded from the northern Interior Plains meltwaters occupied the major depressions of Great Bear and Great Slave Lakes, and flowed northward down Mackenzie River. When the ice front lay along the edge of the Canadian Shield a vast lake extended from Great Bear Lake basin southward through Great Slave Lake basin to the lower Athabasca and Peace River valleys. Craig (1965a) has named this confluent lake phase glacial Lake McConnell. As isostatic uplift drained the lake to lower levels it separated into smaller lakes ancestral to the present Great Bear and Great Slave Lakes. In the latter lake basin strandlines are warped upward to the east. Craig suggests a minimum uplift rate of about 2 feet per mile in Great Slave Lake basin based on the presence of a delta at 500 feet elevation at Fort Simpson and of strandlines up to about 925 feet near the Canadian Shield edge. Great Bear Lake became separated from Great Slave Lake when the water levels fell below about 750 feet, the elevation of the present drainage divide between them.

Pingos and ground ice. Postglacial features of great interest in the lower part of Mackenzie Delta and in Eskimo Lakes–Liverpool Bay region are the pingos or conical ice-cored hills (Porsild, 1938). Mackay (1962, 1963) states that the pingos are elliptical to oval, 100 to 2,000 feet across, 10 to 150 feet high and generally asymmetrical. They occur in former lake basins or drainage channel depressions. The ice cores are believed to roughly conform to the shape of the pingo but perhaps have steeper slopes, and to bottom close to the surrounding ground level. The ice is overlain by a mantle of sand capped by vegetal matter, and is underlain by sand similar to that of the surrounding depression areas. The tops of the pingos are frequently breached or broken. Pingos with diameters of more than about 600 feet do not stand as high as the smaller ones, and those with the greatest base are merely bulge-like swellings in the lake basin sediments. Pingos result from a squeezing process as permafrost encroaches upon unfrozen saturated sands beneath what were formerly relatively large and deep lakes. They may exist for hundreds to thousands of years.

Great lenses of ground ice also constitute important elements of the landforms of this northern region according to Mackay. Fyles (1966) stresses the difficulties of stratigraphic analyses where such ice wedges are prevalent.

Arctic Lowlands

This part of the Arctic Archipelago lies between the mainland and the Queen Elizabeth Islands and includes the northern limit of Laurentide Ice Sheet. The drift generally reflects the variable character of the underlying Proterozoic and Paleozoic bedrocks.

Large morainal belts with fresh glacial landforms characterize an eastern part of Banks Island and are related to the Classical Wisconsin Glaciation (Fyles, 1962). Drift is varied and generally heaped into steep-sided hills but in places occurs as broad, smooth, hills and ridges, with intervening gentle depressions. Relief is several hundred feet. Marine beaches are locally well developed.

The last continental ice sheet reached the south coast of Melville Island perhaps as recently as about 11,000 years ago, and was responsible for the Winter Harbour Moraine (Fyles, 1967b). The moraine is composed of bouldery drift and stratified deposits and comprises a belt of hilly and ridged topography 50 miles long and averaging 2 miles wide. It is less extensive than a former glaciation.

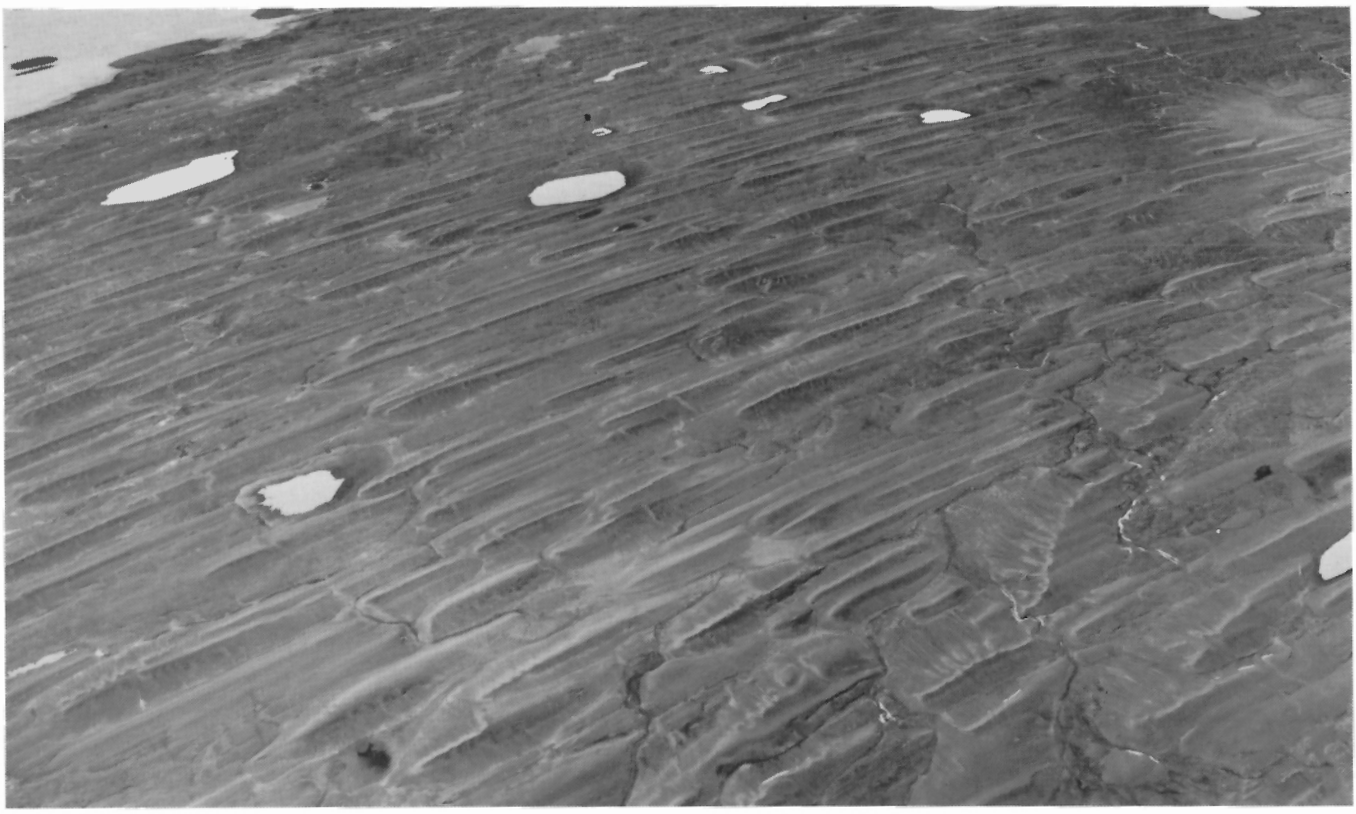


PLATE XII-13. Drumlin belt, Stefansson Island, Northwest Territories. View to the west. Ice flowed northward into Viscount Melville Sound.

Fyles (1963a) reports that the drift on Victoria Island characteristically has a light coloured matrix of slightly sticky, dense, loam or sandy loam. Numerous and exceedingly variable linear landforms are built of glacial debris or glacially deformed strata. These belts include landforms that are generally ascribed to end, kame, and hummocky moraine. The moraine ridges range from major features several hundred feet high, traceable more or less continuously for tens of miles, to miniature forms only a few feet high and a mile or so long. The complexities and interrelations of these great areas of hummocky and ridged moraine make it difficult to delineate end moraines per se. The drift is generally a few tens of feet to more than 100 feet thick and locally exceeds 500 feet. Drumlinized forms are characteristic of most lowland areas (Pl. XII-13) and there the drift is only a few tens of feet thick with bedrock protruding in some places. Eskers are numerous in the low-lying eastern and southern parts of Victoria Island, but they are few and short elsewhere; the longest is 120 miles. The eskers are usually less than 150 feet high but some knots at the junction of two or more esker-tributaries are several hundred feet high. Most consist of sand and gravel but some have material similar to that of the moraines.

An end moraine ridge and kame hills occur on northwestern Prince of Wales Island forming a linear zone of thick drift (Craig, 1964a). Eskers are rare on Prince of Wales and Somerset Islands and on Boothia Peninsula but are numerous on King William Island. They generally parallel the ice-flow direction, but on Prince of Wales Island some are transverse. Kames are striking features of central Boothia Peninsula (a Canadian Shield promontory within the Arctic Lowlands) and western Prince of Wales Island. They occur as broad zones that appear to be ice frontal and also as individual hills and groups of hills and ridges. They range from a few tens to about 200 feet high, and are rarely as much as 400 feet above the general ground level. Glacial landforms are uncommon on the northern parts of Prince of Wales Island and especially so on Somerset Island. These Paleozoic uplands are monotonously featureless but a thin layer of silty, rubbly glacial till occurs in many places, and some ground moraine topography, isolated kames, and meltwater channels are present.

Ice-flow directions. Ice-flow features and transverse morainal elements in the Arctic Lowlands are very complex, but nevertheless the maximum extent of the Wisconsin ice sheet can be delineated in a general way. The Amundsen Gulf ice lobe crossed Darnley Peninsula and reached the eastern side of Franklin Bay but does not appear to have crossed Cape Bathurst. It glaciated at least part of southern Banks Island and the east coast of Banks Island (Craig and Fyles, 1960). J. G. Fyles believes it may have extended westward along M'Clure Strait to the western end of Banks Island. Keewatin ice reached some distance inland on southern Melville Island, and farther

east crossed Somerset Island and Boothia Peninsula (Fraser, 1957; Craig, 1964a). The ice sheet, at the glacial maximum, probably flowed northwesterly and northerly across Prince of Wales and Somerset Islands but later as it thinned and receded the deep channels between the islands influenced the ice flow. On southern Prince of Wales Island the last ice flow was eastward across the north-trending drumlin fields left by the earlier regional flow. A late ice lobe in Peel Sound may account for easterly trending striae on the west side of Somerset Island. Northwest- and northeast-trending drumlin fields occur on parts of King William Island. Craig (1964a) believes that the last ice movements over King William Island and Boothia Peninsula were northeastward.

On Victoria Island the regional flow was westward and northwestward but successive changes of late flow occurred as the ice sheet thinned (Fyles, 1963a). In places diverse drumlinoid trends occur side by side or are superimposed one upon the other. Flow directions are also controlled by relatively minor topographical features, suggesting that the ice sheet was lobate and thin. As the ice thinned, it withdrew from the higher parts of western Victoria Island, retreating to the southeast. Active ice tongues, fed by the main ice sheet to the south and east, remained in the depressions now occupied by the various arms of the sea. The moraines of western Victoria Island are lateral to these ice tongues. In the east-central lowland the ice probably flowed northwesterly in the early stages of wastage but with continued wastage, which took place progressively from west to east and north to south, the ice covering the lowland became thinner, and the direction of ice flow was deflected to an ever increasing degree by minor topographic irregularities. Glacier lobes in the low areas produced fan-shaped areas of ice-flow features trending northward and southwestward.

In the southeastern part of the island the morainal ridges are smaller, lower, and shorter, compared to the great morainal complexes of the western parts. This part of the island was the last to be deglaciated. Strongly flowing ice from the mainland constituting the receding Amundsen Gulf lobe passed down Bathurst Inlet and crossed Coronation Gulf, building a fan-shaped drumlin field across the southwestern end of the island. Apparently the last ice on the island covered the low ground bordering the east and southeast coasts while ice tongues still occupied M'Clintock Channel and Queen Maud Gulf (Fyles, 1963a).

Western Canadian Shield

In Canadian Shield areas west and north of Hudson Bay the character of the drift and the glacial features themselves are rather unlike those of the plains and lowlands. The till matrix is generally silty to sandy and lacks

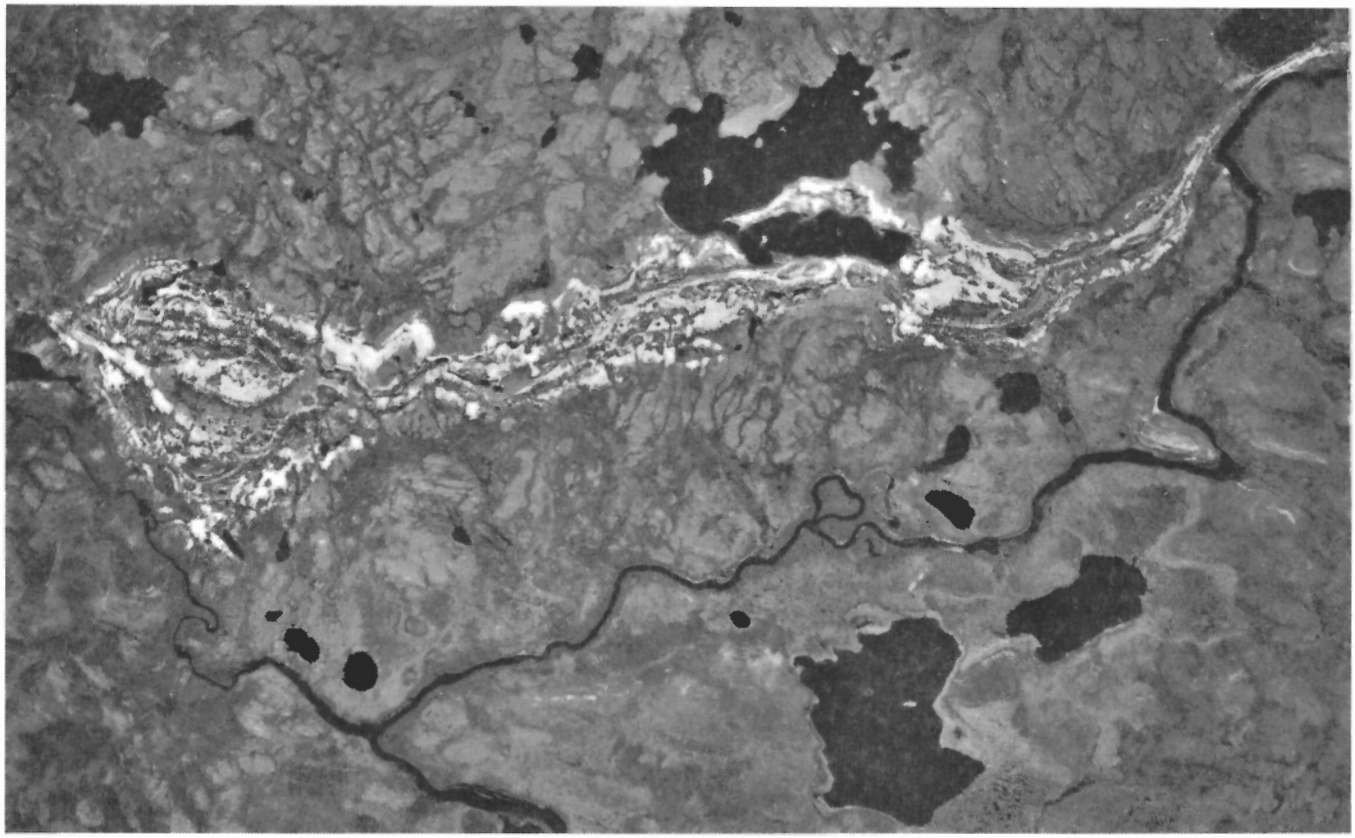


PLATE XII-14. Esker complex west of Ennadai Lake, Northwest Territories. Meltwater flowed southward. The large esker 'knot' is the south end of a system that is more or less continuous northeastward for about 60 miles to Kazan River. Vertical airphoto; scale 1 inch to 3,300 feet.

cohesive properties. The western part of the Shield is commonly mantled with drift in the form of ground moraine and ribbed moraine, which has a somewhat hummocky appearance but is unlike the hummocky terrain of the disintegration or dead-ice moraine of the plains. Eskers are widespread and numerous; they include the simple, sinuous ridges and the great complex multiple systems with widths of a mile or more (Pl. XII-14). The eskers and the remarkably continuous belts of drumlins, drumlinoid features, and glacial flutings delineate ice-flow directions during recession of Keewatin ice. End moraines are not common, but where present are clear linear features transverse to the ice-flow direction. Relief is commonly several tens of feet to about a hundred feet but locally kames, esker knots, and moraine knobs stand several hundred feet above the surrounding terrain.

Ice-flow directions. As Keewatin ice receded from the southern Interior Plains onto the Shield the ice front appears to have formed subparallel to the Shield edge and ice flow was roughly radial to this boundary. A halt in northern Saskatchewan is marked by the Cree Lake Moraine (Sproule, 1939; Tremblay, 1961a, b). It has been traced by means of airphotos northwestward to the western end of Lake Athabasca and eastward into Manitoba for about 500 miles. It varies from an end moraine-outwash complex to a single discrete ridge or a series of minor ridges. Thereafter, retreat towards Keewatin Ice Divide appears to have been uninterrupted by other major halts or re-advances.

In the Great Slave and Great Bear Lakes region recession was also continuous and no major end moraines were formed. The most significant result of this long period of retreat was the development of a marked discordance in the pattern of eskers and ice-flow features between two major ice lobes, a discontinuity that persisted as the ice sheet receded towards Keewatin Ice Divide (Fyles, 1955; Craig, 1957; Craig and Fyles, 1960; Blake, 1963; and Craig, 1964b). The discontinuity in the general radial pattern of flow features suggests that two distinct major glacier lobes formed during recession of the ice sheet. The break in flow pattern may readily be traced from the west side of Dubawnt Lake west-northwest to a point north of Aylmer Lake, and then curving northward past Contwoyto Lake to Coronation Gulf and beyond. To the north and east the flow features lie at a sharp angle to this 'break', whereas to the south and west they tend to be parallel or diverge only slightly from it. The line of demarcation is also revealed by a complex esker system between Dubawnt and Aylmer Lakes which lies a few miles south of the break. This esker continues for about 225 miles and except near the western end all tributary eskers lie only on the south side. Beyond Aylmer Lake a series of closely spaced eskers mark the continuation of the esker system towards Great Bear Lake (Craig, 1964b).

Craig and Fyles (1960) suggest that topographic control was responsible for developing the discontinuity. South and west of the break the ice was covering higher ground, moving slowly and stagnating locally, whereas the ice on the lower ground north of it was actively flowing. Blake (1963) also invokes a difference in activity between two lobes to account for the discontinuity northeast of Contwoyto Lake. Craig indicates that some northward-trending eskers north of it were overrun nearly at right angles by the last-formed lobe of the northern glacier. Southwest- and south-trending drumlinoid ridges west of Dubawnt Lake on the north side of the discontinuity are at right angles to the trend of ice-flow features on either side. These features are in part parallel to older striae recorded by Tyrrell (1897), but Craig considers them related to lobation within the northern part of the ice sheet as it receded towards Keewatin Ice Divide.

Ice retreat from the region north and east of the discontinuity was marked by the formation of end moraines. Two of the moraines, each about 25 miles long, lie east-northeast of Contwoyto Lake and trend slightly west of north. The westernmost moraine, with a maximum relief of nearly 400 feet, lies close to the discontinuity and apparently represents the terminal position of ice moving southwest (Blake, 1963). The eastern moraine is a recessional moraine. Several segments of east-west trending end moraine are evident southwest of Queen Maud Gulf. A prominent moraine extends northeastward from Back River, at the 106th meridian, to MacAlpine Lake for a total, including breaks, of more than 200 miles. This moraine varies from a few tens of feet to 4 miles wide, and from 10 to 250 feet high. The ice front apparently curved eastward to Chantrey Inlet (Blake, 1963; Craig, 1961) and was responsible for two more segments of end moraine some 60 and 25 miles long. Many segments of end moraine that may correlate with that west of Chantrey Inlet extend eastward to Committee Bay for another 225 miles. In this latter area, however, there are several discordant segments and correlation of ice-frontal positions is uncertain. Correlation with end moraines farther east, beyond the sphere of influence of Keewatin ice, is even more tenuous. A younger end moraine of the northern lobe of Keewatin ice also extends northeastward from Back River at the 106th meridian for a distance of 100 miles.

The position of the extension of the ice fronts south of the discontinuity are uncertain, but according to Craig (1964b, Fig. 7) there may be a significant jog eastward. Direct correlation of moraines across the discontinuity has yet to be established.

Glacial lakes. As Keewatin ice receded eastward across the Shield the east and northeast drainage systems became ice free and meltwaters were ponded to form several extensive glacial lakes (Map 1253A). A large lake occupied the Lake Athabasca basin and spilled northward over the rock sill in Slave River at an altitude of 700 feet (Craig, 1965a).

Somewhat earlier a glacial lake formed behind the Cree Lake Moraine in the Cree Lake and Pipestone (Cree)

River valley, the strandlines lying at about 1,800 feet. It may have drained southward through a break in the moraine and then discharged westward down the deep Clearwater River channel to Athabasca River valley. There is, however, an outlet channel from the southwestern end of the glacial lake basin, through which discharge took place into a lake in Richardson River valley at an elevation of about 1,400 feet; this was probably an eastern part of glacial Lake Tyrrell. As the ice front retreated northeast down Pipestone River the glacial lake in that valley discharged laterally to Lake Athabasca basin and successively lower levels were established. When the receding ice fronts cleared the eastern end of Lake Athabasca basin the two water systems merged at about 1,000 feet elevation. During this interval of retreat, a glacial lake formed in the Wollaston Lake basin at a maximum altitude of about 1,600 feet and discharged southeastward into Reindeer Lake basin. Later, lower outlets to Pipestone River valley were established. Reindeer Lake basin was occupied by meltwater when the ice front receded from the moraine at its south end. The outlet through the moraine was at about 1,200 feet a.s.l. and led directly into glacial Lake Agassiz lying south of the moraine.

Extensive glacial lakes also formed in Northwest Territories. Craig (1964b) found evidence of short-lived glacial lakes in the Artillery Lake–Lockhart River basin, east of Great Slave Lake. Beaches occur up to about 1,300 feet around Clinton and Holden Lakes. As the major southern lobe of Keewatin ice receded eastward from this area water was ponded in the headwaters of the northeast-sloping Thelon and Dubawnt Rivers drainage systems. The highest lake phases were about 1,250 feet and discharge was probably southward. Lower lake levels resulted as the ice receded eastward. Retreat of the major northern lobe of Keewatin ice allowed the water in the Thelon basin to discharge northward to the Back River system and lake levels were established at about 800 and 700 feet. At about this time Hyper–Dubawnt was at an altitude of 900 feet. The above lake phases are figured and described by Craig.

South of Dubawnt Lake, glacial Lake Kazan (Lee, 1959) occupied part of Kazan River and Ennadai, Kasba, and other lake basins. It drained eastward towards South Henik Lake and thence presumably into the sea in Hudson Bay. The strandlines around Ennadai Lake (1,070 feet elevation) are at an altitude of 1,260 feet (Pl. XII-15). Later, as the ice melted from Keewatin Ice Divide the lake discharged via Kazan River.

Marine overlap. At the same time as some of the above-mentioned lakes were present the sea was encroaching along the central-north coast (as early as 10,200 years

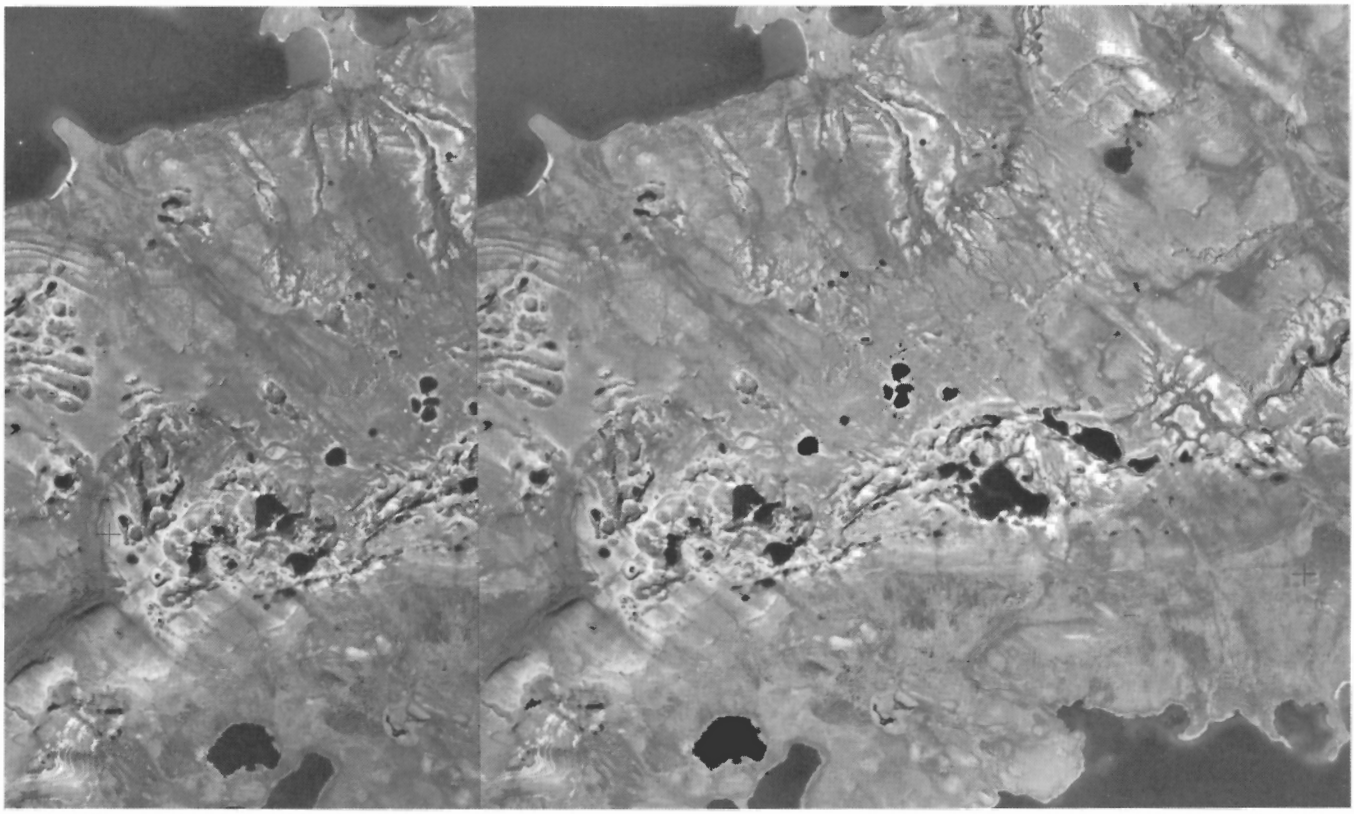


PLATE XII-15. Fluted and kettled ice-contact deposits with raised beaches and frost polygons, Ennadai Lake, Northwest Territories. Ice-contact deposits, 50 miles west of south end Keewatin Ice Divide, were fluted by an ice advance and later terraced and channelled by meltwater. Glacial Lake Kazan washed the lower slopes of the deposits. Some of the buried ice melted during and after the glacial lake phase. Stereoscopic pair; scale 1 inch to 3,100 feet.

B.P., Craig and Fyles, 1960). The sea was also in contact with a part of the ice front as the end moraine was formed at MacAlpine Lake only 8,200 years ago. By means of radiocarbon dating of shells from close to marine limit, Blake (1963) has estimated that the ice retreated about 175 miles from Kent Peninsula to MacAlpine Lake in about 1,000 years. Retreat towards Keewatin Ice Divide was also rapid, for the sea had encroached from Hudson Bay to near its maximum limit in the divide area by 7,200 years B.P. (Lee, 1960). It is clear also that isostatic rebound was rapid during this period of retreat for there is a general decrease in elevation of the marine limit with proximity to the divide. The rate of land emergence has decreased exponentially to the present (Lee, 1960; Craig, 1961). For the Bathurst Inlet area, Blake (1963) suggested an average rate of 1.5 feet per century over the past 2,000 years and probably less at present.

The marine limit west of Hudson Bay appears to decrease from south to north; it is about 650 feet in northern Manitoba, only 480 feet at Kaminuriak Lake, and 400 feet at Wager Bay. Along the Arctic coast the marine limit is greater, being over 650 feet west of Committee Bay and decreasing westward to 500 feet at Chantrey Inlet. It then increases westward to 750 feet at the southern end of Bathurst Inlet and drops off rapidly northwestward to 200 feet in Darnley Bay. The marine limit is not synchronous everywhere and it is probable that more than one centre of postglacial uplift has been active in this vast region.

Foxe–Baffin Glacier Complex

The glacial and deglacial history of Baffin Island is rather complex as a consequence of its physiography, latitude, and climate; it involved a major sector of Laurentide Ice Sheet, several local ice caps, and cirque and valley glaciers. The events in some parts of the present land areas were unlike those in others even where contiguous. On Meta Incognita (formerly Kingait) Peninsula, along the south coast of Frobisher Bay, Mercer (1956) found that unmodified cirques were common near the end of the peninsula. Along the central part of the bay the cirques showed evidence of later modification by ice from the interior, but towards the head of the bay they were absent entirely because that region was covered by interior ice prior to the interval of cirque glaciation. He also showed that cirque glaciers did not form anywhere along Frobisher Bay shore in late Wisconsin time as the interior ice receded. He considered that the interval of cirque formation was long and that it occurred during the Wisconsin. Bird (1963), however, favours development during an interglacial period. Some cirque glaciation, however, did take place in the mountainous terrain along the coast north of Cumberland Sound in late Wisconsin time. Thus despite sound field observation in several areas, any attempt to generalize for the entire island must be made with caution.

It is generally agreed that there was an ice dome over Foxe Basin at some stage during the last major glaciation but there are two views concerning its age: (1) that it existed at about the time of the last glacial maximum; and (2) that it only existed during a later stage of deglaciation. In either, the ice dome was concomitant with many local ice caps as well as cirque and valley glaciers along the eastern and northern highland areas (Ives and Andrews, 1963; Bird, 1963; Andrews and Sim, 1964; Falconer, *et al.*, 1965; Andrews, 1966). The relationships between the Foxe Basin ice dome and other sectors of Laurentide Ice Sheet are not well known. The writer considers that the ice dome was a dispersal centre for a substantial part of Classical Wisconsin time. It was probably confluent with Keewatin and Labrador sectors of Laurentide ice on the west, southwest, and south during a large part of the Wisconsin but maintained its own sphere of influence throughout. Ice from the local ice caps on Brodeur and Borden Peninsulas, together with huge valley glaciers flowing northward along Admiralty and Navy Board Inlets, probably merged with ice from Devon Ice Cap in Lancaster Sound at the last glacial maximum.

Glaciation

It was formerly held that initiation of glaciers on Baffin Island took place in the eastern mountain rim and that migration of the ice towards the interior resulted in piedmont glaciers that later expanded into ice caps and then merged to form an ice sheet. As the mountain rim is the eroded eastern edge of a westerly tilted upland it is unlikely that the mountain glaciers could contribute materially to the formation of the interior ice sheet; rather the ice probably accumulated on the larger central tract of the upland surface. As the westward slope of the upland is gentle and its valleys shallow, the ice must have flowed westward over a broad front, presumably as a plateau ice cap which gradually expanded and its margin moved progressively westward (Mercer, 1956; Ives and Andrews, 1963). Glacierization resulted from depression of the snowline below the level of the uplands as the Wisconsin climatic regimen changed. Rapid, large-scale development of the ice cap on the uplands starved the smaller glaciers in the coastal mountains so that the mountain ice caps and glaciers were never much more extensive than they are today. Perhaps the Penny Ice Cap is an exception. Penny Highlands are a remnant of a dissected older upland higher than the main Baffin surface (Bird, 1959). J. D. Ives believes that, because of its proximity to Labrador Sea and the moisture-laden winds moving up Davis Strait, Penny Highlands may have developed an extensive ice cap earlier than other parts of Baffin Island.

The early-formed interior Baffin Island ice caps presumably merged into an ice sheet which expanded differentially westward and fed ice into Foxe Basin. Glacierization of Melville Peninsula may have contributed eastward-flowing ice to Foxe Basin, and Laurentide ice may have flowed northward into the basin, though there is no evidence of these events. Laurentide ice, however, must

have prevented or hindered early southward ice flow from Foxe Basin which therefore filled and later became a dispersal area (Ives and Andrews, 1963, Fig. 19). An ice dome over Foxe Basin is indicated by the tilting of marine strandlines in central Baffin Island towards the basin (Andrews, 1966). Analysis of carbonate content of drift in many places around Foxe Basin indicates outward flow from the basin (Andrews and Sim, 1964) though sporadic distribution of Precambrian crystalline limestone in central Baffin Island is a possible source of some carbonate. Blake (1966) did not find Paleozoic erratics or any increase in carbonate content of drift east of Amadjuak Lake in southeastern Baffin Island. There is, however, good evidence of early radial flow from Foxe Basin across the narrow central part of Baffin Island (Ives, 1962, 1964; Ives and Andrews, 1963; Andrews, 1963). There the ice sheet reached elevations of over 3,000 feet and escaped eastward through major valleys in the mountain rim and onto the continental foreland. Paleozoic erratics found north of Barnes Ice Cap were probably transported northeastward from Foxe Basin and survived through the later short period of outflow from proto-Barnes Ice Cap (Ives and Andrews, 1963, Fig. 22).

Foxe Basin ice also flowed eastward across Baffin Island through low ground to Cumberland Sound and probably southeastward to Frobisher Bay. It flowed southward across Foxe Peninsula and possibly also across part of Southampton Island (Bird, 1953). Strong and probably long-continued westward flow took place across Melville Peninsula where the bedrock is strongly scoured and *roches moutonnées* are common (Blackadar, 1958; Sim, 1960a; Craig, 1965b). There is no direct evidence of northward flow out of Foxe Basin but it may perhaps be inferred from evidence of northeast ice movements east of Steensby Inlet. It is of course possible that substantial ice caps on northern Baffin Island inhibited or prevented northward flow during the Wisconsin Glaciation.

At the glacial maximum some local ice caps occupied parts of the eastern and northern 'mountain' belt and also Brodeur and Borden Peninsulas, and may have prevented the incursion of interior ice to adjacent parts of the higher ground, and protected parts of the coastal foreland from the ravages of glaciers issuing along the fiords. Thus small parts of the eastern coastal lowland as well as some nunataks remained unglaciated at the last glacial maximum (Løken, 1966).

Deglaciation

Following the last glacial maximum, the Clyde phase of Ives and Andrews (1963), there was a period of glacier thinning and ice-marginal recession. This was followed by still-stands and short re-advances comprising the Cockburn phase of glaciation. The Cockburn Moraine System of end and lateral moraines marks the eastern ice-frontal positions of the interior ice sheet. In northeastern Baffin Island the outermost end moraines of the system lie on the high interfiord mountain flanks and descend into the fiords as lateral moraines; the inner moraines lie mainly on the upland plateau (Falconer, *et al.*, 1965). The ice was 1,500 to 2,000 feet thick in the heads of the fiords at this stage. Radiocarbon datings on shells from close to the marine limit established upon withdrawal of the ice from the coastal areas suggest that some of the eastern moraines formed about 8,200 years ago. The marine limit varies from about 220 feet elevation at some fiord heads to about 50 feet at the outer coast. The inner moraines on the plateau must be somewhat younger. Blake (1966) reported that a major end moraine in southeastern Baffin Island, believed a part of the Cockburn Moraine System, was forming for several hundred years after 8,200 years B.P. and suggested that a major end moraine on Foxe Peninsula was another correlative moraine. The Cockburn Moraines extend across northern Baffin Island and along the west side of Melville Peninsula (Falconer, *et al.*, 1965). Parts of this moraine in northern and northwestern Baffin Island may have formed somewhat earlier, perhaps as much as 9,000 years ago. The Cockburn Moraine System therefore appears to occupy a range in time from about 9,000 years to less than 8,000 years ago. This must have been a period of waning ice and consequent rapid change in ice-dome configuration. According to J. T. Andrews the east coast mountain glaciers were less extensive during the Cockburn phase of glaciation than they are at present. The Penny Ice Cap and several other major bodies were separated from the interior ice during this period.

Following the Cockburn phase of glaciation, the Foxe Basin ice dome underwent rapid disintegration and the sea invaded Foxe Basin about 7,500 to 7,000 years ago. The centre of ice dispersal was thus shifted northeastward onto Baffin Island causing ice flow towards Foxe Basin, whereas it had earlier been away from the basin. On the northeastern side of Baffin Island the ice margin receded west of the drainage divide and glacial lakes formed. Along the western side of central Baffin Island glaciers now occupied the fiords and inner bays. Marine overlap left strandlines ranging from about 350 feet near the coast to 100 feet farther inland, while glaciers in the valleys remained in contact with the sea. An end moraine in lower Isortoq River valley is associated with a sea level coincident with the marine limit (Andrews, 1966). Andrews equated the marine limit 'still-stand' with an increase in ice load on Baffin Island and a concomitant reduction in isostatic recovery. He named this period of thickening ice and moraine formation the Isortoq phase of glaciation. It would appear to correspond with the proto-Barnes Ice Cap phase of glaciation (Ives and Andrews, 1963, Fig. 22). Radiocarbon dates indicate that it occurred less than 7,000 but more than 5,500 years B.P. Andrews found that the marine limit was about 230 feet elevation some 15 miles from the mouth of Isortoq River and about 300 feet at the mouth, indicating a differential upwarp in direction S35°W towards the centre of Foxe Basin, at a rate of about 5.3 feet per mile.

The ice receded from Foxe Basin coast west of Net-

tilling and Amadjuak Lakes by 6,700 years B.P. and the marine limit was established west of these lakes at about 350 feet. Ice apparently remained in Amadjuak Lake basin during the Isortoq phase of glaciation; it was gone before 4,500 years B.P., and possibly as early as 5,500 B.P.

As the proto-Barnes Ice Cap shrank, glacial lakes remained along its northeastern side and continued to discharge into Baffin Bay. Glacial lakes were ponded in the middle Isortoq River valley, north and northwest of Barnes Ice Cap, during the period from about 5,000 to 2,000 years ago. As recently as 1,000 years ago a minor expansion of the ice cap blocked the river again and gave rise to another glacial lake, which drained in relatively recent times as the ice cap assumed its present position, size, and shape. The southwestern edge of the present ice cap with recent end and marginal moraines is shown in Plate XII-16. The east coast mountain glaciers are believed to have expanded in the last 2,000 years, reaching their maxima in the last few centuries (Harrison, 1964). Within the last 200 to 400 years a large part of the interior uplands was covered with thin ice and snowfields, but this has successively melted off until today only 2 per cent is covered, exclusive of Barnes Ice Cap (Ives, 1962). Falconer (1966) showed that in northern Baffin Island a small ice cap of about 3 square miles and never more than about 200 feet thick, developed as recently as 300 years ago in an area of patterned ground first deglaciated about 1,000 years ago. This small ice cap was inactive and served to protect the underlying polygonal structures and their vegetal cover—now emerging as the ice recedes. Deglaciation is still in progress and hence the term postglacial is inappropriate on Baffin Island.

In northernmost Baffin Island, Craig (1965b) reports that the glacial landforms and nearly all the surficial deposits are due to the last or Classical Wisconsin Glaciation. There is, however, a dearth of glacial landforms particularly on the Paleozoic rocks of Jones–Lancaster Plateau. Erratics are present throughout the region but are scarce on the northern part of central Brodeur Peninsula. Linear morainal zones with a hummocky surface and poorly defined higher ridges occur in a few places and although some mark terminal positions others afford little evidence of the position and movement of the ice that formed them. The most extensive linear zone is along south side of Bernier–Berlinguet Inlet, where it has been traced for 90 miles and in places rises 200 feet above adjacent land. The glacier appears to have moved westward and expanded radially. On the north side of the inlet there is also evidence of later ice flow from an ice cap to the north-northeast. Glaciers may have flowed eastward into Admiralty Inlet and northward along the inlet according to Craig. Also glaciers may have flowed westward off Bylot

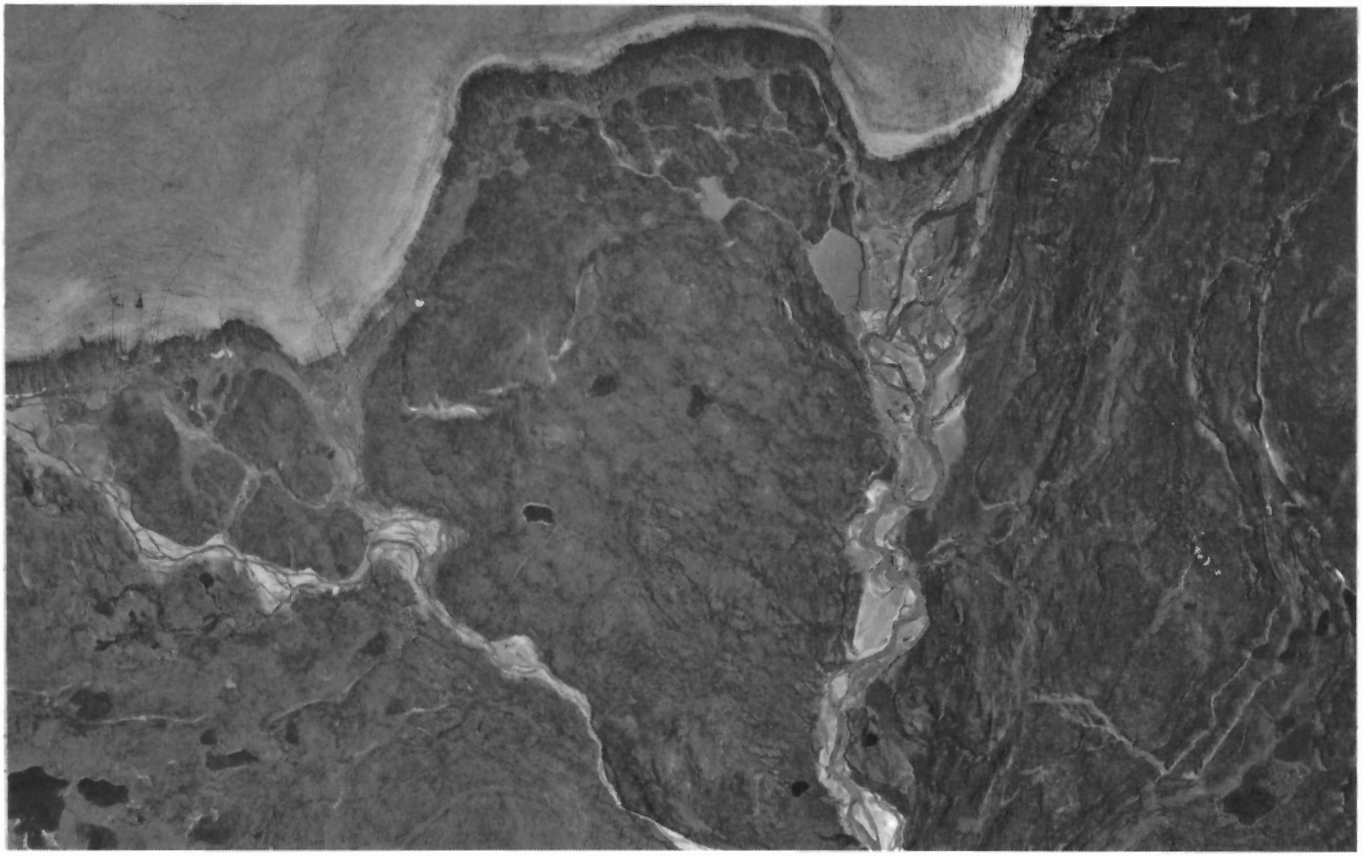


PLATE XII-16. Barnes Ice Cap with recent marginal and end moraines, Baffin Island, Northwest Territories. Vertical airphoto of upper reaches MacDonald River, southwest side Barnes Ice Cap. Scale 1 inch to 4,700 feet.

Island. Large areas of hummocky moraine are found in northern Baffin. These seldom afford any evidence of former ice positions or movement but along the west side of Brodeur Peninsula were probably formed by glaciers from the plateau. Radiocarbon dates on marine shells from raised beaches in northernmost Baffin Island indicate that maximum marine overlap occurred between about 9,000 and 7,000 years B.P.

As the above events were taking place on Baffin Island a somewhat comparable sequence of events occurred on Melville Peninsula (Sim, 1960; Ives and Andrews, 1963; Falconer, *et al.*, 1965; Craig, 1965b). Following the period of regional northwest ice movements over the peninsula, probably during both the Clyde and Cockburn phases, there was extensive thinning of the glacier complex and later incursion of the sea into western Foxe Basin. Sim reports flooding of the east coast of Melville Peninsula to about 500 feet a.s.l., and Craig reported marine overlap of the southeast coast to 485 feet. An ice cap probably occupied most of interior Melville Peninsula before breaking up into two or three local ice caps. Radial outflow was probably maintained by the larger southern body. In the centre of the peninsula a myriad of meltwater channels indicate a former small ice cap, and farther northeast eskers suggest a southeasterly flow of meltwater from another ice remnant (Sim, 1960; Craig, 1965b).

Deglacial events in northern Southampton Island, which may have been covered in part by Foxe–Baffin Glacier Complex, are little known. Bird, however, reports a late ice cap in the central part of northeast coast from which radial flow took place.

Queen Elizabeth Islands Glacier Complex

At present ice fields and valley glaciers are common features of the higher eastern part of the Queen Elizabeth Islands and small ice caps occur on Meighen and Melville Islands. Not all are remnants of the former more extensive ice complex; some have developed in recent times. Relatively fresh glacial features are generally lacking, possibly due to the inactive character of the cold arctic glaciers as compared to temperate-climate glaciers, but mainly to the character of the bedrock and to intense solifluction processes (Fyles and Craig, 1965). At its maximum development the glacier complex probably occupied the entire archipelago except for the coastal part of southern Melville Island which was covered by the Laurentide Ice Sheet. Active ice flow in the western part of the archipelago was largely restricted to the channels between the islands. Present data on ice-flow directions and on marine overlap suggest that the ice was thicker over the eastern and southern parts of the archipelago and that recession was towards these regions; perhaps the glacier complex preferentially developed adjacent to the Greenland and Laurentide Ice Sheets. Final recession and wastage may have taken place on the individual islands. Blake (1964) considers that some parts of Bathurst Island were deglaciated by 10,000 years B.P. and that extensive areas were ice free by 9,000 years B.P.

Prince Patrick Island. Glacial striations and other features at the head end of Mould Bay were formed by ice moving southward from the interior of the island (Tozer and Thorsteinsson, 1964). These features are attributed to the action of a small local ice cap, presumably of Wisconsin age, that developed subsequent to the establishment of the main drainage system. Mould Bay and the adjacent inlets were deepened and scoured. J. G. Fyles has noted boulders at the southwestern end of the island that were transported from the east and probably indicate a westward channelling of ice flow from Prince Patrick and Melville Islands. He also has noted that marine overlap is greater on the eastern side of the island, suggestive of thicker ice eastward.

Melville Island. Tozer and Thorsteinsson (1964) have concluded that the many fiord-like inlets in western Melville Island were probably originally stream-cut valleys that have been scoured by ice from a local ice cap. They also noted many glaciated valleys in interior western Melville Island. J. G. Fyles found erratics on both northern arms of the island that were far north of their outcrop area and presumably were deposited during the last regional glaciation. The present glaciers on western Melville Island are not remnants of the Wisconsin ice cap, for they occur in areas with youthful, little modified, postglacial ravines. The ice caps are probably less than 200 feet thick.

The maximum marine overlap is on the eastern side of the island, the highest raised shore features lying at 235 feet elevation (Henoch, 1964). Shells from the highest deposits were dated at $9{,}075 \pm 275$ years (I-730) and more from 175 feet of elevation were dated at $8{,}275 \pm 320$ years (I-GSC-21). The marine limit appears to decrease westward as no evidence of marine overlap was noted in western Melville Island. Henoch suggests that the depression of the east coast was produced by an ice cap centred near Sabine Peninsula, whereas depression of the south coast was caused by Laurentide ice.

Bathurst Island. The Queen Elizabeth Islands Glacier Complex occupied the whole of Bathurst Island and apparently restricted the northward expansion of Laurentide ice. Blake (1964, 1965) found abundant meltwater channels, patches of drift with ice-flow features, some dead-ice topography, rare striations, wave-modified esker and end moraine remnants, and other data indicative of former glacier cover. The ice may have been thicker over the northern end of the island than the southern as marine limit is about 450 feet in the north, only about 350 feet in the western and southwestern parts, and is rarely more than 300 feet along the east-central and southeast coasts. Recession appears to have been towards a remnant glacier in the interior of the island. Ice flowed northward off the island and may have reached Lougheed Island for

Fyles (1965) has noted erratics of rock types found on Bathurst Island. Marine overlap on Lougheed Island reaches an altitude of 300 feet. Shells from 25 to 100 feet above sea level were dated at 8,200 ± 180 years (I-GSC-24).

Cornwallis Island. According to Thorsteinsson (1958) the island was glaciated at least once. Striae at Resolute Bay on the south coast and Read Bay on the east coast indicated ice flow from an island ice cap. U-shaped valleys and other evidence of glacial scour in eastern coastal areas also indicated ice flow from the island. A series of rock basin lakes associated with *roches moutonnées* some 12 miles inland from Read Bay indicates northward ice flow. As glacial erratics of both igneous and metamorphic rocks occur in many places at elevations above the marine limit it is probable that Laurentide ice reached the island prior to the ice-cap stage but the age of this event is not known. Thorsteinsson places the postglacial marine limit at 425 feet elevation and notes that the well-defined beaches occur below 275 feet.

Devon Island. Ice caps on the highlands and plateaux of Devon Island cover about a quarter of its surface. Though the main ice cap has a maximum altitude of about 6,100 feet it is presumed to have a thickness of only 1,000 to 2,000 feet. It is passive except for a few distributaries in the eastern highland which flow in deep channels. Elsewhere the role of the ice caps appears to have been that of a protective cover and the smaller ice caps are known to be frozen at their base. Meltwater is carried by supraglacial streams. The physiography of the present unglaciated parts led Roots (1963) to conclude that the former, larger ice cap was also largely protective rather than an agent of erosion. He noted, however, that Griffin Inlet on the west coast was glacially sculptured by west-flowing ice and that north of there the ice margin was locally active, accounting for a series of lakes. Lakes in the west-central part of the island were also formed by west-flowing ice from the larger ice cap as it diverged southward to Maxwell Bay and north to the north shore of the island. Near the northwestern edge of the large ice cap a succession of stream channels and numerous ice-marginal meltwater channels indicate former melting but the ice near the present edge does not display these features. In the northwestern part of the island a more extensive ice cap than the present one was responsible for some 'hummocky' moraine south of Norwegian Bay and along the west coast south of Grinnell Peninsula, but the peninsula itself, like many of the islands, shows no evidence of having been glaciated. Glacial striae and crystalline erratics on Beechey Island, at the southwestern end of Devon Island, are believed due to northwestward ice flow. A granitic erratic found west of Jones Sound was also attributed to westward-flowing ice.

Brock, Borden, and Mackenzie King Islands. On Brock Island, Fyles (1965) observed an arcuate belt of ridges that he regarded as a thrust moraine formed by a glacier flowing westward through Wilkins Strait. No other glacial landforms were recognized on Brock or Borden Islands although a few erratics are present. The marine limit on Mackenzie King Island lies between 50 and 100 feet elevation and on Brock Island at about 50 feet elevation. It is unknown on Borden Island, but a beach crest at the west end is 10 feet above sea level.

Ellef Ringnes Island. Evidence of glaciation on Ellef Ringnes Island is rare, in large part due to solifluction and also to the nature of the bedrock. Crystalline erratics occur northwest or west of their outcrop and well above all evidence of marine overlap. Two gravel ridges trending west and one trending southwest are interpreted as eskers (St. Onge, 1965; Fyles, 1965). The marine limit is 150, 110, and 75 feet on the east-central, south, and west-central coasts of the island respectively. There is no information on the Pleistocene of Amund Ringnes Island.

Meighen Island. Fyles and Craig (1965) report the occurrence of numerous, large striated boulders on Meighen Island and the occurrence of striae, ascribed to westward-moving ice, on nearby Fay Islands. They believe that late Pleistocene glacial ice emanating from Axel Heiberg Island extended westward to cover these islands. The present ice cap has a summit elevation of just over 900 feet and a maximum thickness of about 400 feet (K. Arnold, 1965). Arnold has determined over the past ten years about 15 feet of thinning at the summit of the ice cap and at the northern edge, but notes that a slight deterioration of climate could cause an increase in the ice cap. He observed postglacial marine features to about 50 feet above sea level but places the marine limit at about 80 feet.

Axel Heiberg Island. On west-central Axel Heiberg Island meltwater appears to have been greater in early postglacial times than during the present arid climate (Müller, 1963). Because of the frozen ground and lack of vegetation, when run-off occurs now it is rapid, eroding channels and depositing valley alluvium and fans. The short valleys that occur between the glaciers and the sea are U-shaped only near the ice front and become V-shaped with valley trains and small fans in their lower ends (Rudberg, 1963). Some partly dissected trough-shaped valleys also occur in parts of southern Axel Heiberg not now occupied by glaciers. Abundant striae in the west-central part of the island indicate two periods of ice movement, the first from the north and east and the second from the southeast and east-southeast. Rudberg did not find undisputed evidence of far-travelled erratics in this area but red granite, presumably from Ellesmere Island, occurs on Schei Peninsula on the northeastern side of Axel Heiberg. There also, inland ice-flow features indicative of late ice movements trend northward. South of Schei Peninsula, J. G. Fyles observed red granite erratics that had been transported northwestward, roughly parallel to Eureka Sound; such movements represent former major trunk glaciers seeking

escape from between Ellesmere and Axel Heiberg Islands. Elsewhere striae appear to record local, late, ice flow down small valleys and fiords as the glaciers receded inland towards the higher areas.

Two main stages of glaciation, the younger definitely Wisconsin (Boesch, 1963) and the older unknown, are recognized in west-central Axel Heiberg. As the Wisconsin glacier waned and receded onto the island the sea encroached on the uncovered ground, reaching altitudes now some 250 feet above sea level. Shells from the highest features have been dated at 9,000 ± 200 years (L-647F). Isostatic rebound was fairly rapid as the basal layer of a peat deposit lying at 80 feet elevation was dated at 4,210 ± 100 years (Bern). The Thompson glacier has not advanced more than 2 kilometres in the last 4,000 years according to Müller, and also the climate of the last 9,000 years never favoured a glacial advance of more than a few hundred feet beyond present positions. It is currently forming a push moraine of huge blocks of stratified material up to 60 feet thick. Organic matter in the lower part of some blocks has been dated at 6,200 ± 100 years (L-647A) and a piece of driftwood from a high part of the push moraine was dated at 5,325 ± 270 years (Gx-0144). This seems to indicate that formation of the push-moraine commenced after the hypsithermal period and that 5,300 years ago the sea occupied the area now covered by the snout of Thompson glacier. This implies that sea level was at least 50 feet higher than at present. Marine shells from a terrace 120 feet above sea level have been dated at 5,330 ± 195 years (Gx-0143).

Ellesmere Island. Ellesmere Island is by far the largest of the Queen Elizabeth Islands and supports the largest ice caps in Canada. In its mountains and plateaux the former ice cover probably formed a complex of coalescent valley glaciers and mountain ice caps, which conformed in a general way to the topography and were similar to the major glacier complexes on the island today (Fyles and Craig, 1965). In some places the distribution of erratics indicates a former ice cover that was thick enough to maintain a regional flow across the grain of the topography. Christie (1967) reports that boulders of granite and gneiss, probably from Greenland occur over a 10-to-25-mile-wide coastal belt of Hazen Plateau. The maximum westward extent of the Greenland ice is not known, as late Ellesmere ice has since advanced southeast and east across Hazen Plateau to the coast. At the glacial maximum the whole of northern Ellesmere Island was under ice cover and shelf ice probably extended well beyond the present north shore (Hattersley-Smith, 1961). The coastal ice caps on northern Ellesmere Island west of the northernmost tip, and disposition and shape of stream valleys and fiords which indicate former trunk glacier flow in western and northern parts, may be a result of westerly and northwesterly snow-bearing winds and high altitude. Taylor (1956) and Smith (1961) however, have suggested that the fewer fiords on Hazen Plateau may be partly a result of confluence of Greenland and Ellesmere ice.

The valleys of northern Ellesmere are regarded as normal preglacial valleys that have been straightened and deepened by strong glacier flow. Disraeli Fiord has a water depth in excess of 965 feet, and seismic data indicate a submarine canyon, trending northwest from the fiord mouth, about 2,800 feet deep (Crary, 1956). Despite this great depth the glacier in Disraeli Fiord appears to have overridden Ward Hunt Island, giving it a *roches moutonnées* shape and emplanting gabbro erratics from Ellesmere Island to an elevation of 1,100 feet (Hattersley-Smith, 1961). The present streams are deep and mostly V-shaped, but near their head, at the glacier margin, they are U-shaped, and tributary valleys are hanging relative to the trunk valleys.

As the ice receded from the north coast the sea inundated the land to a maximum altitude of 400 feet, as on Ward Hunt Island, but the marine limit varies greatly from place to place presumably due to the presence of late ice. Marine shells from 125 feet above sea level have been dated at 7,200 years. Cyclically-bedded silt, sand, and organic materials, the last as much as 4 inches thick, occur at elevations up to 70 feet at the head of Clements–Markham Inlet. These have been interpreted as deposits formed during the climatic optimum but J. G. Fyles reports that they also occur at various altitudes in many valleys and may have a considerable age range.

The character of the northern valleys has led all workers to conclude that the present ice cap is little changed since the climatic optimum. At present there are again extensive areas of shelf ice up to 50 feet thick and 12 miles wide, and also land ice near the coast. Hattersley-Smith (1961) considers these to be expressions of the colder climate following the climatic optimum (about 5,000 B.P.) whereas the interior ice fields are largely remnants (maximum of 2,700 feet thick) of the main Wisconsin glacier complex. He has recorded the advance of valley glaciers over raised beaches and into V-shaped valleys, and the growth of low ice caps on areas of raised beaches. He concludes that the climate has ameliorated in the past few decades and, although the main, high-level ice caps and outlet glaciers in northern Ellesmere Island show little or no change in areal extent from year to year, thinning is taking place in the lower reaches of the outlet glaciers and both thinning and recession of the small low-level ice masses have occurred. On the north coast of the island the surface wastage and occasional calving by the ice shelf reflects the warmer trend of recent decades.

In the Lake Hazen and Hazen Plateau regions Ellesmere ice sheet receded towards the United States Range (Christie, 1967). Directional ice-flow features and distribution of erratics indicate east and southeast ice flow across the plateau. Even though ice-flow features are present, the stream valleys show little other evidence of major trunk glaciers; truncated spurs, hanging valleys, and over-deepening are rare. A piedmont glacier must have formed on the plateau by coalescence of glaciers

stemming from the United States Range. Lake Hazen basin was excavated along one contact of a wedge-shaped body of weak rocks overlying stronger rocks, and the lake is at least 864 feet deep.

In the western part of northern Ellesmere Island fiord-like valleys, glacial striae, and the distribution of erratics also show outward movement of trunk glaciers. Erratics at high elevations indicate regional flow across a region of ridges and valleys with a relief of more than 3,000 feet (Fyles and Craig, 1965). They report that during deglaciation the complex of glaciers and ice caps in western and central Ellesmere Island retreated progressively away from the outer coasts, up fiords and valleys, and into various high plateau and mountain areas. They also note that end moraines are not abundant and occur mostly in valleys within a few miles of an existing ice cap or glacier. The inferred period of shrinkage, on the basis of a few radiocarbon dates, is thought to be late Classical Wisconsin. An end moraine in central Ellesmere Island about 3 miles from the western margin of the ice cap was built more than 6,400 years ago, and another less than a mile from a modern glacier, more than 4,200 years ago; this indicates that the rate of recession over the past 6,000 years has been rather slow. Müller (1963) reports shells from an altitude of 350 feet at Eureka that dated 9,550 ± 250 years (L-647b).

Cordilleran Glacier Complex

The Cordilleran Glacier Complex is the system of intermontane, piedmont, and valley glaciers that developed in the Cordilleran physiographic province. Growth of the glacier complex stemmed from the passage of moisture-laden Pacific air masses over the various mountain systems.

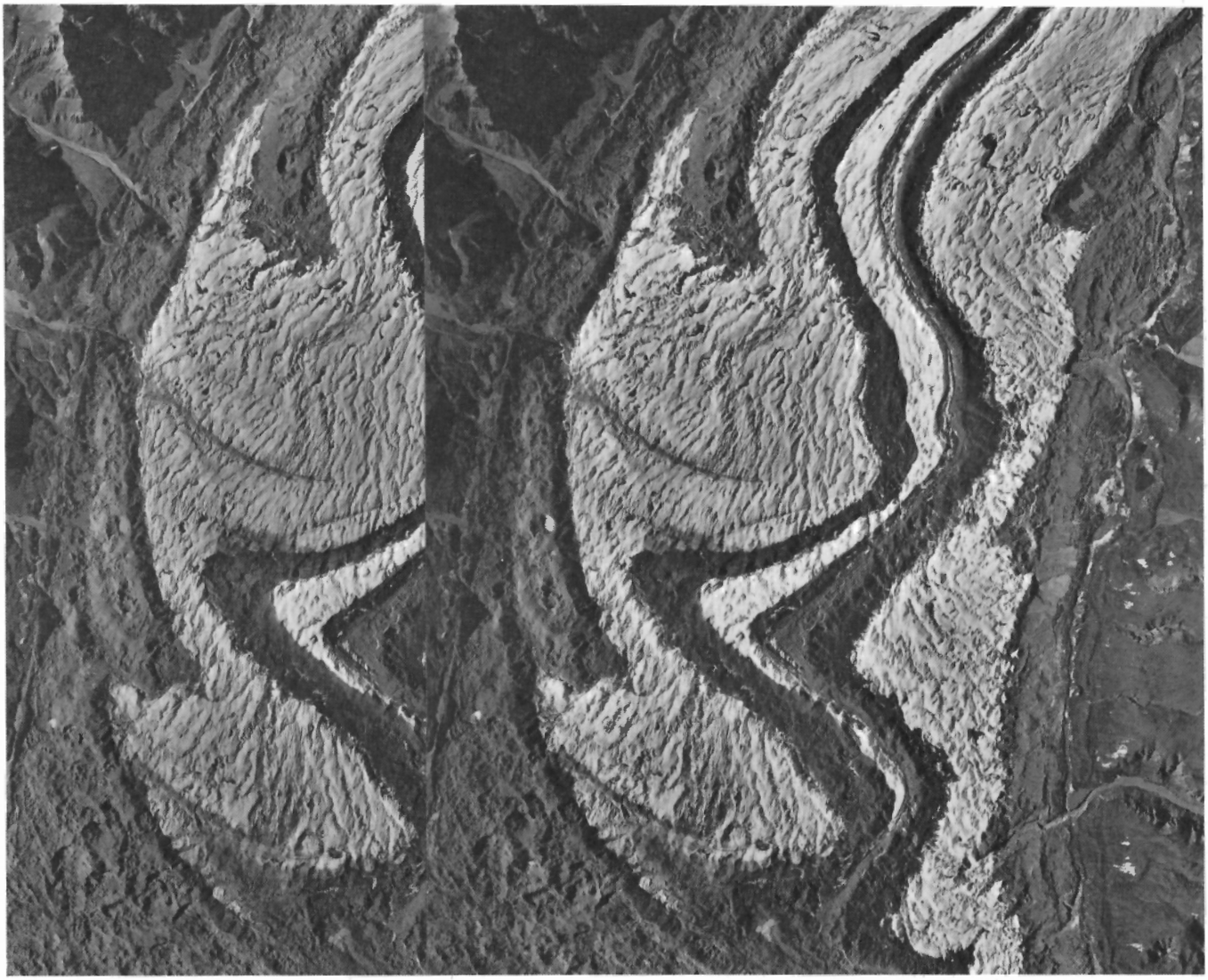


PLATE XII-17. Klutlan Glacier, St. Elias Mountains, Yukon Territory. Note the marginal and medial moraines, drift-mantled terminal zone with small ponds, meandering meltwater channels on glacier surface, and termini of valley-side streams at glacier margin. Glacier flows east and north at head-end Generc River valley. Vertical airphoto; 1 inch to 5,200 feet; stereoscopic pair.

It is probable that extensive alpine glaciation preceded development of the major ice fields. Snowfall was, without doubt, heavier on the Coast Mountains than on mountains in their lee and it was also heavier on the western sides of all mountain ranges than on their respective eastern flanks. As glacial conditions continued, however, glaciers formed on the eastern flanks and later contributed to in-filling of interior valleys, plains, and plateaux.

Glaciation

Coast and St. Elias Mountains. The westward-moving glaciers of the Coast and St. Elias Mountains were largely dissipated in the Pacific Ocean. Farther south a glacier probably occupied much of Hecate Strait and extended to Queen Charlotte Islands. A major glacier from the southern part of the Coast Mountains occupied the Strait of Georgia, probably nourished in part by glaciers from the high central part of Vancouver Island. It moved southwestward across the southern part of the island where it scoured mountains to at least 5,100 feet (Fyles, 1963b), and southward to beyond Seattle, Washington (Armstrong, *et al.*, 1965). In Fraser Valley the last major glacial period is termed the Fraser Glaciation (Armstrong, *et al.*, 1965) and it probably represents the same geologic-climatic episode as the Classical Wisconsin Glaciation of the mid-continent region. The build-up and advance include a period of alpine glaciation, the Evans Creek Stade and an ice sheet glaciation, the Vashon Stade, when ice occupied the lowlands of southwestern British Columbia and adjoining parts of Washington. On Vancouver Island all the deposits relatable to the last regional glaciation are referred to as Vashon drift (Fyles, 1963b). The last regional ice sheet is believed to have entered the northern end of Strait of Georgia after 25,000 years B.P. but it did not reach southern Vancouver Island until after 19,000 years B.P.

Central interior. Except for those along the Pacific Coast, the glaciers from all parts of the mountain systems flowed into the intermontane areas and sought escape along the river valleys. Because of topographic confinement, the interior glaciers thickened until they covered all but the highest peaks of the Interior Plateau, a belt of relatively less precipitation. In central British Columbia the ice attained a thickness in excess of 6,000 feet according to H. W. Tipper, and may have thickened until its surface became a major area of accretion standing somewhat higher than the confining mountains. Ice flow was mainly eastward from the Coast Mountains across Nechako Plain to where it was deflected north and south along Rocky Mountain Trench. Ice flow at the Wisconsin maximum paralleled the trend of southern Cassiar Mountains, but locally the ice flowed across the ranges towards Liard Plain.

Southern interior. The first ice to enter Okanagan Valley probably stemmed from gathering grounds in Monashee Mountains to the north and east (Nasmith, 1962). The southern tributary valleys show little evidence of glaciation; apparently there was little tributary ice from the adjacent plateaux and highlands. The valley glacier was then joined by the major ice sheet from Coast and Cascade Mountains which spread south and southeast over Columbia Plateau and reached a terminus some 90 miles south of the International Boundary. At the boundary the ice sheet reached an altitude of 7,200 feet, and it was accordingly some 6,300 feet thick over Okanagan Valley. Meltwater flowed south down Columbia River to the Pacific. A great thickness of lake sediments resting on outwash has been drilled in Okanagan Valley and in one place the valley bottom lies below sea level. The valleys, therefore, probably attained their present form prior to the last glaciation.

Eastern region. The glaciers in the eastern part of southern Rocky Mountains were formed by local precipitation and were not in contact with the ice sheet west of the continental divide. However, north of Athabasca Valley eastward movement of ice across the divide did occur and some glaciers from the upper part of the Fraser River system flowed into Athabasca River valley. The southern limit of strong ice flow across the Rocky Mountains lies about 100 miles south of Peace River.

Piedmont glaciers undoubtedly were present locally south of Athabasca Valley, but most ice was contained in the valley glaciers that flowed eastward through the Foothills onto the Plains to where the ice either dissipated or merged with Keewatin ice. The time relations between the two glacier systems have not been established everywhere. South of Crowsnest Pass the valley glaciers only reached the Foothills and had receded into the mountain valleys before the Keewatin ice had spread that far west. Keewatin drift extends some miles up the vacated valleys overlapping the Cordilleran drift (Stalker, 1960, pp. 72-73). Farther north, the Cordilleran glaciers and Keewatin ice did not merge during the Wisconsin but were separated by Porcupine Hills, a narrow, elongate belt of relatively high hills (Douglas, 1950). In the Foothills west of Calgary, Cordilleran glaciers had also attained their maximum development and receded before Keewatin ice reached the area and overlapped some 5 to 20 miles onto the recently vacated ground. Farther north Cordilleran ice in Athabasca Valley merged with Keewatin ice and was deflected southward east of the Foothills as earlier mentioned.

In Peace River valley the Cordilleran ice also appears to have reached a limit some distance east of the later western limit of Keewatin ice. There, however, a readvance of Cordilleran ice is recognized; a relatively late Cordilleran glacier from Rocky Mountains, reached to within 15 miles of Fort St. John after Keewatin ice had begun to recede (Mathews, 1963). North of Peace River the relationships of the Cordilleran and Keewatin glaciers are little known.

In the Liard Plateau and the adjacent plains Cordilleran and Keewatin ice made contact in only a few places. The Cordilleran glaciers were not very extensive

in the Mackenzie Mountains nor along the eastern side of Selwyn Mountains as this was apparently a region of low precipitation. Glaciers from the eastern Selwyn Mountains and to a lesser extent from the western side of Mackenzie Mountains partly filled the intermontane area. Only elongate valley glaciers flowed eastward and northward towards the unglaciated or older glaciated terrain. Valley glaciers that developed on the dry, eastern side of Mackenzie Mountains were short and seldom reached low altitudes. In places they extended somewhat beyond the western limit of Keewatin ice.

Northern interior. The limit of the Wisconsin Cordilleran ice in western Yukon is shown on Map 1253A. The limit probably corresponds with the McConnell Moraine and glacial advance of Bostock (1966) in central Yukon Territory. This limit is marked by fresh ice-marginal landforms, particularly in Stewart River valley a few miles below Mayo. Glaciers moved northward from areas of heavy snowfall in the St. Elias, Coast, and Cassiar Mountains, westward from the Selwyn and Pelly Mountains towards the dry Yukon Plateau, reaching somewhat short of the limit reached by pre-Wisconsin ice sheets. Near the ice margin innumerable monadnocks protruded above the glacier surface. The Ogilvie Mountains, which separate the Yukon and Porcupine Plateaux, also nourished glaciers; these moved outward from several of the higher areas within the mountain complex but did not reach the plateaux. Meltwaters on the northern side of Ogilvie Mountains flowed northeastward via upper Peel River until they encountered Keewatin ice in the lower part of the river basin; then, joined by meltwater streams from the Keewatin ice, escaped northward along Eagle River to Porcupine River, Yukon River, and finally Bering Sea. As deglaciation proceeded and Keewatin ice vacated lower Peel River valley, meltwater from Ogilvie, Selwyn, and Mackenzie Mountains joined that from the Keewatin ice and followed Peel River to Mackenzie Delta and Beaufort Sea.

Deglaciation

West coast. The last regional glaciation (Vashon) on southern Vancouver Island is represented by a till and associated deposits (Fyles, 1963b). Both are generally sandy, especially in the lowlands where southwest-moving ice from Strait of Georgia overrode the sand deposits of the Quadra non-glacial interval. As the ice thinned, southwest flow was maintained for a time through mountain passes and valleys and received additions from glaciers stemming from high parts of the island. As recession continued island-derived ice became dominant. According to J. G. Fyles, ice retreat from Strait of Georgia was probably underway by 14,000 years B.P., and by 12,800 years B.P. the western side of the strait was open so that much of southern Vancouver Island was ice free. Along the mainland coast, thinning and marginal recession was prolonged by active accumulation in the high parts of Coast and Cascade Mountains. The last major ice lobe in Fraser Lowland was in contact with the sea until after 11,000 years B.P., when a major re-advance into the sea—the Sumas Stade—deposited a widespread layer of stony and locally shell-bearing marine clay varying from a few feet to over 500 feet thick. Deposition was probably from icebergs except near the ice margin where shelf ice was present (Armstrong, *et al.*, 1965). North along the Pacific Coast the ice progressively lingered longer in contact with the sea, probably to about 10,000 years B.P. Along the Alaskan coast the glaciers had receded to or near their present positions by 8,000 to 7,500 years B.P.

Southern and central interior. Deglaciation of the southern part of Cordilleran Glacier Complex was accomplished largely by downmelting and stagnation of the ice mass as a whole, with no clearly defined halts or re-advances (Nasmith, 1962). Lowering of the ice surface ultimately left the plateaux and highland areas bare, and when the remaining ice was confined to the valleys its surface area was greatly reduced. It seems likely therefore that withdrawal of the ice proceeded much more rapidly in the early stages of melting than in the later stages because the sharp reduction of the surface area may have been sufficient to bring ablation into equilibrium with ice accumulation. In Okanagan Valley lateral meltwater channels were cut at altitudes of 3,700 and 2,800 feet near the International Boundary. Later melting may have produced a nearly flat ice surface with meltwater flowing on top of the ice. The valley glacier was still thick enough to flow, however, and a northward ice-frontal retreat was maintained. Farther north up the valley the Okanagan ice lobe was more active and lateral meltwater channels were cut. A minor climatic change resulted in development of a few morainal ridges. Several glacial lakes formed in Okanagan Valley region. Glacial Lake Penticton was the largest lake and had the longest history. A beach of this lake is tilted upward to the north at a rate of 3.5 feet per mile. Volcanic ash in talus, alluvium, lake beds, and peat deposits covers a broad region and is correlated by Nasmith with the Glacier Peak eruption about 6,700 years B.P.

The last glaciation northwest of the Okanagan Valley is represented by a single till (Fulton, 1965). In many places, this rests on proglacial sediments that pass down into non-glacial sediments about 20,000 years old, suggesting correlation with the Fraser Glaciation of the Coast Mountains and Vashon Glaciation of Vancouver Island. According to Fulton, regional downwasting of the ice sheet followed, without oscillatory marginal fluctuations. This was the result of confinement of the ice to intermontane positions. In general, the southern margin first retreated northward and later west of north. As the ice sheet thinned it broke up into a number of stagnant or semi-stagnant lobes occupying the major valleys and lowlands. During the early phases of retreat meltwater was ponded in the lowlands and discharge was to the south. Upon further recession the meltwater flowed eastward into the Okanagan River system and thence via Columbia River to the Pacific Ocean. This route was used during de-

glaciation of most of the southern part of the Interior Plateau. Finally, however, recession of the ice sheet into Coast and Cascade Mountains opened the Fraser Valley and discharge took this more direct route to the Pacific (Mathews, 1944; Fulton, 1965).

Elsewhere in southern British Columbia less is known of the general pattern of deglaciation. The Cordilleran glaciers retreated to a number of mountain ranges where snowfall was heavy, maintaining for a time active glaciers in the major valleys well beyond their area of nourishment. As recession progressed, the valley glaciers retreated to within the confines of the surrounding mountainous belt and finally left the valleys entirely.

Relatively late ice is thought to have remained in northeastern Columbia Mountains, plugging the Rocky Mountain Trench in the vicinity of Big Bend, and in adjacent parts of Rocky Mountains. These mountains have glaciers at present, but whether they are remnants of the former sheet or were newly formed is unknown. A region of late ice is also postulated for a large part of southern Coast Mountains. H. W. Tipper (GSC) believes that as recession progressed the eastern flow from Coast Mountains lost contact with westward flow from Columbia Mountains. The latter, then directed largely by topography, turned north-northwest and south-southeast. The south-southwest trending ice divide between longitude 120 and 121 degrees (*see* Map 1253A) marks this line of separation. Most of the flow from Coast Mountains was northeast towards the Interior Plateau and upper Fraser Valley. Nechako Plain was covered by a broad, east to northeast flowing ice sheet from central Coast Mountains (Armstrong and Tipper, 1948). This part of the mountains at present has fewer glaciers than parts farther north and south. Glacial lakes were ponded in the valleys of Fraser, Nechako, and Stuart Rivers by active glaciers in the lower part of Fraser Valley; discharge was north along Parsnip River in Rocky Mountain Trench and down Peace River into lakes dammed by Keewatin ice in Alberta. It is thus possible that meltwater from glaciers originating in the Coast Mountains may at one time have flowed eastward via Peace River to the Interior Plains and thence via the Great Lakes system into the Atlantic Ocean.

Northern interior. Centres of late glacial activity persisted in the northern parts of Coast Mountains, and in the Cassiar and Skeena Mountains. In the Cassiar Mountains, reversals of ice-flow direction occurred during deglaciation, evident from the distribution of erratics and from ice-flow trends and ice-marginal meltwater channels (Watson and Mathews, 1944). Glaciers from the western side of Cassiar Mountains reached to the Pacific, but the relationships of these ice movements with those of ice in the Coast Mountains astride the outlet valleys have been little studied. During glacier recession, the Cassiar Mountains had an ice cap above 7,000 feet elevation from which glaciers flowed southwest and northeast. In the southern part, ice from the higher Coast and Skeena Mountains to the west passed through the Cassiar Mountains, along Dease River valley, flowing in a northeasterly direction and reaching to about 7,000 feet elevation. Alpine glaciation both preceded and followed the last major glaciation (Gabrielse, 1963).

In parts of the southwestern Yukon the last major glaciation, the Kluane (Denton and Stuiver, 1967), probably correlates with the McConnell advance in central Yukon Territory. Radiocarbon dates on organic materials from beneath Kluane drift indicate that the glaciation began about 30,000 years B.P. and persisted to at least 12,500 years and perhaps to less than 10,000 years ago. The Kluane glaciers, stemming from ice fields at altitudes of 12,000 to 15,000 feet on the eastern flank of St. Elias Mountains, moved northeastward through Kluane Ranges. It then flowed northwestward along Kluane River valley to Donjek and White Rivers, and northward to beyond Snag (Bostock, 1952; Krinsley, 1965). In the valley southeast of Kluane Lake the elevation of the surface of the glacier was 6,100 feet, near Donjek River it was about 5,000 feet and at its northern limit about 2,500 feet. The Kluane glaciers deposited three distinct tills which are associated with outwash and lake deposits. The glacier that deposited the uppermost till sheet in Kluane Valley had a minimum elevation of 2,500 to 3,200 feet. The glacier retreated along the paths of its earlier advance without any significant pauses or halts although large masses of ice stagnated in Kluane Lake area. Loess, derived from Kluane outwash, was deposited as a thin blanket over all older deposits up to an altitude of 4,500 feet. The Kluane Glaciation was followed by a period of soil formation (Denton, 1967). The soil was developed prior to the neoglaciation of St. Elias Mountains, about 2,800 to 2,600 years ago, which Denton correlates with the Little Ice Age. Following a relatively warm interval the glaciers were again active from about 600 years ago to the present century but are currently undergoing a fluctuating recession. The Donjek and Kaskawulsh Glaciers attained their neoglacial maxima about 300 years ago.

ECONOMIC CONSIDERATIONS

The surficial deposits or overburden that mantle the bedrock in most parts of Canada are mainly the result of glaciation. The direct glacier deposits or till, together with varied mixtures of ill-sorted or well-sorted materials, comprise the moraine that characterizes much of Canada's surface. Some large areas of gently rolling ground moraine in southern Canada have proved suitable for agricultural purposes but, in general, the irregular surface and variably stony materials limit or hinder large-scale farm operations. The more hilly areas of ground moraine, along with dis-

integration, interlobate, end, and kame moraines, therefore serve as ranchlands or forest sites, and are important also for water storage. Some of these morainal tracts with their myriad lakes provide game preserves, park lands, and recreational areas important to our tourist industry. Areas of glaciofluvial deposits are similarly important, particularly in regard to water supply.

Vast regions in southern Canada were formerly covered by glacial lakes and the resulting flat areas of fine-grained soils comprise the farm lands that supply most of the grains and vegetables essential to support the ever-expanding urban and industrial areas. In many parts of Canada glaciation was instrumental in providing better soils than we might otherwise have had, for glacial processes determined the mixing and size-sorting of materials that characterize some of our best farm lands. Elsewhere the glaciers probably removed excellent soils and transported them beyond our borders, leaving only bouldery or rocky terrain in their place.

Similarly, glaciation has played a vital role in our mineral and mining development. In parts of the Canadian Shield, Cordilleran, and Appalachian Regions the glaciers have stripped-off the weathered mantle of many orebodies, removing and dispersing the valuable secondary ones. Glaciation thus exposed the primary mineral deposits that foreshadowed the age of the prospector in Canada with consequent development of our mining industry. Elsewhere, orebodies perhaps once exposed by glacier action have been buried by younger glacial deposits and remain to be found by drilling and geophysical surveys. Others may be found through study of the surficial geology. Boulder tracing along the line of the last ice movements, as deduced from striae and other ice-flow features, is a simple technique that has been practised and proven successful on many occasions. More refined systems such as the glaciofocus method of localizing the source areas of minor constituents in drift (Lee, 1964, 1968) should prove successful also in the search for orebodies. A comprehensive knowledge of glacial events and glacier transportation will be even more necessary in future years as Pleistocene deposits are thoroughly examined for clues as to the loci of valuable minerals.

Glaciation has played a key role in the development of our towns, cities, and industrial areas. These are generally located in drift-covered parts of plains and lowlands, or in valleys in mountainous or rolling areas, rather than on adjacent more rocky and irregular-surfaced terrain. Because of the relative ease of excavating in unconsolidated materials as compared to bedrock there is still a tendency, despite modern heavy equipment, for urban expansion to avoid the areas of rock outcrop. This practice is engulfing the better arable lands at an alarming rate. Though some of our cities are in areas of little or no drift, most include areas of heavy drift. In some cases the bedrock is deeply buried and only the largest buildings are founded on bedrock, all others being anchored in the drift mantle. Where possible, spread-footings are located on till or gravel layers, and experience has shown that great care must be exercised to evaluate the physical properties of the drift mantle. The character and thickness of the drift mantle also have a direct bearing on the costs of water and sewer installations, and the construction of roads and sidewalks. Drift mantle may yet prove to be the deciding factor with regard to the type of rapid transit systems to be planned for many urban areas.

Sand and gravel are in great demand in all urban areas, and as a result supplies are being rapidly depleted within many miles and generally some tens of miles of these centres; whereupon, due to excessive transportation costs, crushed rock is used as a substitute. Most supplies of sand and gravel are from glaciofluvial deposits but locally river, lake, and seashore materials are employed. Despite competition from crushed rock, the use of sand and gravel for construction purposes alone remains at a high level. The 1965 production and value of sand and gravel used in road construction, railway ballast and in concrete, asphalt, and mortar mixes is given in Table XII-3. These materials are in large part of Pleistocene age, but some preglacial gravels are included.

A Pleistocene gravel deposit of unusual interest, which was not used for construction but rather as an iron ore, was that at the Canadian Charleston mine south of Steep Rock Lake, Ontario. There a gravelly end moraine included sufficient 'float' to be mined on a commercial basis as iron ore. Pebbles, cobbles, and boulders of high grade hematite and goethite were the main components of the gravel deposit. These were deposited from an ice lobe that had gouged them from beneath Steep Rock Lake. A high grade iron pellet concentrate was produced on the property. Although operations were confined to the summer months, and there was no production in 1961, some 642,957 tons of high grade iron concentrate was produced from about 4.5 million tons of gravel between 1959 and 1964 when operations ceased.

Clay deposits of both lacustrine and marine origin were used extensively in former years in the manufacture

TABLE XII-3 *Production and value of sand and gravel (by F. E. Hanes, Mines Branch)*

Province	1965	
	short tons	$
Newfoundland	4,063,734	3,684,891
Prince Edward Island	412,064	374,081
Nova Scotia	6,574,387	4,498,803
New Brunswick	4,491,514	2,594,846
Quebec	40,507,369	19,583,351
Ontario	75,082,026	55,297,474
Manitoba	9,757,104	6,767,068
Saskatchewan	8,570,008	5,615,794
Alberta	13,163,941	10,661,383
British Columbia	20,484,706	12,662,016
Totals	183,106,853	121,739,707

of brick, tile, and related products. Their use for brick and tile has been largely discontinued in areas where shale is available, and the amount still being used for this purpose is a small part of the over-all production of clay products valued at somewhat more than $31 million annually. Bricks made from Pleistocene clays are found in most brick buildings in the older parts of our cities and in the older farm houses.

Major construction projects such as water storage and power dams, superhighways, new railways, causeways, and port facilities must take surficial deposits into account, as some need to be removed and alternate materials hauled to the sites. Landslides involving these materials may occur where highways cut through hills or valley walls or where stream erosion undercuts valley sides. Clays or clay tills may be sought for impermeable blankets in water reservoirs, or permeable fill may have to be found for some earthfill dams. The glacial history of such regions should be known, for it may dictate the amount of testing necessary to evaluate the stratigraphic sequence at the construction site proper.

The effects of differential uplift need to be evaluated where major construction projects with a long-term lifespan involve large bodies of water. Differential uplift between opposing ends of the storage basins of only a small fraction of an inch per year may be important, and its postglacial record is requisite. The rate of differential uplift in Hudson Bay, based on short-term tide gauge records, may be about 2 feet per century (Barnett, 1966). Such information, together with the rate of postglacial uplift, would provide a useful curve of land–sea relations important in planning major harbour facilities in the region.

Waste disposal is becoming an ever-greater problem in industrial and mining areas. Areas of relatively impermeable clay are at times sought for waste disposal pits for corrosive or noxious materials, but the mineralogy of the clay should be determined with regard to possible chemical reactions within the deposit. Waste products that are less objectionable and that will break down in a relatively short time may be pumped into some drift-covered waste lands, but great care has to be exercised to prevent contamination of the groundwater.

Surficial deposits have a direct bearing on Canada's water resources for they determine the rate of run-off and amount of water storage. They hold much water that would otherwise be lost by rapid run-off, and they grudgingly but positively discharge this to the rivers and streams. Without this constant to semi-constant supply, many of our river systems would vacillate between raging torrents and dry valleys.

Thus our heritage of glacial deposits is fundamental to the very character of this land and affects the progress of our agricultural, forestry, mining, and fishing industries, urban and industrial developments, and recreational facilities, all of which influence our daily way of life and our future plans.

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XIII. Groundwater Geology

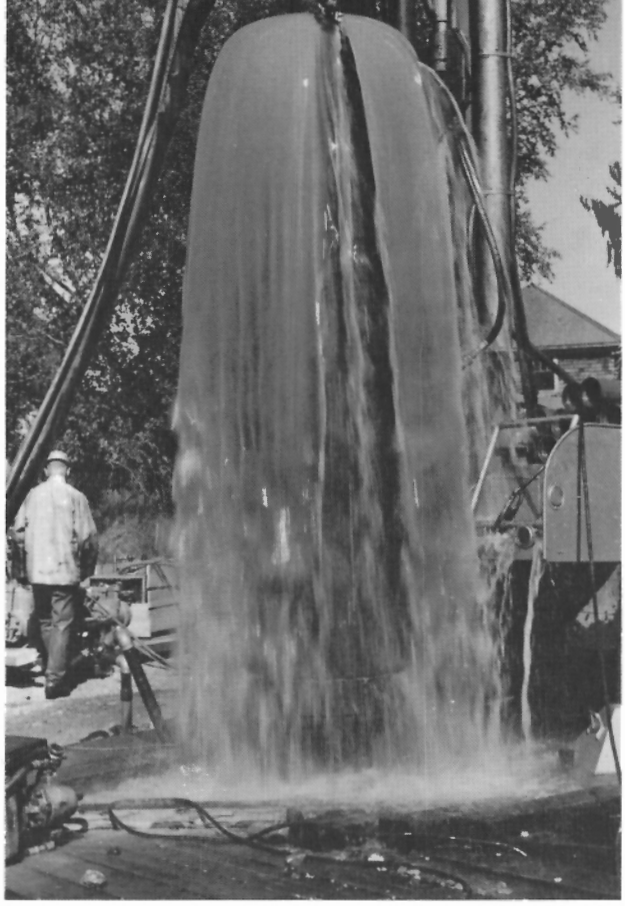
Introduction 766

Hydrologic Cycle 767

Hydrogeological Regions 771

Selected References 791







INTRODUCTION

In Canada groundwater contributes about 10 per cent of the water supplied by municipal water supply systems serving communities with a population of more than 1,000 and an equal or greater proportion of that used by individual homes (Table XIII-1). Quantities of industrial and irrigational water are also obtained from groundwater.

Groundwater is defined as all water in the zone of continuous saturation except that water chemically combined in minerals.

The upper limit of completely saturated material is the *water-table* and for practical purposes may be determined by the level of water standing in open holes penetrating the saturated zone. It may be more accurately defined as the surface at which water is at atmospheric pressure (Fig. XIII-3). The water-table may be at surface in wet periods when the ground is fully saturated, or it may be at depths of many hundreds of feet in mountainous and arid areas. The lower limit of groundwater is determined by the depth at which pressure closes all openings in the rocks. This depth depends mainly on the structural strength of the rocks, and generally is not known.

The first investigation by the Geological Survey of Canada specifically mentioning groundwater was in 1875 when exploratory drilling was done in the Northwest Territories (now Saskatchewan) to obtain water for railway construction. In a short report on drilling for groundwater at Deloraine and Morden, Manitoba, A. R. C. Selwyn (1890) gave a summary of the groundwater possibilities from Red River to the Pembina Escarpment. The first Geological Survey report fully devoted to groundwater was prepared in 1901 by F. D. Adams and O. E. LeRoy, on the artesian wells of Montreal Island. Mapping of the Milk River sandstone in southern Alberta by D. B. Dowling in 1915 indicated that it should be an extensive artesian aquifer; this was substantiated by a drilling program the following year. In the 1920's and 1930's reconnaissance mapping on the Prairies for oil, gas, and water produced some wide-ranging, excellent reports. In 1935 a massive groundwater inventory of the wells in southern Saskatchewan and east-central Alberta was undertaken by the Geological Survey of Canada and 274 water supply papers were published. This type of well inventory was continued until about 1959 in Ontario, Quebec, Prince Edward Island, and the Prairies, with emphasis on the quantitative and chemical aspects and on appraisal of older data. Since then the Survey has studied groundwater flow systems and the fundamental behaviour of groundwater; well inventory data are gathered almost entirely by provincial agencies many of whom require drillers to submit reports on all wells drilled. These data are kept readily available and are an important source of information.

[1] Inland Waters Branch



XIII Groundwater Geology

I.C. Brown[1]

Artesian groundwater flow, Coldstream ranch, Vernon, B.C. From left, uncontrolled flow from 6" casing; installation of 30" flow-control casing; controlled flow from 30" casing.

HYDROLOGIC CYCLE

The hydrologic cycle is the complete cycle of processes through which all water on earth is continuously moving (Fig. XIII-2). The energy for this movement is provided by radiation from the sun which evaporates water from all land, water, vegetative and animal surfaces into the atmosphere. The sun's energy plus the rotation of the earth moves this atmospheric water to locations where conditions are such that it condenses in the form of either solid or liquid precipitation and is pulled back to the surface of the earth by gravity. Some immediately evaporates again, some runs off into streams and lakes and eventually into the ocean, some sinks into the soil and is either extracted and transpired by plants or evaporates directly, some sinks more deeply into the ground to the water-table, where it becomes groundwater. Groundwater, acting largely under the influence of gravity and the pressure of surrounding water, continually seeks areas of lower pressure or lower potential energy and follows complex pathways through the materials of the earth. Some does not travel far before it returns to surface and joins the other parts of the cycle, some travels to great depths and remains underground for very long periods, but all is eventually discharged on the surface or into streams, lakes, or the ocean from where it is again evaporated and continues on around the cycle. The amount of water in active storage in the ground is about 30 times that stored in streams and lakes, but is a little less than a quarter of that stored in ice and about 1/150th of that in the oceans.

Groundwater Fluctuations

Variations in groundwater flow both in time and space are considerably slower and more uniform than variations in either surface or atmospheric water. The rates of groundwater flow depend on the hydraulic gradient and on the permeability of the materials and are commonly slow, as the hydraulic gradient is usually considerably less than the surface slope in areas of high relief and is, of course, very slow in flat areas. Actual rates of movement vary from almost zero to a maximum of more than 100 feet per day, but a normal range is from 5 feet per year to 5 feet per day. These low velocities combined with the vast storage available in the geological materials of the earth combine to reduce the variations in groundwater flow to a minimum. Thus, whereas the water-table may fluctuate rapidly, depending on the precipitation, deeper parts of groundwater flow systems will vary much more slowly.

Groundwater fluctuations, which vary greatly in magnitude and timing, are due to many causes both natural and artificial (Figs. XIII-5 and XIII-6). Small rapid variations in the level of the groundwater-table can be caused by any force that affects the materials in which it occurs, such

TABLE XIII-1 *Canadian water supplies*

Province	No. of systems	No. supplying groundwater entirely or partly	Water supplied in 1,000's of gal. per day		Population served in 1,000's	Per capita supply gal. per day
			Groundwater	Total water		
Alberta	64	26	3,755	90,537	810	112
British Columbia	83	20	9,664	160,711	1,278	126
Manitoba	42	10	1,333	85,819	1,048	82
New Brunswick	23	16	10,185	57,019	250	228
Newfoundland	21	1	300	23,217	155	150
Nova Scotia	34	14	4,660	38,408	341	113
Ontario	294	120	97,824	669,638	6,443	104
Prince Edward Island	3	3	2,908	2,908	29	100
Quebec	272	81	28,156	573,660	4,286	134
Saskatchewan	47	33	10,691	51,348	254	202
TOTALS	883	324	169,476	1,753,265	14,894	Average 135
Groundwater supplied % of total 9.9						

Water supplied by Canadian municipal water supply systems serving populations of 1,000 or more
Does not include large industrial or private supplies not depending on municipal utilities
(Compiled from statistics in Canadian Municipal Utilities, Waterworks Manual and Directory for 1963)

Type of supply	Number of dwellings	Average occupancy	Estimated water supply gpd per dwelling	Estimated water supply 1,000's gpd
Piped into dwelling	759,349	3	200	151,870
Not piped into dwelling	377,401	3	100	37,740
Totals	1,136,750			189,610

Estimated water supply of dwellings not supplied by municipal water supply systems. Sources of water unknown but probably largely from wells. (Compiled from statistics in Canadian Housing Statistics, 1962, Central Mortgage and Housing Corporation, March, 1963)

as vibrations from traffic, earthquake forces, crustal tide forces, and air pressure changes. These short-term fluctuations are usually small and are measured in inches or fractions of inches, though those due to earthquakes are as much as a few feet. Longer term fluctuations are due to variations in supply of recharge water, and are more likely to be seasonal. They vary in magnitude depending on the rate of movement in the flow system and the frequency and amount of recharge, and in most areas of Canada vary from 5 to 20 feet. Fluctuations in the water-table are also caused by man's interference with any part of the flow system by pumping, irrigation, and drainage, altering river or lake levels, changing or removing the natural cover, and many other causes. In Canada such fluctuations are known to vary in amplitude from a few inches to many tens of feet and in extent from a few hundred feet to many miles.

In general throughout Canada recharge is greatest and groundwater levels are highest during spring break-up (Fig. XIII-6). From this peak they fall gradually to a low late in the summer; only prolonged and widespread rainy periods during the summer will reverse the trend, as most summer precipitation is returned to the atmosphere by evapotranspiration. The lowest levels are reached in the early autumn. The autumn rains come when evapotranspiration is at a minimum and considerable recharge takes place. In much of Canada, water-tables rise until freeze-up after which most precipitation is in solid form and little soaks into the ground. During the winter there is a gradual falling of groundwater levels until spring break-up occurs again.

Groundwater Flow Systems

Gravity is the principal force moving groundwater for it produces major pressure differences. Because the downward pull of gravity prevails through widely varying materials, pressure differences are set up and groundwater flow (though in general from topographically high areas to low areas) can be in any direction, including straight up. The flow can be influenced by other less continuous and universal forces such as temperature gradients, pressure due to underground bodies of gas, tectonic forces, capillary forces, and possibly osmotic pressures.

Although it may seem anomalous that water introduced into the groundwater system at atmospheric pressure can flow to great depths and thence return to discharge at

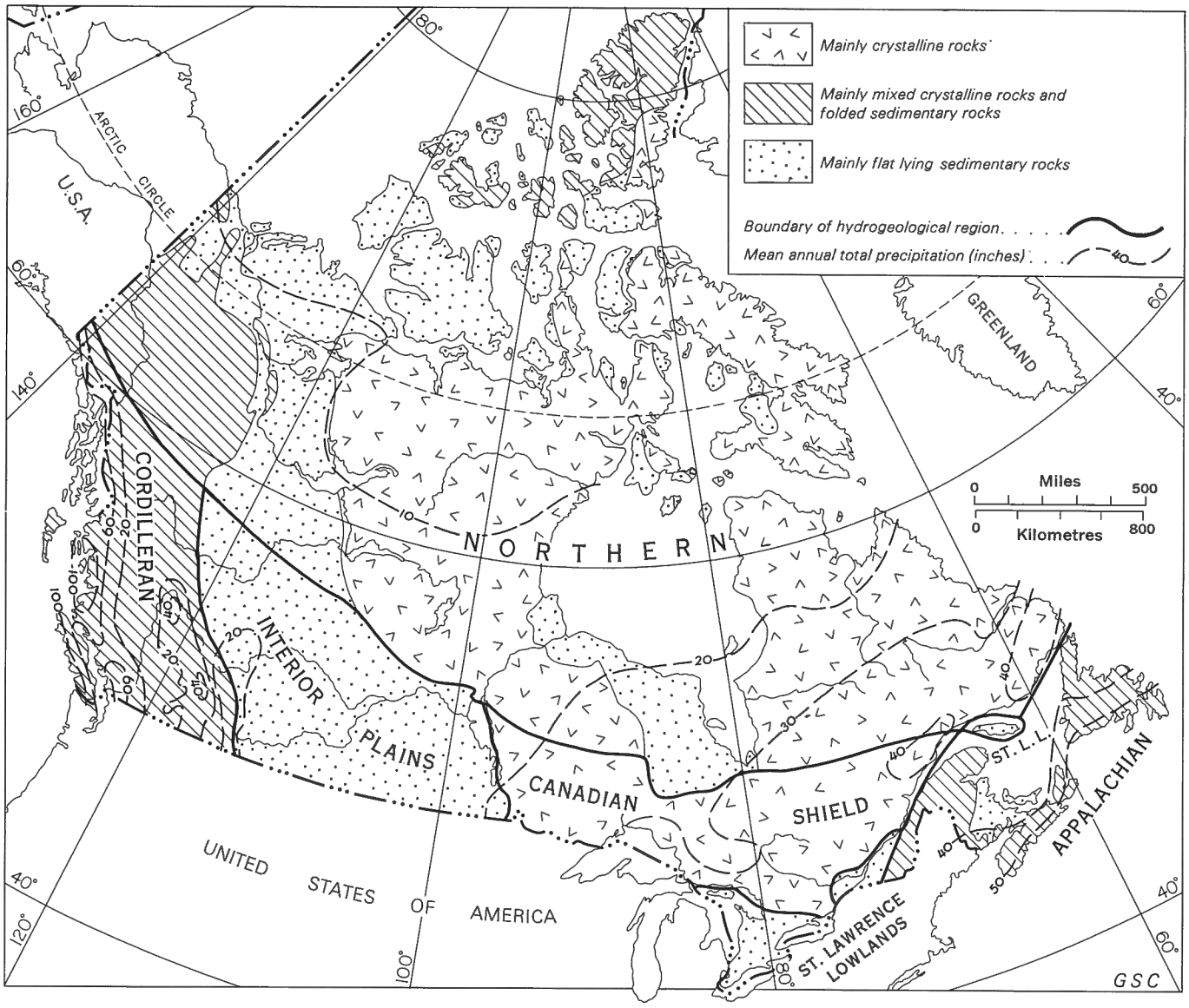


FIGURE XIII-1. Hydrogeological regions of Canada.

atmospheric pressure, groundwater is always moving down a hydraulic gradient. The pressure head, therefore, as measured along the flow system continually decreases from recharge to discharge.

Darcy's formula for the flow of fluids through porous media may be stated as follows: the quantity of water flowing through a given cross-sectional area equals the permeability times the hydraulic gradient. In groundwater flow systems permeability is a fixed property of the material, and consequently both hydraulic gradient and the quantity of water flowing are dependent on the geology of the material.

Water moves through the intergranular spaces in materials and also through open fractures that vary from minute size to large solution cavities and channels. Movement through intergranular spaces in porous materials is diffuse and fairly uniform within uniform material but most permeable materials have directional properties inherent from their conditions of deposition and permeability in one direction may be many times as great as that in another. Normally the horizontal components are greater than the vertical but this may vary greatly and differ even between individual layers of a single deposit. In some geological materials permeability may be due largely to fractures or bedding and the direction of maximum permeability may lie in any direction.

Groundwater flow systems may vary in scale from those related to a single small isolated pond to systems many hundred miles in extent and even to very deep flow systems of continental size. It is the larger flow systems with their comparatively constant flow that provide the water that keeps rivers and streams flowing during long

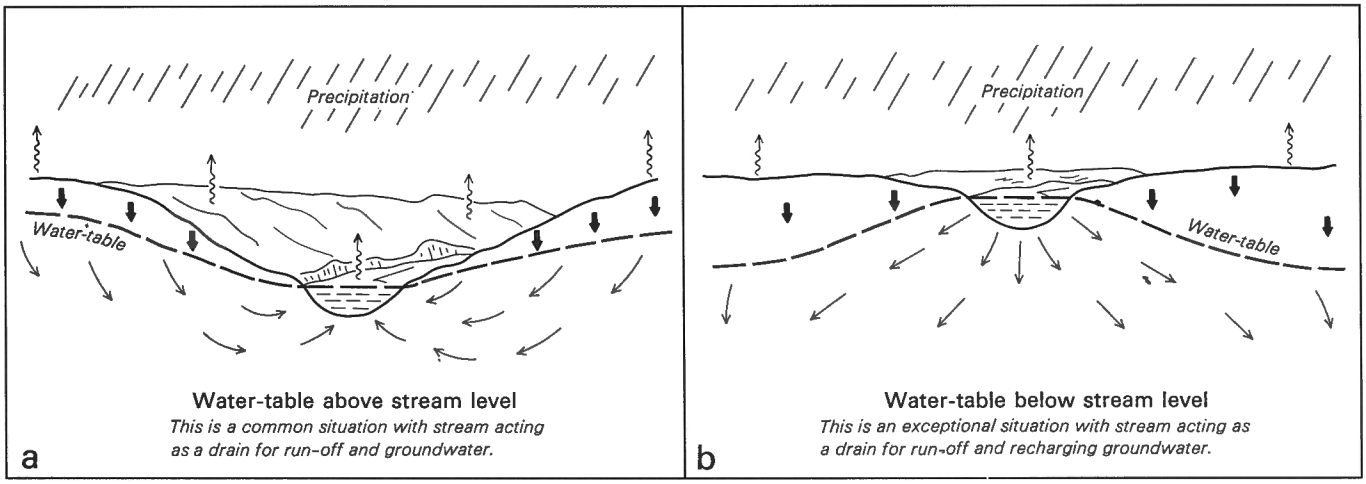


FIGURE XIII-2. Hydrologic cycle: a, influent stream; b, effluent stream.

dry periods. Thus, measurements of the low flow of rivers give a close estimate of the amount of water travelling through the related groundwater flow systems.

Flow systems may be detected by many methods. Direct ones are springs and seeps, pressure head measured in wells, and tracers of various types. Indirect methods are much more varied but largely depend on the chemistry of groundwater. On a very large scale, and regardless of

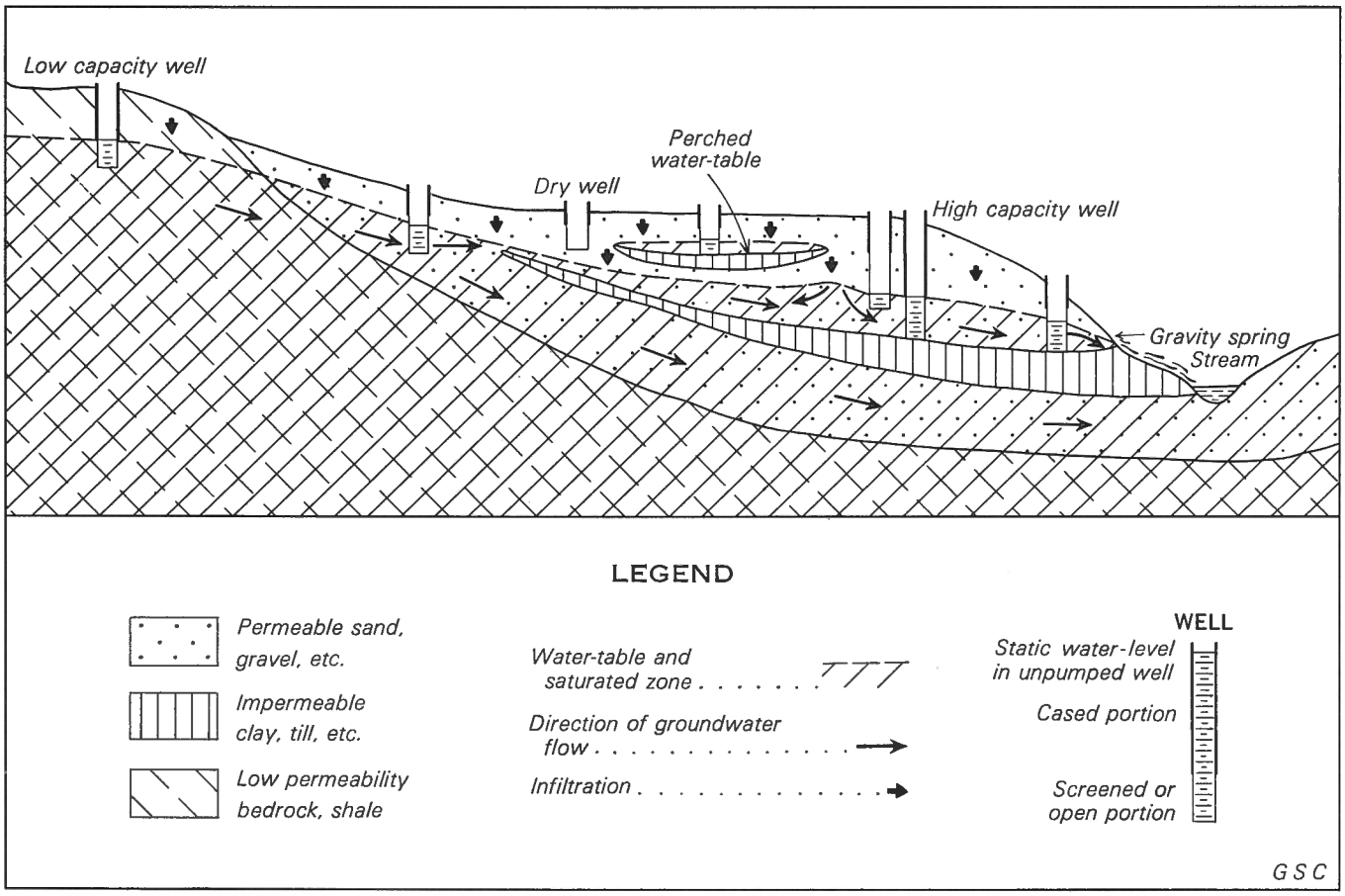


FIGURE XIII-3. Water-table conditions (unconfined aquifers).

the material through which it travels, groundwater starts out as bicarbonate water and changes through a sequence of carbonate plus chloride plus bicarbonate to chloride plus sulphate or sulphate plus chloride to finally a predominantly chloride water approaching the composition of sea water. This full sequence is not always present but has been found very useful in indicating flow systems in several areas in Canada. The composition of water also depends on the materials through which it flows, for example water in limestones and dolomites will be high in carbonate and that in gypsum-bearing rocks will be high in sulphate. In clean sands and gravels very little material will be dissolved. In general, water will be of better quality in areas of recharge than in areas of discharge. Though it is known that the rate of change of dissolved material in groundwater can be rapid, changes, for example, in composition of groundwater depend closely on the containing materials even though the flow system may cross the boundary between them.

Plant growth aids greatly in understanding groundwater flow systems. Salt-tolerant plants indicate areas of discharge from long flow systems whereas non-salt tolerant plants are found in recharge areas or over areas of lateral flow. Phreatophytic plants, or those depending on a continuous supply of moisture, normally grow where their roots can reach the water-table. Non-phreatophytic plants depend largely on soil moisture and can survive considerable periods of drought without contact with the water-table. Thus in areas where the water-table is generally deep, phreatophytic plants generally indicate discharge areas and non-phreatophytic plants recharge or lateral flow areas.

HYDROGEOLOGICAL REGIONS

Certain broad features of the geology, precipitation, climate, and topography serve to divide Canada into six regions of differing hydrogeological conditions (Figs. XIII-1 and XIII-7). The boundaries between the regions are not sharp but rather are zones of varying width wherein one group of conditions changes to another. In defining these

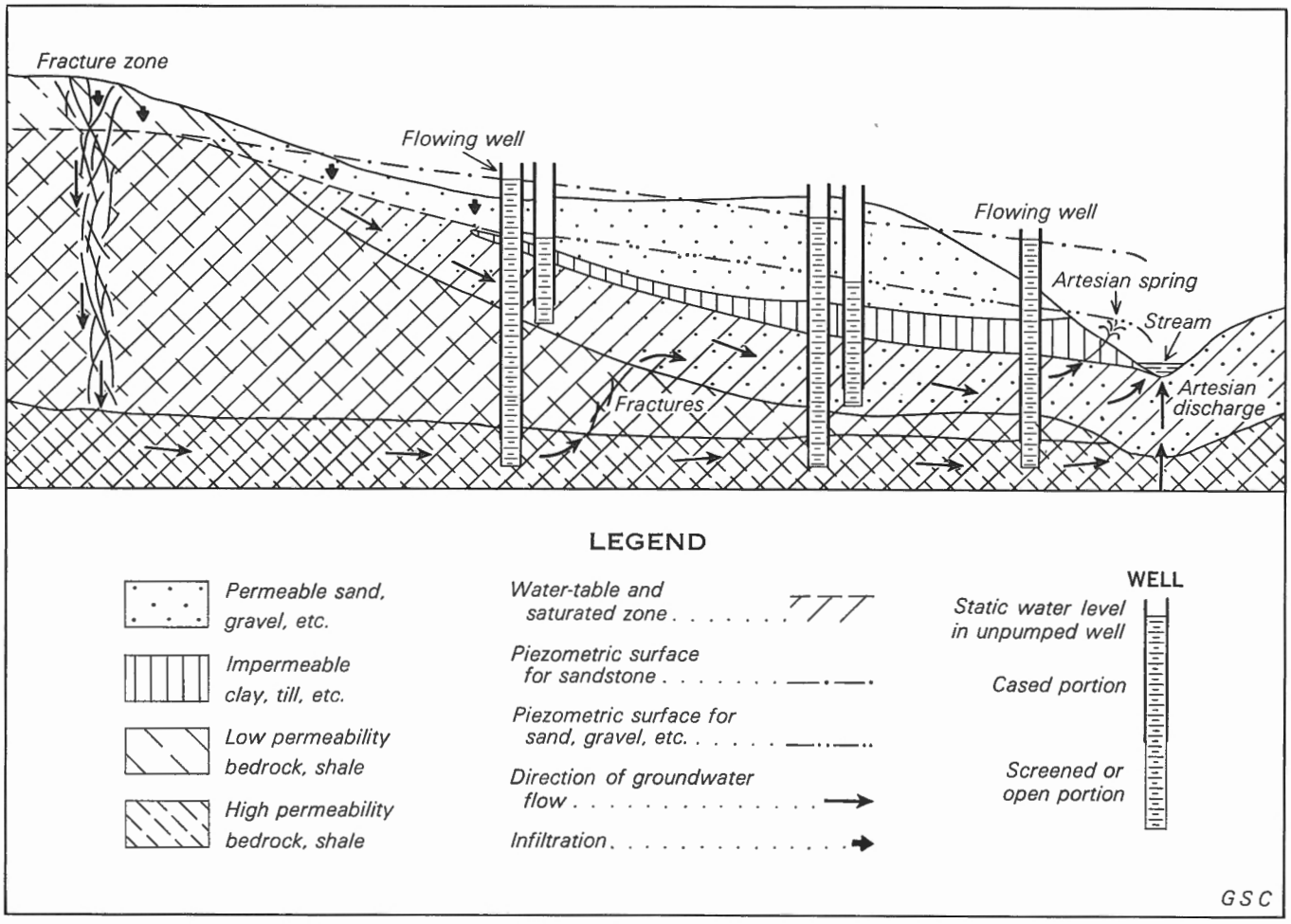


FIGURE XIII-4. Artesian conditions (confined aquifers).

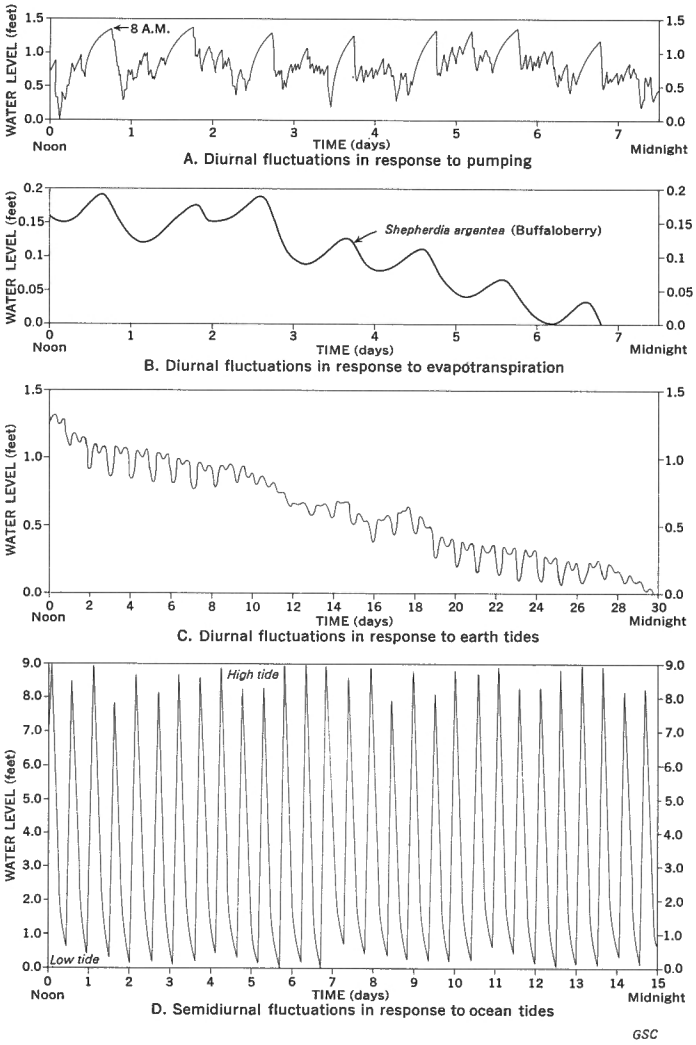


FIGURE XIII-5. Diurnal and semidiurnal fluctuations in groundwater levels.

regions, geology is the most important factor as it determines the amount of water that can be stored and transmitted. Precipitation and climate control the amount of water available for recharging groundwater flow systems. Topography influences the length and depth of the systems.

Appalachian Hydrogeological Region

This region is characterized by a mixture of crystalline, folded to flat-lying sedimentary rocks, and comparatively thin surficial materials. It is an upland sloping gently to the southeast, dissected by valleys and broken by broader lowland areas developed on belts of weaker rocks. Relief in the Gaspé Region exceeds 2,000 feet but elsewhere it is seldom more than 1,000 feet and commonly only a few hundred. The climate is humid continental. Total annual precipitation ranges from 30 to 50 inches with the maximum along the Atlantic coast, and is exceeded only by that in the western Cordilleran Region.

Hydrogeology

Groundwater in the consolidated rocks moves largely through fractures of various types and relatively little moves through the pore spaces of these rocks. The hydrogeological properties are, therefore, depicted by hydrostratigraphic units consisting of rocks yielding approximately the same quantity and quality of groundwater (Fig. XIII-8). This division is possible only in New Brunswick, Prince Edward Island, Magdalen Islands, and Nova Scotia, as sufficient data on which to make this division are not available for Newfoundland, Gaspé, and the Eastern Townships of Quebec.

New Brunswick, Prince Edward Island, and Nova Scotia

The basement complex comprises all rocks from Precambrian to Devonian age. These rocks seldom yield more than 5 gallons per minute. Wells drilled in excess of 100 feet, and particularly those over 1,000 feet, commonly yield saline water. The better wells occur in valleys at shallow depths. The groundwater is of calcium bicarbonate type and generally contains 50 to 150 ppm of total dissolved solids. Locally iron and manganese concentrations are excessive, especially in the Meguma Group of Nova Scotia. In the Truro area iron content has been reported up to 14 ppm.

Pre-Windsor sediments comprise all the Mississippian conglomerates, sandstones, siltstones, and shales that lie between the basement complex and the Windsor rocks. Well yields range up to 25 gpm (rarely to 50) and the water is predominantly calcium bicarbonate type containing 100 to 300 ppm of total dissolved solids, except where groundwater has passed through Windsor sediments before entering this unit and then the quality may be poor.

The Windsor Group comprises Mississippian limestones, dolomites, anhydrite, gypsum, salt, shale, and siltstone. Well yields are extremely variable and the water is excessively hard, commonly saline, and often contains greater than 10,000 ppm total dissolved solids. The main constituents are sodium, calcium, chloride, and sulphate.

The post-Windsor sediments comprise the late Carboniferous and Permian sandstones and claystones with some conglomerate and siltstone. Yields of 25 gpm are generally obtained and large capacity wells have been developed in some areas. The Imperial Pollett River No. 1 well reported a flow of 840 gpm fresh water from the Boss Point Formation at 390-foot depth. Four wells at Fredericton drilled to between 187 and 400 feet yield between 150 and 500 gpm. A spring 10 miles northeast of Sussex flows continuously at 700 gpm. Wells at Sackville, Shediac, and Amherst yield between 150 and 670 gpm. The water is generally a good quality calcium bicarbonate type containing from 100 to 300 ppm total dissolved solids. A few isolated occurrences of saline water exist but these are presumed to be due to contact with the salt of the Windsor Group. Salt water intrusion, due to over pump-

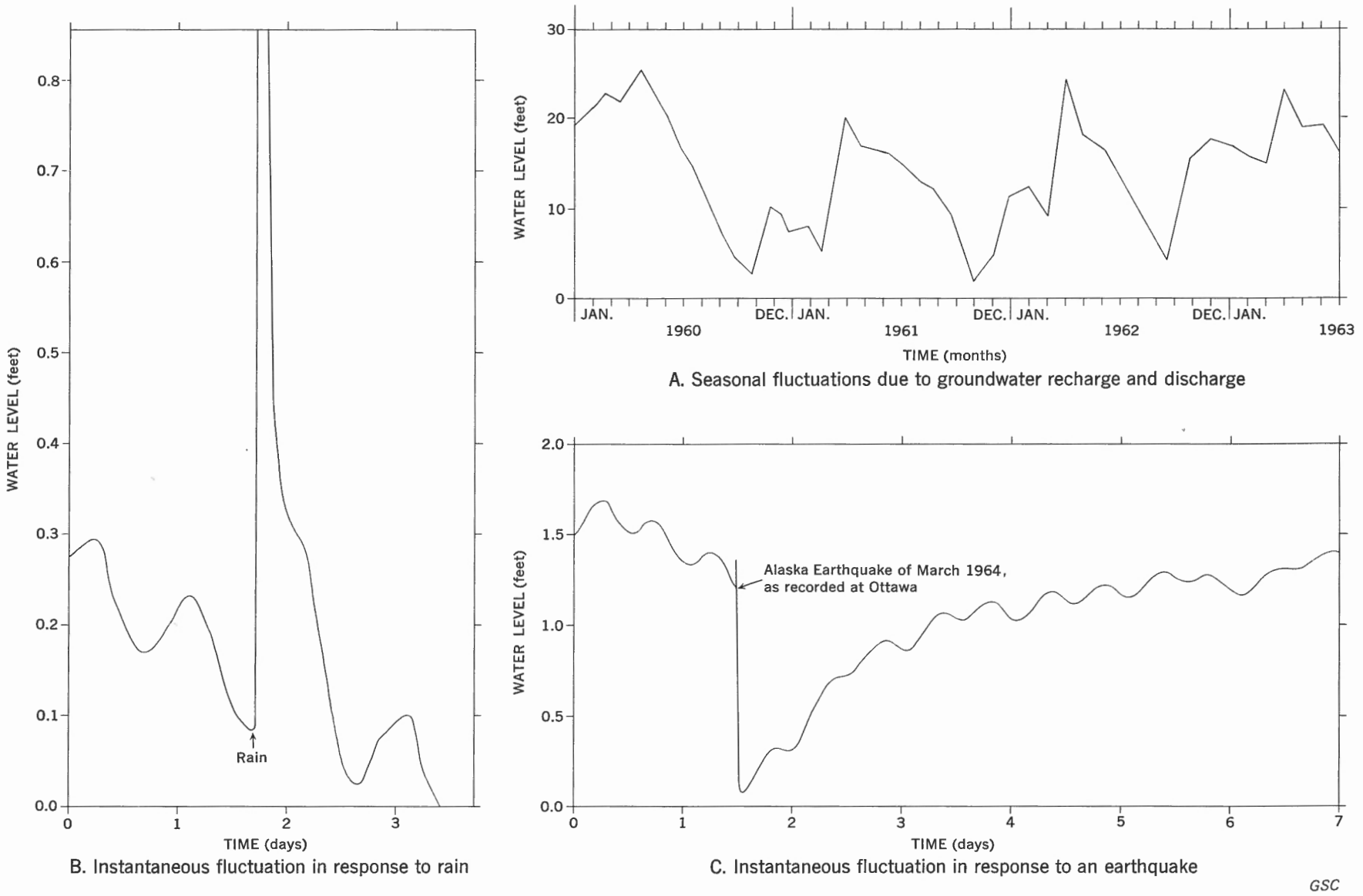
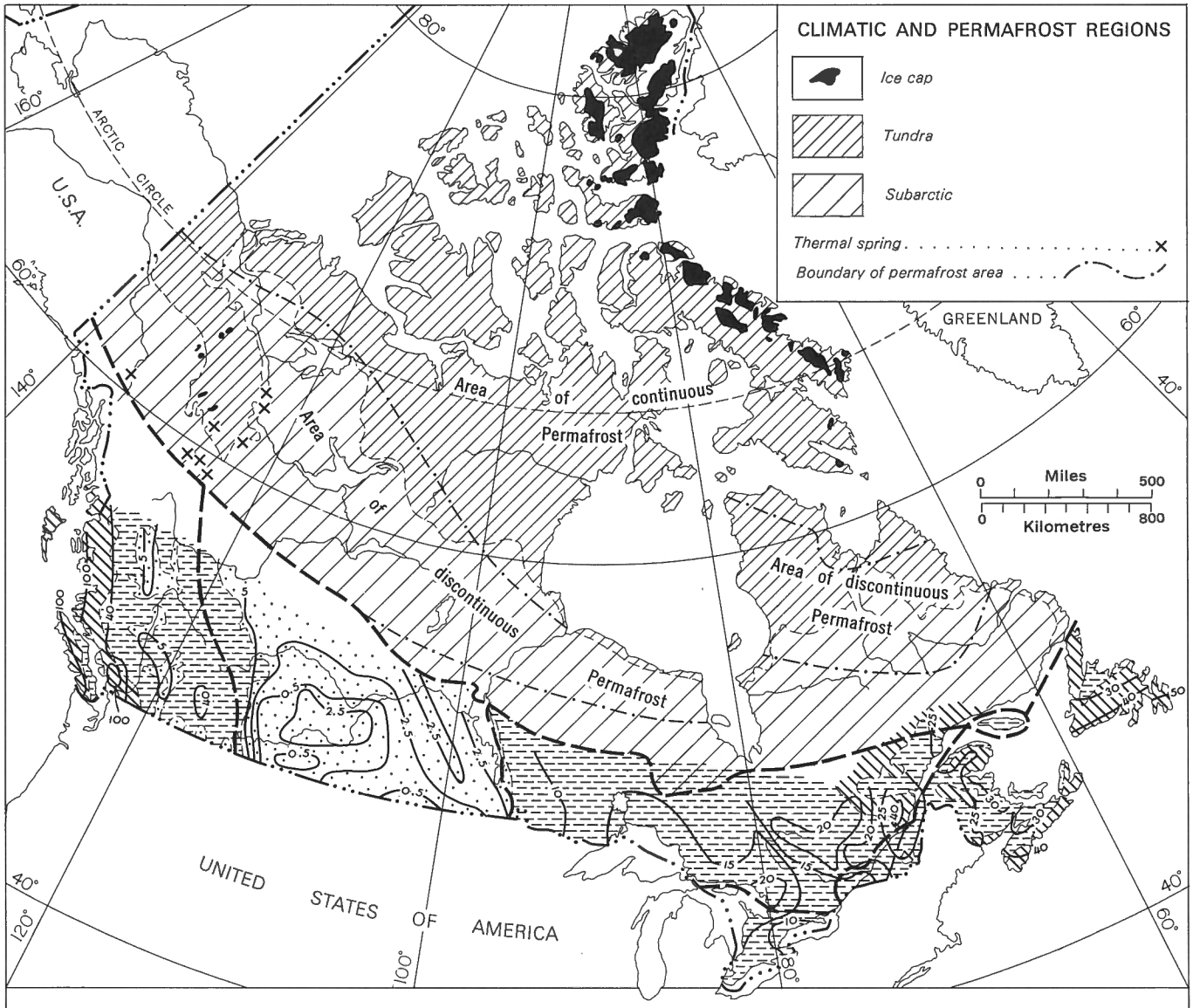


FIGURE XIII-6. Seasonal and instantaneous fluctuations in groundwater levels.

ing, has been induced in some wells adjacent to the coast of Prince Edward Island and New Brunswick.

Triassic rocks comprise sandstone, shale, basalt, and diabase and are probably the best bedrock aquifers in Nova Scotia. The basalt and diabase generally yield less than 3 gpm but yields of up to 400 gpm are obtained from



* Hydrogeological Region	Groundwater	Surface	Total
Cordilleran	8	16	24
Interior Plains	0.12	0.35	0.47
Canadian Shield	3	14	17
St Lawrence Lowlands	4	9	13
Appalachian	6	20	26

* Refer to Figure XIII-1

FIGURE XIII-7. Climatic regions of Canada, permafrost regions and annual run-off in southern Canada.

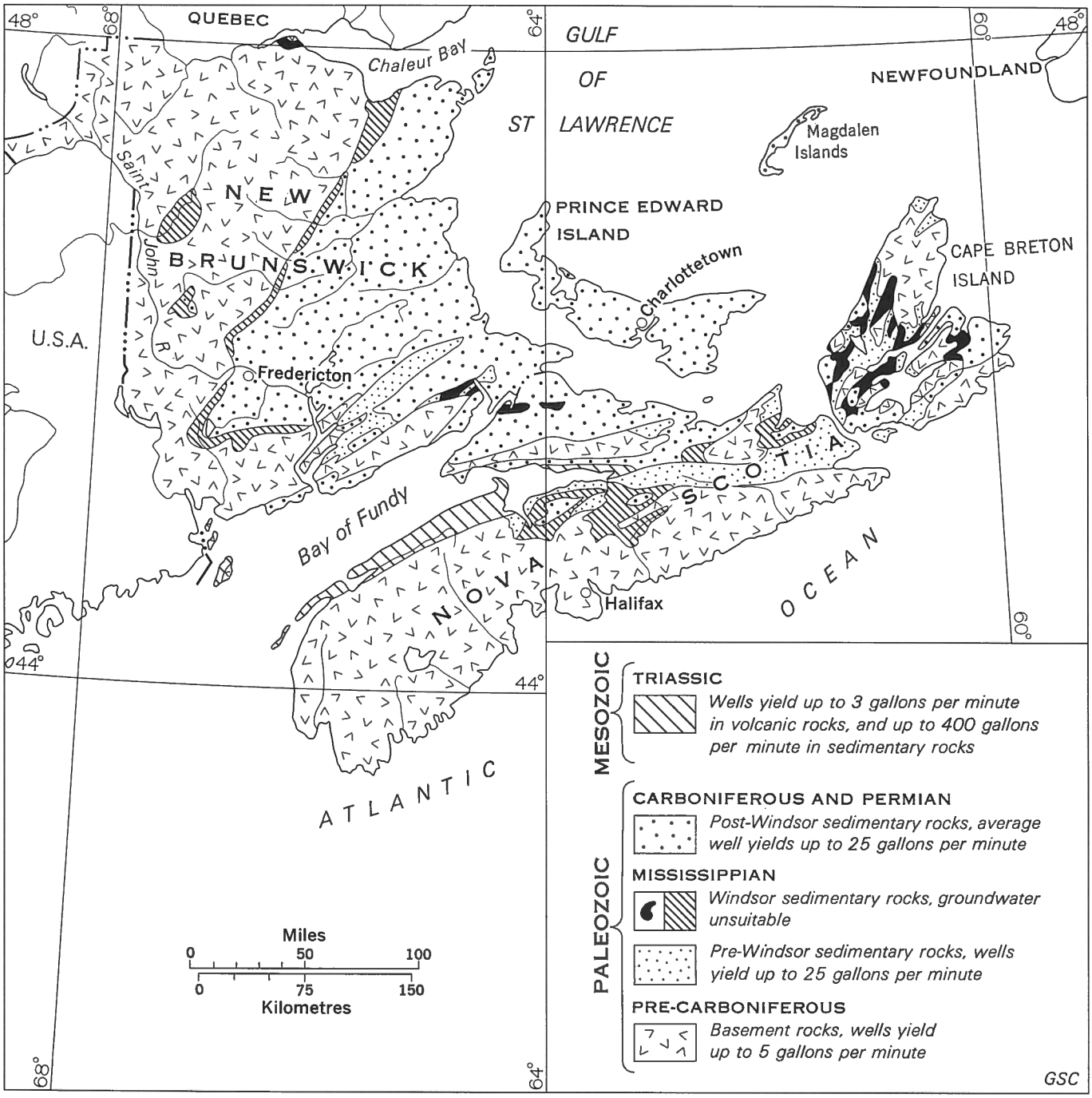


FIGURE XIII-8. Bedrock hydrostratigraphic units in New Brunswick, Nova Scotia, and Prince Edward Island.

the other rocks. The water contains between 50 and 200 ppm total dissolved solids. Some sea water intrusion has occurred at Kingsport, Nova Scotia, where a sample of groundwater contained 1,400 ppm total dissolved solids, mainly sodium chloride.

The Pleistocene sand and gravel deposits occur as eskers, kames, and outwash plains and although of small size and limited distribution are good aquifers. The thin layer of till that covers most of the Maritimes is a poor aquifer.

Insufficient data are available to determine the average yield of surficial materials but the following examples probably indicate the maximum limits. The town of Sussex obtains its water supply from a 12-inch-diameter well rated at 500 gpm with a specific capacity of 17 gpm per foot of drawdown. Three municipal wells at Fredericton, 113 to 146 feet deep and 12 to 18 inches in diameter, had initial capacities between 590 and 1,125 gpm with specific capacities between 32 and 165 gpm per foot of drawdown. Three miles north of St. Stephen a sand and

gravel aquifer yields 1,400 gpm. Groundwater from these outwash deposits generally contains less than 200 ppm total dissolved solids and is normally a calcium bicarbonate water. These aquifers are commonly confined to valleys adjacent to rivers and consequently can rely on induced infiltration to sustain their yields.

Recent alluvium of sand and gravel occurs in valleys of some of the larger rivers such as the Saint John. It is usually derived from Pleistocene glaciofluvial material and its hydrologic properties are similar to theirs. Extensive tidal mud flats surrounding Bay of Fundy have low permeability and contain saline water and consequently are poor aquifers. Beach sands are abundant but little fresh water can be pumped from them before sea water is induced.

Gaspé and Eastern Townships

Only limited information is available concerning the groundwater of this area and none of it has been published. The best aquifers so far developed are in Pleistocene sand and gravel deposits. The Quebec Department of Natural Resources reports that twenty-one pump tests of bedrock wells drilled to 135 feet yielded an average of about 20 gpm with occasional yields as high as 65 gpm, whereas eight pump tests on wells to 55 feet in surficial deposits showed yields averaging 60 gpm with a high of 136 gpm.

Newfoundland

No groundwater studies have been done here to date; most water supplies are drawn from surface resources. Much of the island is underlain by crystalline rocks and the few wells developed in these indicate only low yields suitable for domestic supplies. Surficial deposits, particularly outwash sands and gravels, provide the best aquifers. About 3.6 per cent of the total area of Newfoundland is underlain by potential sand and gravel aquifers. Very large capacity wells at Stephenville have been developed in these materials.

Groundwater Chemistry

Due to the high rainfall, comparatively short groundwater flow systems, and the preponderance of crystalline rocks and outwash deposits groundwater in the Appalachian Hydrogeological Region is generally bicarbonate type low in total dissolved solids. The three main causes of poor water quality are: high iron, manganese, and sulphate content derived from pyrite in some of the pre-Carboniferous rocks especially in Nova Scotia; high calcium sulphate and sodium chloride content derived from evaporitic rocks, especially the Windsor Group; and induced salt water intrusion in the coastal areas where there is over pumping.

Water Balance

The water balance for any simple drainage basin is expressed as $P=E+Rs+Rg$, where P is the total precipitation in inches, E the total evapotranspiration in inches, Rs the total run-off of the stream, and Rg the groundwater discharge directly out of the basin. Preliminary calculations done for three small basins in Prince Edward Island give a general idea of the relative sizes of the quantities involved in the Appalachian Hydrogeological Region: $49''(P)=13''(E)+28''(Rs)+8''(Rg)$.

St. Lawrence Lowlands Hydrogeological Region

The St. Lawrence Lowlands, an area of about 43,000 square miles, is underlain by unfolded Paleozoic rocks with a generally thick cover of surficial materials. Relief is low, seldom exceeding 200 feet. The climate is humid continental, with 30 to 40 inches of total precipitation per year. The Lowlands is divided into three parts separated by Precambrian rocks. The western part, most of southern Ontario, is separated from the central part by the Frontenac Axis. The central part consists of the Paleozoic rocks of the Ottawa area and these are separated from those of the St. Lawrence River area by the Beauharnois Axis at the confluence of the Ottawa and St. Lawrence Rivers. The eastern part consists of Anticosti Island, Mingan Islands, and a narrow strip of the north shore of St. Lawrence River, and is separated from the central part by about 360 miles of the St. Lawrence River.

The hydrogeological characteristics of the region are determined largely by the prevailing climatic and geologic conditions (Fig. XIII-9). The climate is generally milder than that of areas immediately to the north because of the moderating influence of the Great Lakes and the sea. However, one of the major storm tracks on the continent passes through the region and produces rapid variations in weather particularly in the winter.

Hydrogeology

The porosity and permeability of bedrock in the St. Lawrence Lowlands Hydrogeological Region is largely due to fractures of various types, and at present the best method of assessing this is through well-yields in various areas (Fig. XIII-10). The surficial deposits consist of glacial and non-glacial sediments deposited during the last glacial stage of the Pleistocene Epoch except for relatively small areas of Recent sediments that are generally associated with modern streams and rivers. Their aquifer potential is assessed through well-yields as shown in Figure XIII-11.

Ontario

Between the Frontenac Axis and the Niagara Escarpment sedimentary rocks are not good aquifers, though in some areas fracturing allows yields of 10 gpm and more. The most productive rocks are commonly limestone.

Along Niagara Escarpment springs occur where fractured dolomite of Middle Silurian age overlies shales of the Lower Silurian, Clinton Group. The Middle Silurian

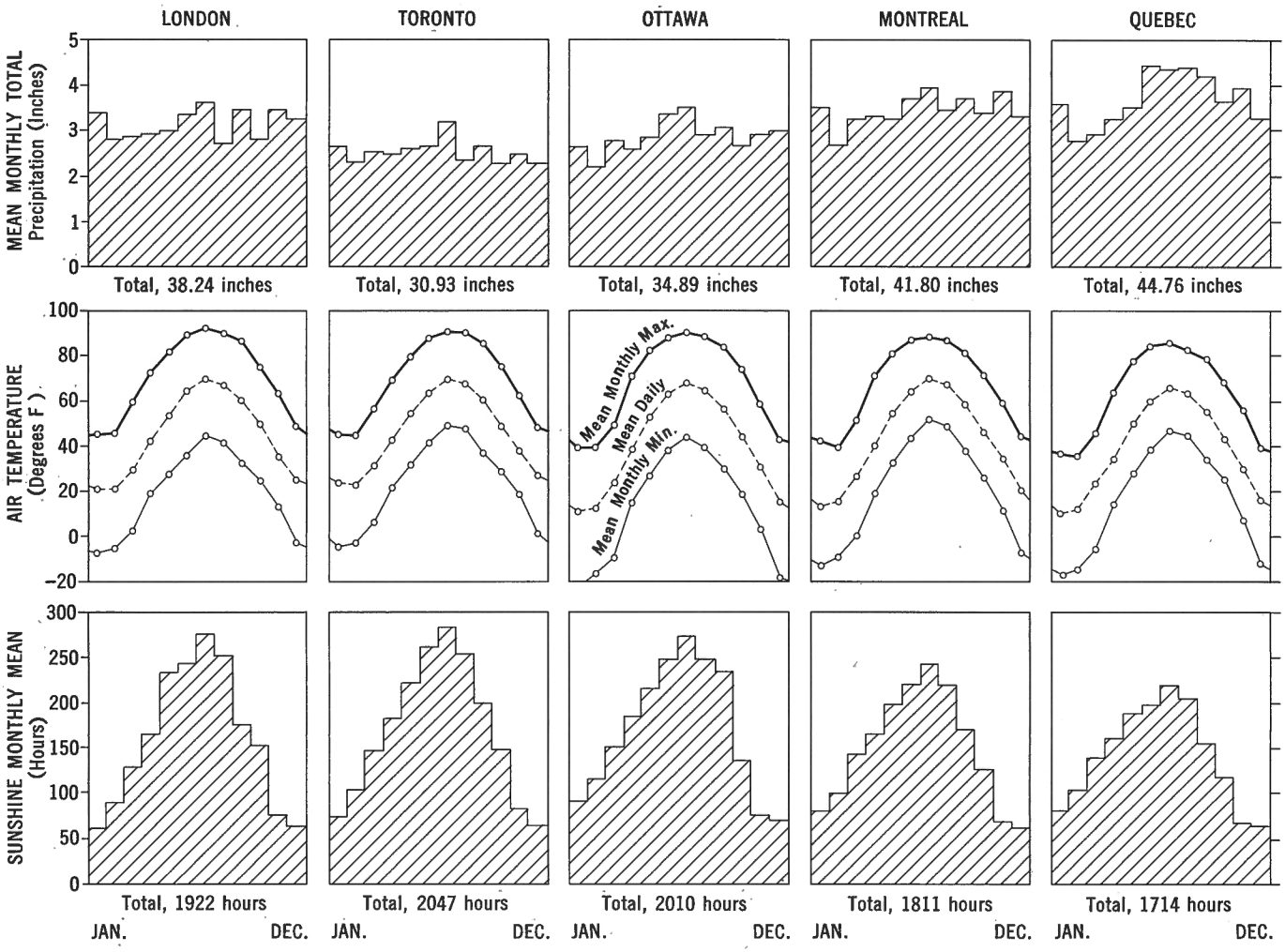


FIGURE XIII-9. Climatic data for stations in St. Lawrence Lowlands Hydrogeological Region.

formations are well jointed and constitute some of the best aquifers in the western part of the lowlands. Wells yield 5 to 15 gpm and locally in excess of 600 gpm. The Salina Formation is a poor aquifer, as the gypsum, anhydrite, and salt produce highly mineralized water. The best aquifers in the Devonian are the Bois Blanc and Detroit River Formations. Well-developed joint systems, thin to medium bedding, and some cavernous reef structures provide well-yields averaging 10 to 30 gpm. The water is usually of good quality with a tendency to become more highly mineralized at depth.

East of Frontenac Axis in the Ottawa area the oldest Paleozoic rocks are the Cambrian sandstones which directly overlie Precambrian crystalline rocks. They are generally a good source of groundwater, particularly near Kingston and Ottawa. Yields of ten to several hundred gpm are obtained from jointed and fractured zones in the upper and lower parts of the formations. Overlying the sandstones are thick sequences of Ordovician limestones and dolomites with some shale and sandy zones. Fracture zones developed in these rocks provide yields of 5 to 15 gpm, though most are below 10 gpm. Water is generally of good quality, but hard. The brown and black pyritiferous shales, however, may yield slightly larger quantities of water but highly mineralized and sulphurous.

Most coarse granular surficial deposits that are significant aquifers in the western part of the region were produced during the northward recession of the glaciers when meltwater was most active in transporting and sorting sediments. The confluence of two main glacial lobes is marked by thick elongate kame moraines composed mainly of sand and gravel in the area north of Lake Ontario and between Orangeville and London. These form some of the most extensive and important aquifers in the region. The finer grained sediments deposited in the offshore parts of the lakes marginal to the glaciers generally have little potential as aquifers. However, the more extensive shoreline or near shoreline deposits—beaches, beach bars, and spits—may be important local aquifers.

East of the Frontenac Axis most sand deposits now found on surface originated either as Champlain Sea beaches or as high-level terraces formed during the early stages of the Ottawa and St. Lawrence Rivers. Average

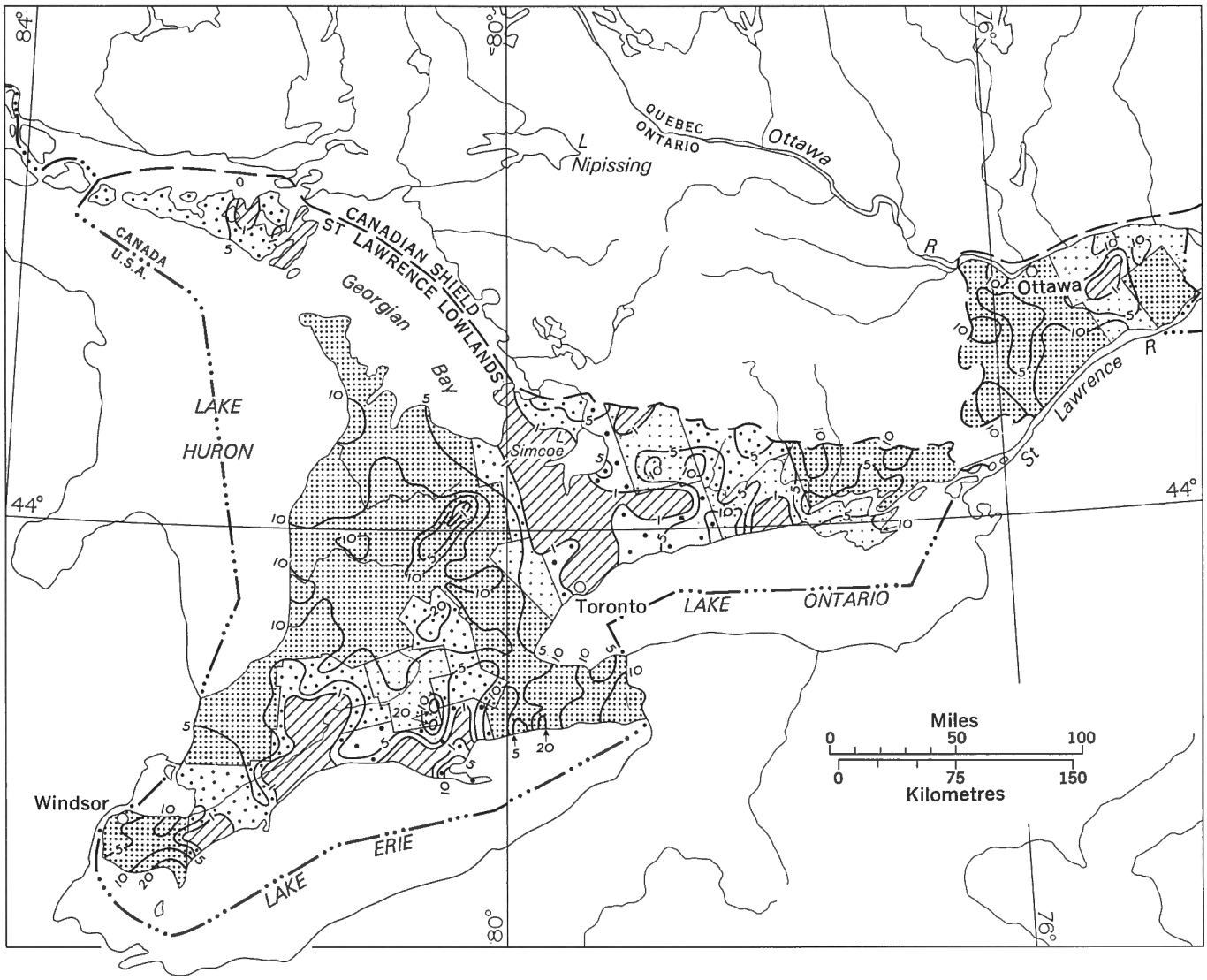


FIGURE XIII-10. Average yield of bedrock wells in Ontario part of the St. Lawrence Lowlands Hydrogeological Region.

well-yields from surficial aquifers range from one to thirty gpm. The most productive are the complex of spillway, beach and kame deposits in the area south from Elmira to Lake Erie.

Deep preglacial valleys were eroded in bedrock at various places in Ontario, particularly along the Niagara Escarpment and in the Georgian Bay area. Where filled with permeable materials, the yield is several hundred gallons per minute.

Quebec

Very few published hydrogeological data are available for the central part of the St. Lawrence Lowlands Hydrogeological Region, thus it has not been possible to show

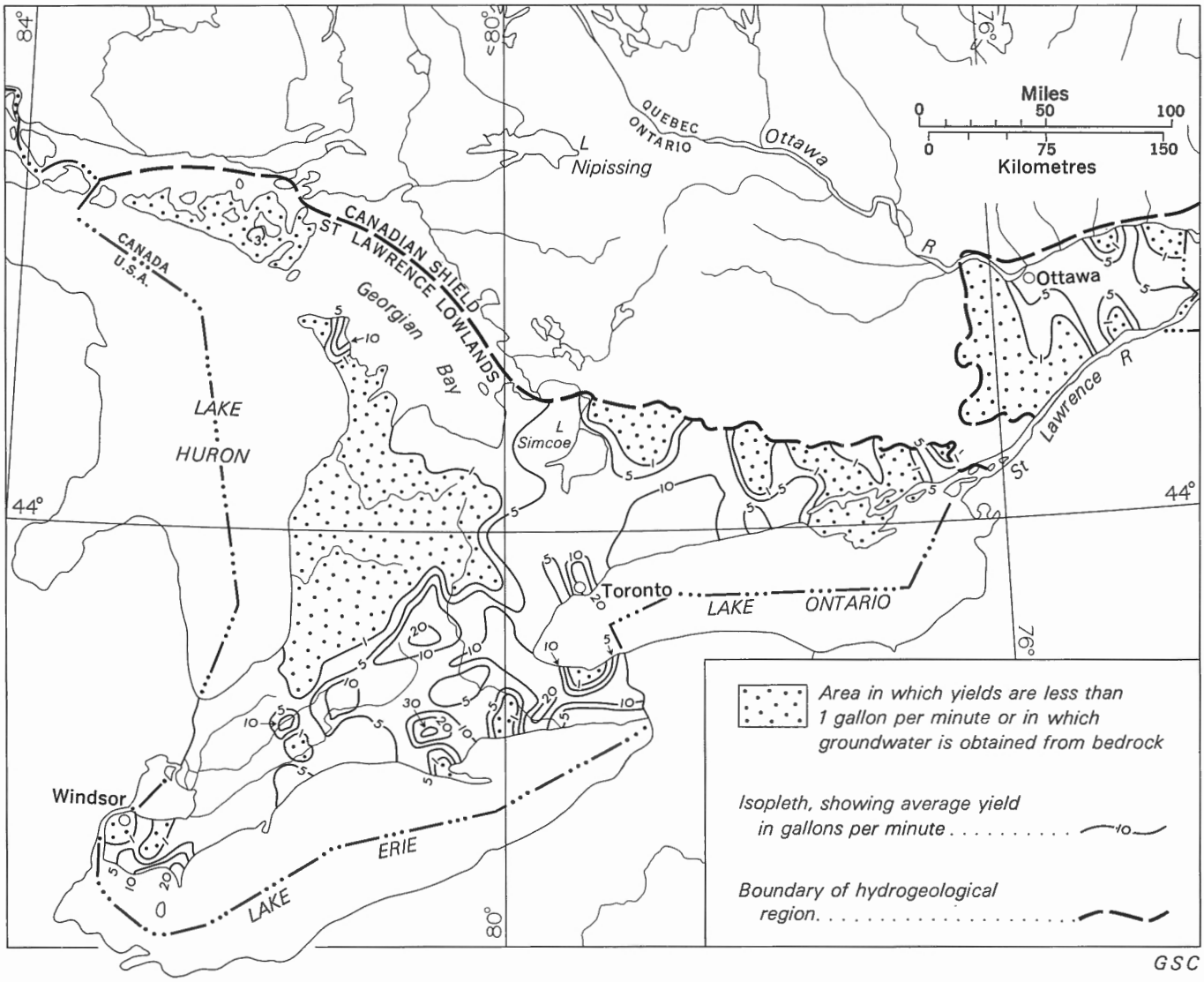


FIGURE XIII-11. Average yield of wells in surficial deposits in Ontario part of St. Lawrence Lowlands Hydrogeological Region.

the distribution of average well-yields for this area.

The Precambrian crystalline rocks are overlain by Cambrian sandstones and conglomerates that constitute one of the best aquifers in the area. Well-yields of more than 400 gpm are known. The overlying Lower Ordovician carbonate rocks are irregularly jointed and fractured so that whereas most of these rocks have comparatively low yields of 5 to 15 gpm, some high-capacity wells have been developed with yields of 70 to 500 gpm. Upper Ordovician strata are mainly shale with minor interbeds of limestone and sandstone, and seldom yield more than local domestic supplies.

The Paleozoic sediments were intruded during Cretaceous time by plug-like masses of alkaline igneous rocks forming the Monteregian Hills. Local aquifers, formed by fractures in the zone of contact between the intrusive and sedimentary rocks, yield 50 gpm.

Generally the oldest glacial deposit is till lying on the bedrock surface. In some places, however, as in the Becancour area, stratified glacial lake sediments are found below the till. Above the till are stratified sediments of non-glacial origin overlain in turn by younger till and thence stratified deposits of drift, kames, and outwash. Associated with these deposits, particularly along the northern margin of the region, is glacial outwash of sand and gravel. The coarser and better sorted materials are potential aquifers and because of the interlayering of permeable and impermeable materials artesian conditions are likely to be encountered.

Overlying the glacial materials are marine and brackish water silts, clays, and minor sands that were deposited by the Champlain Sea. Associated with these are terrace and alluvial sands in which wells yielding up to 1,000 gpm have been developed.

Buried valleys, probably tributaries to the preglacial river system, have been detected beneath the drift by

detailed seismic surveys in the area southeast of Rigaud and in the Beauharnois and Chambly areas south of Montreal. Wells yielding several hundred gallons per minute have been developed in the buried valley near Rigaud.

Anticosti Island

Anticosti Island is underlain by about 2,400 feet of Ordovician and Silurian rocks, but as the water supply for the local population is obtained entirely from surface sources no water well data exist. The presence of groundwater is indicated by perennial streams which derive much of their baseflow from groundwater discharge.

Above an elevation of 250 feet there is only a thin veneer of till with no groundwater potential. Below this elevation a succession of wave-cut terraces mantled by stratified marine sediments occur; where these are sand and gravel it is probable that groundwater supplies could be developed.

Groundwater Chemistry

Throughout most of southern Ontario the groundwaters are calcium and magnesium bicarbonate water, which reflects the chemistry of the limestone and dolomite aquifers. Similarly a zone extending from Caledonia northwest to Southampton is characterized by water high in calcium sulphate due to the underlying gypsiferous Salina Formation. Areas underlain by black shale yield water containing sodium bicarbonate and hydrogen sulphide. Sodium chloride is a common chemical constituent of bedrock waters and is generally the dominant component in waters from the deep bedrock aquifers.

In the western part of St. Lawrence Valley in Quebec groundwater from bedrock shows the same association of bicarbonates of calcium and magnesium with limestone aquifers, high sodium bicarbonate, and hydrogen sulphide from shales.

In some places throughout St. Lawrence Valley groundwater has a high content of sodium chloride. Some of this may be derived from the leaching of salt in the extensive deposits of marine clays in the area. Although this water is generally undesirable for domestic water supplies, spring waters of the sodium chloride type containing various trace elements are marketed for medicinal purposes.

Canadian Shield Hydrogeological Region

The Canadian Shield region is underlain almost entirely by mixed crystalline rocks with only minor areas of sedimentary rocks. Surficial materials are distributed very irregularly with little or none in the upland areas and thick deposits largely confined to the valleys. The topography is extremely rugged though relief is seldom more than a few hundred feet except for a few areas where it exceeds 1,000 feet. The climate is humid continental, and total annual precipitation (much as snow) varies from 22 inches in the west near the Interior Plains to 44 inches in the east near the Gulf of St. Lawrence. In the western part about one third of the mean annual total precipitation falls as snow and in the eastern part nearly one half. This precipitation is ample to fulfil, at least theoretically, the recharge needs of all groundwater basins, so except for local areas of high pumpage these can always be considered to be at their seasonal maximum.

No well inventories or groundwater studies have been published for the region. Available data are, therefore, scanty and deductions must be drawn from known geology, comparison with other areas, and information from the numerous mines.

Because of the abundance of good surface water, groundwater is not widely used. Only some 20 per cent of the total population, including rural, uses groundwater. Approximately 150 million gpd of water (not including industrial supplies) are used by municipalities and of this an estimated 11 million gallons are groundwater.

Hydrogeology of Surficial Deposits

The best sources of groundwater are the coarser grained surficial materials deposited by meltwaters from the glaciers. Because the crystalline rocks are resistant to erosion these materials are commonly found along the already existing valleys in the bedrock.

Large areas of the Shield region were covered by glacial lakes, such as Barlow-Ojibway, and the clays deposited in these lakes are poor sources of groundwater though they may overlie more permeable material. Some wells have been developed in these materials. It is possible that large scale water supplies could be developed from aquifers of this type.

Over much of the Canadian Shield bedrock is covered by ground moraine consisting basically of clay till and it is unlikely that other than domestic supplies could be obtained.

Hydrogeology of Bedrock

For practical purposes all groundwater occurs in fractures in the crystalline rocks and as their distribution and size vary widely the groundwater potential varies accordingly. Small supplies generally can be obtained, but the large quantities needed for industries and towns can be found only in exceptional areas.

The quantity of water that can be obtained can be estimated by considering the quantities of water pumped from mines in the Canadian Shield. For example, at Hollinger Consolidated Gold Mines, Limited almost 2 million gpd is pumped on the average and nearly double this amount in the spring when recharge is at a maximum. It would appear from this figure that large groundwater supplies are available. Hollinger workings, however, cover about 920 acres to a depth of more than 5,000 feet, and as the size of the cone of depression surrounding this

withdrawal area is unknown the total area actually supplying this water is also unknown.

Crystalline rocks can also be virtually impermeable as illustrated by Sullivan Consolidated Mines, Limited whose workings are situated beneath Lake Dubuisson and extend to within some 100 feet of the lake bottom, yet only 200,000 gpd is pumped from this mine.

Mining experience shows that fractures are present and numerous at depth but that in general they are tighter and carry less water. Unfortunately little quantitative data are available on this decrease with depth, but three uranium mines at Elliot Lake provide an interesting example. As the orebodies are nearly flat, all the workings except the shaft are at about the same depth. Nordic mine at 1,400 feet pumps some 400,000 gpd whereas Denison at 2,465 feet and Milliken at 3,000 report little water pumped. It should not be assumed that these conditions prevail everywhere, as some large flows of water have been met in mining at great depth but they are much less common than those at shallower depths.

Groundwater Chemistry

Both crystalline rocks and the surficial materials formed from them are composed of nearly insoluble minerals, therefore, in general, groundwater in the Canadian Shield is similar to the acidic surface waters and has a similar low pH. It is notable that although both the organic content and the stream stage fluctuate widely the chemical content remains nearly constant throughout the year, indicating that groundwater discharged to streams is similar in composition to the surface water and suggesting strongly that groundwater flow systems are comparatively short and shallow.

Poor quality water may be encountered near some mineral deposits. In most base metal mines the orebody contains sulphides and the groundwater is strongly acidic and corrosive. In most gold mines however sulphides are rare and the groundwater is potable.

Interior Plains Hydrogeological Region

The Interior Plains Hydrogeological Region is that part south of the southern limit of discontinuous permafrost. Except for a narrow folded belt along the western edge, the entire area is underlain by nearly horizontal strata of Paleozoic, Mesozoic, and Tertiary age. It is almost entirely covered by a thick layer of surficial deposits and bedrock outcrops are rare except along deeply incised river valleys. Elevation varies from about 4,000 feet in the west to 500 feet along the eastern edge. Most of the surface is rolling with relief of 100 to 200 feet. Exceptions are the very flat glacial lake basins. The climate is humid continental, except for the area between Moose Jaw, Lethbridge and as far north as latitude 52°N where it is semi-arid. Total annual precipitation varies from 10 to 20 inches, with the eastern and western parts of the region receiving the greater amount.

Hydrogeology of Surficial Deposits

Surficial deposits fall into three categories: *till,* an unsorted mixture of sand, silt, clay, and boulders; *lacustrine sediments,* mainly silt and clay; and *outwash deposits,* deltaic sands and gravels or blanket outwash sands. The regional distribution of these is shown on the Glacial Map of Canada, and indicates that 60 per cent of the plains is covered by till, 40 per cent by lake deposits, and that outwash deposits cover less than one per cent.

The till is commonly calcareous, its texture ranging from loam, clay loam to sandy loam. Over-all thickness is about 75 feet but increases markedly from west to east, and thicknesses of 1,000 feet have been measured. Near surface, the permeability ranges from 0.3 to 0.5 inch per hour but decreases by a factor of 10 or more in the interval 10 to 100 feet below surface.

An estimated 60 per cent of all farm water supplies in the Prairie Provinces is drawn from the till, so it is an important aquifer even though shallow wells yield less than 5 gpm and the till cannot be relied on for large sustained yields. The most reliable yields are from stratified sand and gravel deposits occurring within the till. Their aquifer potential depends on their areal extent. Such intertill aquifers near Regina have a combined safe yield of nearly 20 million gpd.

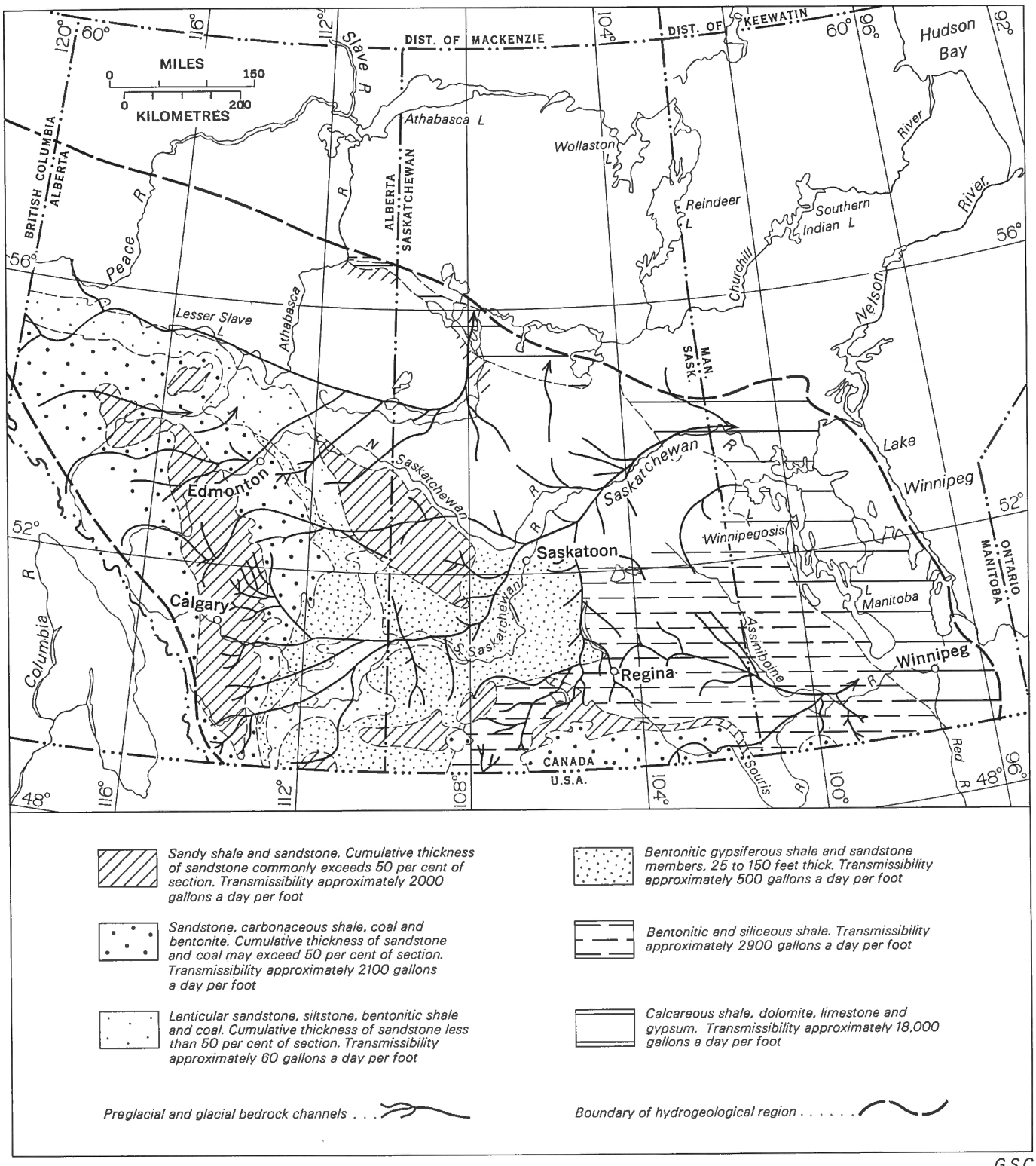
Most of the till occurs as ground moraine with a flat to gently rolling surface with relief of 5 to 30 feet. Owing to low permeability of the till and the presence of more permeable strata in the underlying bedrock, groundwater generally moves directly downward. Annual recharge in these areas is about 0.5 inch per year. A typical landform of many topographically high areas is hummocky moraine. These areas contain innumerable small closed depressions in which snow accumulates and would seem to provide the best recharge environment. Infiltration, however, is slow, and phreatophytic plants growing around ponds intercept much of the water and the cone of depression formed around their roots may reverse the downward flow, with the result that the depression becomes a discharge area.

Lacustrine deposits are made up chiefly of montmorillonite clay which is characteristically calcareous, and unoxidized except near surface. Their thickness is about 40 feet but diminishes towards the periphery of the lake basins. Permeability is low, ranging from 0.1 to 0.2 inch per hour near the surface to 0.001 to 0.002 inch per hour from 10 to 100 feet below surface. These deposits are aquicludes and are only locally sufficiently permeable to be used for domestic supplies.

The effect of lacustrine deposits in areas where groundwater discharge is restricted to deeply incised river valleys is to retard recharge over the lake plain. Where groundwater discharge is more diffuse lacustrine clays retard discharge and this may result in artesian conditions, as in Red River valley.

Outwash deposits are found along the shores of the

former glacial lakes and as alluvial sands at the mouth of meltwater channels. They consist chiefly of well-sorted sand and fine gravel, and are commonly less than 15 feet thick, though some are more extensive and may be 50 or 100 feet thick. Their permeability is 1 inch to 10 inches per hour regardless of depth. Though limited in extent their aquifer potential is the most important of the prairies. Industrial wells on the Carberry sand plains of Manitoba



GSC

FIGURE XIII-12. Bedrock hydrogeological divisions and buried valleys of the Interior Plains Hydrogeological Region.

produce over a million gpd without apparent depletion during the few months of the year that they are in use.

No principal elements of the drainage system that existed prior to or during the last glaciation of the prairies have been reconstructed from subsurface information (Fig. XIII-12). The valleys may be either partly or completely filled with glacial drift or with outwash and alluvial deposits. They may contain important aquifers if filled with permeable materials and if their base-level is low enough that they do not drain into present streams. Buried valleys that contain coarse deposits may show as a depression in the water-table or in the piezometric surface indicating that they act as subdrains. They tend to be broad, commonly between 2 and 10 miles wide with gently sloping sides and a depth of 100 to more than 300 feet. Groundwater therein characteristically contains less total dissolved solids than that in the surrounding till.

Hydrogeology of Bedrock

The nature of the upper 500 feet of sedimentary formations is such that its potential as aquifers is better illustrated in terms of transmissibility and lithology than by the distribution of formations. Such bedrock hydrogeological divisions are presented in Figure XIII-12.

As much of the sedimentation in the Cretaceous occurred under non-marine conditions, none of the Cretaceous formations contains extensive uniform aquifers. Nevertheless, many sandstones of limited areal extent have sufficient permeability to be useful aquifers. Regionally, the cumulative thickness of sandstone beds in the stratigraphic column and their individual permeabilities decrease from west to east and consequently aquifer conditions also become poorer. Most bedrock aquifers appear to be confined, that is they are not intersected by the phreatic surface, but under conditions of continuous pumping many sandstone aquifers receive leakage from overlying silty shale so it is a matter of relative permeability rather than complete confinement. Due to these conditions, areas with flowing artesian wells are common throughout western Canada.

Bedrock aquifers of the Interior Plains are everywhere sufficient to meet domestic water requirements, but sandstone aquifers that can yield 50 gpm or more are limited to the Paskapoo Formation, parts of the Edmonton Formation, the Birchlake and Ribstone Creek Members of the Belly River Formation, and perhaps some parts of the Eastend and Ravenscrag Formations. At present only the Paskapoo Formation and locally the Edmonton Formation are tapped for municipal use and the production of 125,000 gpd (85 gpm) appears to be about the ultimate safe yield of these formations.

Although the hydrologic divisions of the prairies make it appear that no good aquifers can be found east of Last Mountain Lake, the siliceous shale of the Riding Mountain Formation is sufficiently fractured to yield from 5 to 50 gpm. The Paleozoic limestones underlying the Red River valley have high transmissibilities believed to be due to a relatively thin solution channel and fracture zone parallel to the surface.

Groundwater Chemistry

Surficial Deposits

Groundwater acquires its principal chemical composition during infiltration in the soil zone. The net amount of downward flow is the balance of infiltration and evaporation. In temperate climates the principal process is dissolution whereas in semi-arid climates precipitation of salts in soil is of greater importance as a result of more salts being deposited by evaporation of capillary water than are removed by the infrequent rainfall. In areas of light rainfall followed by strong evaporation water that finally reaches the zone of saturation will be more mineralized than water that infiltrates in areas of temperate climate. Thus, rainfall and temperature are major factors in establishing the climatological macrozonation of groundwater chemistry. This zonation is best expressed by the concentrations of the most soluble salts, principally calcium sulphate ($CaSO_4$) and sodium chloride (NaCl) and it is shown in the Interior Plains Hydrogeological Region by the chemical zonation of low-flow river water (Fig. XIII-14).

In western Alberta where rainfall is more than 18 inches groundwater in surficial deposits contains less than 800 ppm of total solids, predominantly calcium-magnesium bicarbonates. In east-central Alberta and central Saskatchewan where rainfall is less than 16 inches the total dissolved solids increase to 2,500 ppm. The predominant constituents are sodium, calcium, and magnesium sulphates. In southwest Manitoba the average total dissolved solids diminishes to 2,000 ppm and in south-central Manitoba to 1,100 ppm. There rainfall is slightly greater, with calcium and magnesium sulphates being the predominant dissolved salts. In central and eastern Manitoba rainfall is considerably greater and the amount of total dissolved solids falls to 912 ppm with calcium and magnesium bicarbonate dominating. The chemistry of shallow groundwater on the prairies maintains this consistent relationship generally because of the uniformity of the surficial deposits. This simple relationship is modified however where the groundwater infiltrates into the less chemically uniform bedrock.

Shallow Bedrock

The hydrochemical zonation of the Upper Cretaceous and younger shallow bedrock aquifers is shown in Figure XIII-13. Most of this zonation appears to be due to chemical interactions between the sulphate or bicarbonate drift waters and typical bedrock materials such as lignite, coal, methane, bentonite, and selenite.

The sodium bicarbonate water of southern Alberta is believed to be due to the combined effects of three processes that produce the highly alkaline bedrock waters in all formations that contain coal, lignite, and bentonite.

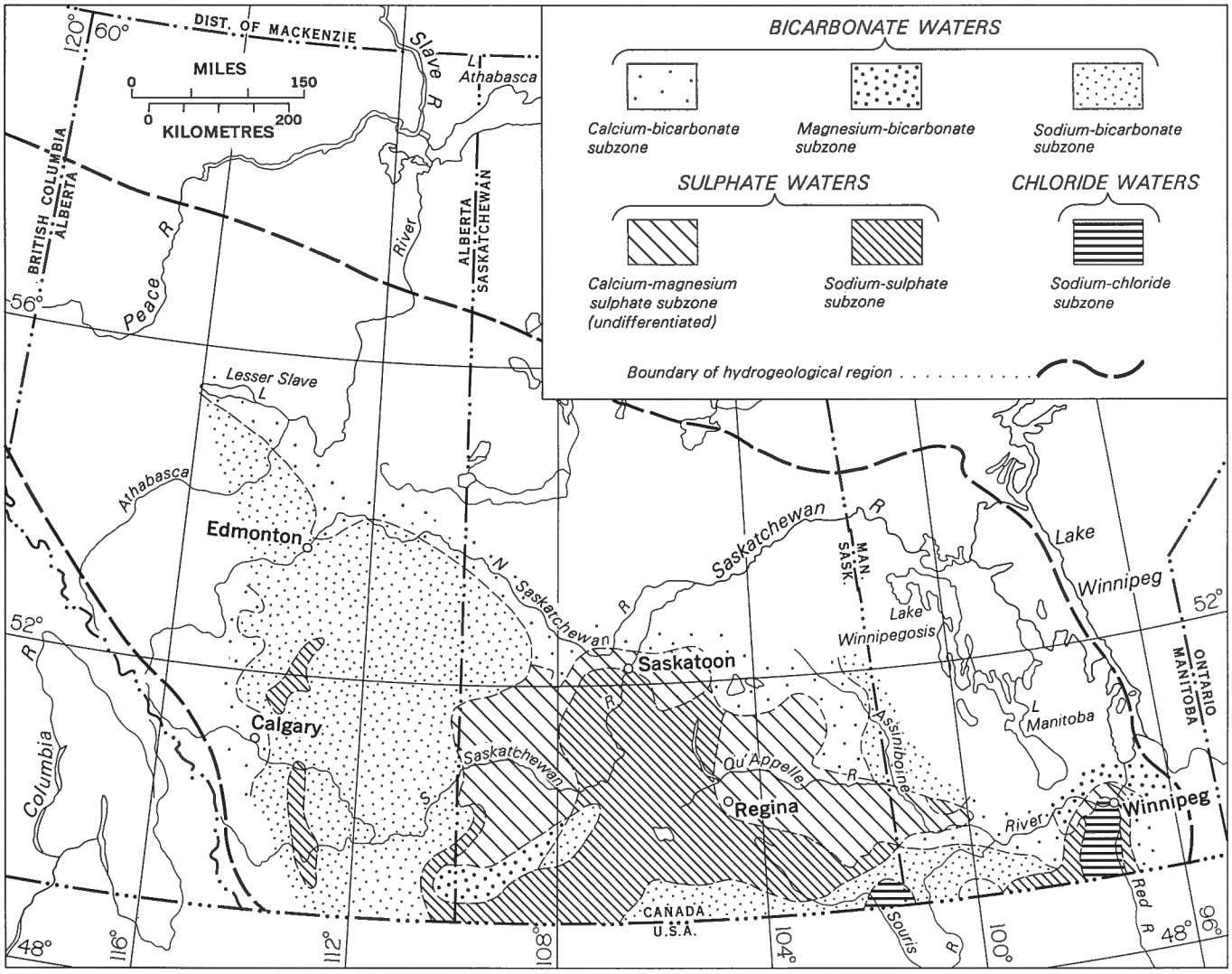


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FIGURE XIII-13. Hydrochemical zonation of shallow bedrock aquifers, Interior Plains Hydrogeological Region.

The shallow calcium-magnesium sulphate water or sodium sulphate water of the surficial deposits may be changed to bicarbonate water by bacterial sulphate reduction in the soil zone. Secondly, the large amounts of carbon dioxide present in the coal measures result in above normal bicarbonate content of water penetrating them. Thirdly, base exchange with bentonite replaces calcium in solution by sodium thus increasing the sodium content of the water.

In Saskatchewan, areas of gypsiferous shale contribute to the sulphate character of the drift water and produce a distinct zone of calcium-magnesium sulphate water surrounding the central area of sodium sulphate water in the marine shales.

The magnesium bicarbonate zone in the Cypress Hills area is due to higher rainfall. The sodium bicarbonate waters in the Shaunavon and Riding Mountain areas are probably related to base exchange with the bentonitic shales.

The occurrence of sodium chloride water at shallow depth is uncommon on the prairies, except in the limestone aquifers of the Manitoba Lowland where it may be related to very saline water from deeper formations being discharged into Red River valley.

Deep Bedrock

The study of data from the deep bedrock, Lower Cretaceous and older, has just been started and only one major feature is evident.

Of all the formation waters studied the highest concentrations of total dissolved solids, up to 300,000 ppm, are found within the deeper parts of the original depositional basins, the Alberta Trough, the foothills belt, and the Williston Basin. The lowest concentrations, often below 500 ppm, occur near the Alberta Trough and over the Sweetgrass Arch. In areas of high concentration the total dissolved solids increase rapidly with depth. For example, an increase of from 107,000 ppm to 256,000

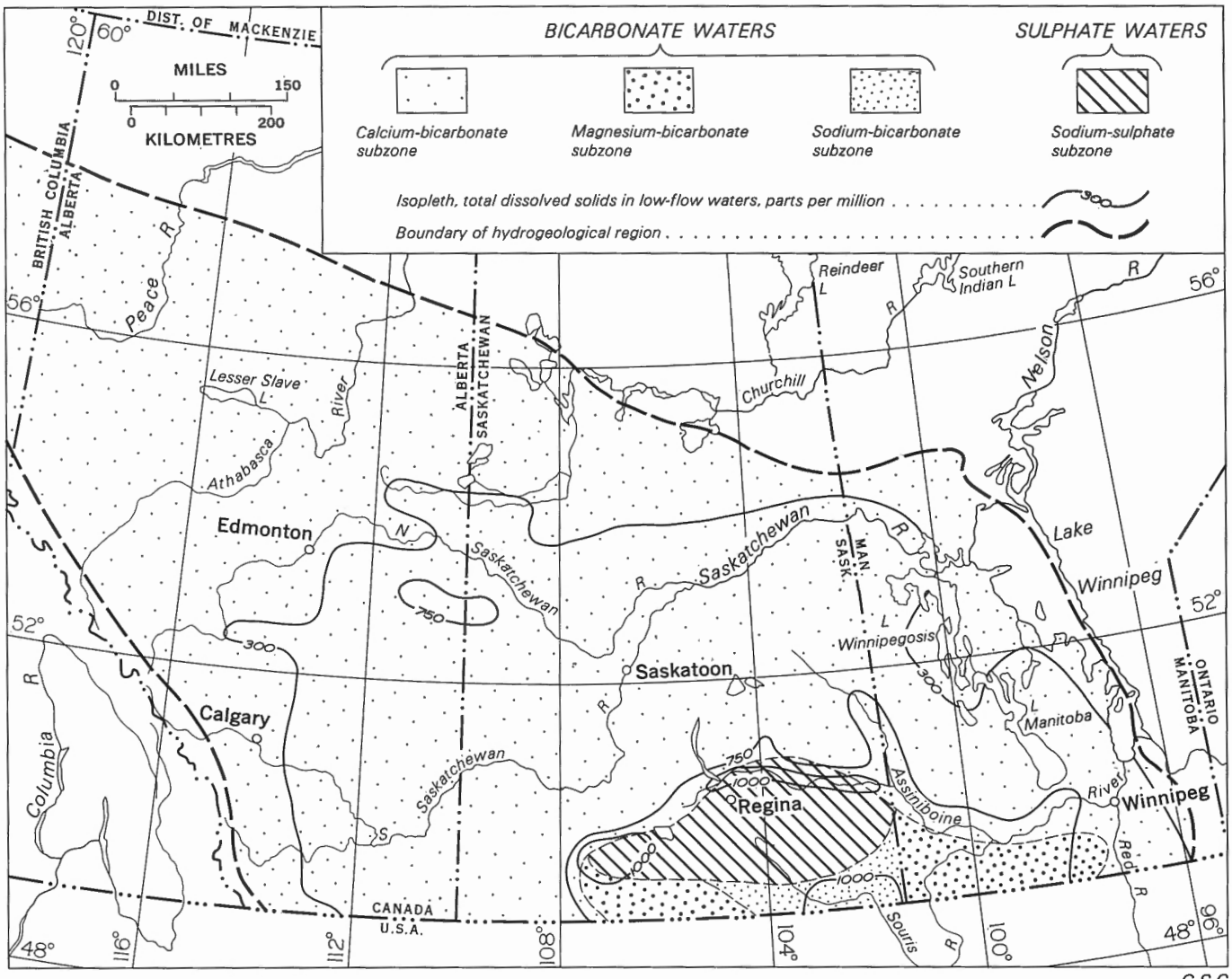


FIGURE XIII-14. Hydrochemical zonation of low-flow river water, Interior Plains Hydrogeological Region.

ppm over a vertical distance of 75 to 100 feet is found in the Frobisher area. In general, these high values are probably related to depth below surface, distance from the outcrop, and proximity to Devonian halite deposits.

Brine Springs

Numerous brine springs containing more than 25,000 ppm total dissolved solids issue from the Paleozoic strata near the Canadian Shield and are in strong contrast to the sulphate-dominated Interior Plains water. The springs fall into two groups. Those in the light grey upper Manitoban limestone contain less than 40,000 ppm NaCl; those in the lower Manitoban limestone and the Winnipegosis dolomite contain more than 40,000 ppm NaCl.

Temperature measurements suggest that the brines originate in an even larger stratigraphic interval. Along the west shore of Lake Winnipegosis there is a 10 degree difference in average brine temperature between the two groups which would indicate a difference in depth of origin of about 500 feet.

Brine springs possibly similar to those in Manitoba are present in the northern Interior Plains along Slave and Mackenzie Rivers.

Water Balance

The order of magnitude of the amount of annual groundwater replenishment has been determined from hydrometric records using Wundt's method of analyses. This is based on the assumption that most groundwater discharge from an area takes place in the rivers and that in a steady state system the amount of discharge is equal to the amount of recharge. Hydrometric observations do not account for underflow or for evapotranspiration but merely give an index of groundwater recharge from which an absolute value has to be estimated. The amount of natural groundwater replenishment ranges from 1.8 inches

to 0.18 inch per year, the latter value being the most common.

Minimum stream flow is taken as the measure of groundwater discharge and a distinction can be made between bank storage, contact springs, and artesian leakage. About 75 per cent of the low flow of prairie rivers is made up of bank storage, a figure also suggested by study of the hydrochemistry of low-flow river waters (Fig. XIII-14). The main type of low-flow water on the prairies is calcium-magnesium bicarbonate water, which cannot originate from the predominantly sulphate shallow surficial groundwater but occurs where flood waters infiltrate the river banks and return to the rivers comparatively unchanged. Groundwater chemistry is reflected in the internal drainage basin of Old Wives Lake and in the headwater regions of Qu'Appelle and Souris Rivers.

Assuming that 70 per cent of the total base flow is derived from bank storage, the index of groundwater discharge from Pleistocene and bedrock aquifers in excess of underflow and evapotranspiration is within the range 0.5 inch to less than 0.05 inch per year. Also assuming underflow and evapotranspiration are twice natural groundwater replenishment then the estimated total groundwater recharge, including recharge to bank storage, varies from 3.6 inches to 0.36 of an inch per year. Recharge to Pleistocene and bedrock aquifers alone amounts to 30 per cent, that is 1.08 inches to 0.1 inch per year. Thus total recharge in the areas of 12 to 16 inches precipitation is estimated to lie within 3 to 22 per cent of the annual precipitation, and recharge to Pleistocene and bedrock aquifers alone varies from 7 to less than 1 per cent of the total precipitation.

Cordilleran Hydrogeological Region

This region is largely underlain by crystalline rocks and steeply folded sedimentary rocks with some flat-lying sedimentary and volcanic rocks in the central plateau and coastal areas. Surficial deposits are confined largely to the valleys and lowland areas. Relief is many thousand feet in the more mountainous belts but only a few hundred feet in the central plateau except where valleys have been cut below this general surface. Total annual precipitation varies from less than 10 to more than 100 inches and extremes can occur a few tens of miles apart. The total annual precipitation on the region is estimated at 800 million acre-feet, whereas the run-off for the same area amounts to 650 million acre-feet. Some 150 million acre-feet are evaporated or represent change in storage in glaciers and as groundwater.

Because of the extreme variability of geology, topography, and precipitation the main hydrogeological characteristic of this region is its variability.

Physical Features and Climate

The Cordilleran Hydrogeological Region includes three major physiographic subdivisions: the Western, Interior, and Eastern Systems (Fig. XIII-15).

The *Western System* includes the St. Elias and Coast Mountains with elevations of from 13,000 to more than 19,000 feet, the Queen Charlotte and Vancouver Island Ranges, and the Cascade Mountains in which relief varies from 3,000 to 7,000 feet. The higher mountains nourish Alpine glaciers and ice fields. Within this system the Georgia Basin is bordered by the Nanaimo Lowland on the west and the Georgia and Fraser Lowlands on the east. These comprise the humid coastal belt. The *Eastern System* includes the Rocky Mountains and the Rocky Mountain Foothills belt and contains peaks of more than 11,000 feet in elevation that support Alpine glaciers. The *Interior System* comprises several major and minor mountain ranges and plateaux including the Columbia Mountains and the extensive Interior Plateau.

Climate is dependent to a large extent on these physiographic systems as air masses that originate in the Pacific and Arctic are forced to ascend over the high mountains of the Western System and in so doing are cooled and loose their moisture as heavy precipitation along the coastal belt. Precipitation averaging 60 to 100 inches but as much as 251.3 inches has been recorded at Henderson Lake at the head of a valley off Barkley Sound. After crossing the mountain barrier dryer air, sometimes heated, descends on the lee side and produces an interior dry belt where precipitation varies from 12 to 16 inches annually and in the deep valleys to less than 10 inches. Precipitation on the remote high plateau and mountain areas is believed to be somewhat greater but data are lacking. The eastern moving air masses again rise over the Columbia and Rocky Mountains and produce a second rain belt with precipitation in places exceeding 40 inches and again there is a dryer belt in the lee of the mountains, the foothills belt and adjacent parts of the Interior Plains.

Because of high relief and steep slopes run-off is rapid. In some drainage basins records indicate that run-off exceeds precipitation. This is probably due to difficulty of gauging steep streams and to inadequate records of precipitation at high altitudes. Evaporation exceeds precipitation during the summer growing season at all weather stations in the coastal humid belt except Agassiz. With the high run-off and high evapotranspiration the major source for groundwater recharge is the heavy winter precipitation which accumulates as snow in most places and, though chiefly lost by direct run-off on melting, part at least passes through a groundwater storage phase.

Hydrogeology

Most groundwater in bedrock occurs in fractures of all types, though in some local areas rocks have enough intergranular porosity and permeability to allow groundwater storage and movement. In much of the Cordilleran Hydrogeological Region the groundwater potential is unknown (Fig. XIII-15). In the mountainous areas surface water supplies are sufficient to meet most demands and no records are available on the successful completion of

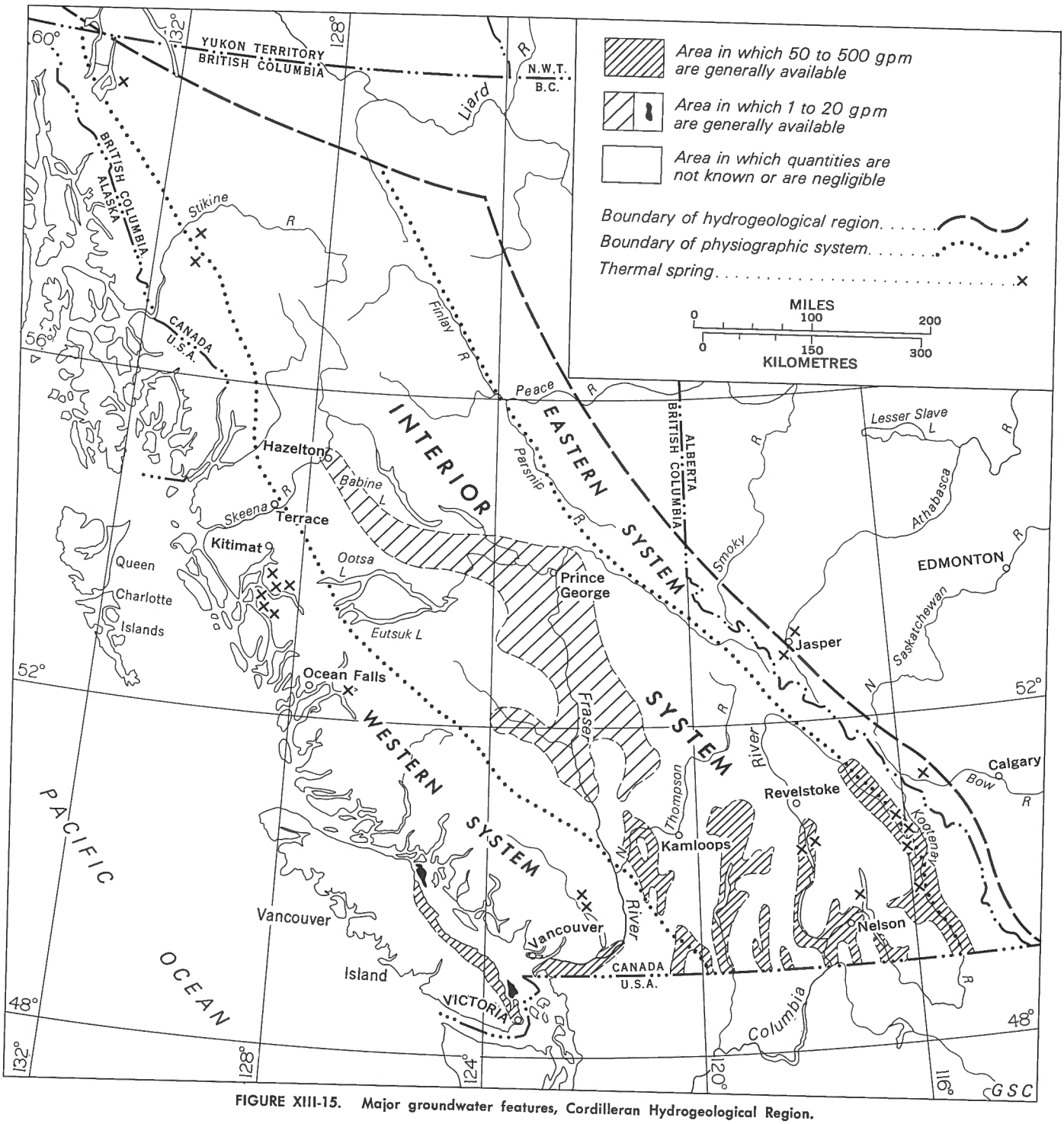


FIGURE XIII-15. Major groundwater features, Cordilleran Hydrogeological Region.

wells in bedrock that yield large volumes of water. All mines in the region are damp and report some groundwater, whereas a few have intersected solution channels or major fracture zones and have had great difficulties with water.

In the sedimentary basins groundwater occurs in the gently dipping or folded sandstones, conglomerates and shales, and drilled wells commonly provide domestic supplies. The operation of coal mines in many of these basins has been curtailed by groundwater inflow, and upon cessation of mining the lower levels were quickly flooded. Plateau lavas in the central part of the interior system are potential aquifers and discharge groundwater by springs at the base of the lavas. These maintain their flow during seasons of little or no run-off. Surficial deposits provide the best natural aquifers but in general are thin on the

upland areas and reach their greatest thickness in the valleys. Glacial, glaciofluvial, lacustrine, alluvial, and wind blown materials mantle the floors and lower slopes of the intermontane valleys. The present rivers are entrenched into these deposits in places to depths of 200 feet or more.

Permeable outwash gravel and sand that is perhaps more than 300 feet thick fills Rocky Mountain Trench. Drillholes have penetrated as much as 250 feet of these deposits without encountering bedrock. Not all valleys are filled with permeable sediments and more than 300 feet of till has been reported filling the Clearwater and North Thompson valleys. At Enderby in the Okanagan Valley test wells have penetrated more than 1,300 feet of unconsolidated sediments. Because of interstratification of permeable and impermeable materials in these deep valley fills, high artesian pressures are sometimes encountered.

In the northern part of the interior system permeable sand and gravel fill valleys to depths of more than 50 feet and cover wide plains. These deposits supply large volumes of water especially along the main transportation routes.

Humid Coastal Lowland

The Fraser Lowland consists of wide relatively flat uplands and wide flat-bottomed valleys tributary to the main east-west postglacial valley occupied by Fraser River. Tributaries to Fraser River are estimated to carry 30 million gpd of water supplied by groundwater during periods of low flow. Unconsolidated surface deposits are commonly more than 600 feet thick and deep test holes have indicated thicknesses of more than 1,000 feet in places and of as much as 1,800 feet at Tsawwassen.

The Pleistocene geology is complex with at least three major glaciations and a fourth minor valley glaciation. The principal water-bearing materials include outwash sand and gravel, fluvial sand and gravel between two stony clays, fluvial sand and gravel confined beneath ground moraine or till, and interconnected lenses of coarser material included in the silt, clay, and fine sand that fills some valleys to depths of more than 900 feet. Flowing artesian wells with flows of from 15 to 20 gpm exist in confined aquifers in the valleys.

As less than 7 per cent of the annual total precipitation falls during July and August, nearly all communities and farms south of Fraser River rely on groundwater for domestic use and many crops are irrigated. It is estimated that groundwater consumption in the area exceeds 20 million gpd.

In the Georgia Lowland surficial deposits are thin and impermeable and except for a few local sand and gravel aquifers do not even provide domestic supplies.

The Nanaimo Lowland has thick unconsolidated deposits. Till is the most widespread of these and mantles much of the lowland but in places is separated from the underlying bedrock by fluvial sands and gravels up to several hundred feet thick. These sands and gravels constitute the largest aquifer of the lowlands and large capacity wells tap it to supply water systems at Comox, Bowser, Qualicum, Lantsville, Cowichan Bay, and Sidney. Overlying the till the principal aquifers include ice-contact gravel and sand of varied form, deltaic gravel and sand accumulated where rivers reached former higher sea levels, and modern river bottom, deltaic and alluvial fan deposits. These deposits are rarely more than two miles across but many of them are several tens of feet or even 100 feet thick.

Groundwater Chemistry

In general, groundwaters are low in mineral content and hardness. The harder waters generally are higher in iron and manganese. Groundwaters unsatisfactory for most uses do occur in certain geological environments especially where leaching produces a concentration of salts, where oxidation increases the iron content, and in coastal areas where over pumping may induce sea-water intrusion.

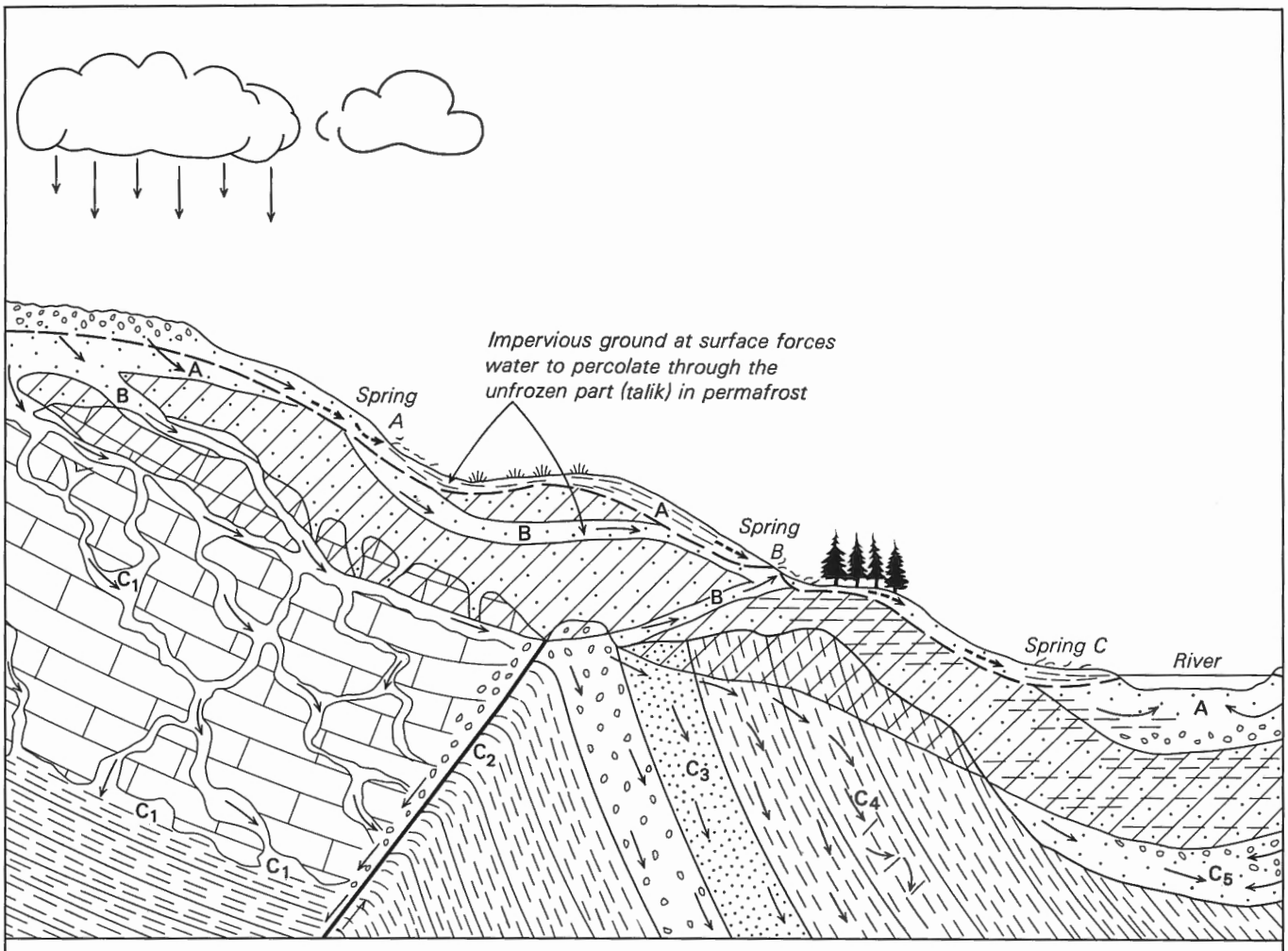
In thermal springs (Fig. XIII-15) calcium sulphate and calcium bicarbonate are the chief constituents and hydrogen sulphide is released on exposure to open air. Sampling of a spring at Banff over an 8-year period indicated little change in chemical composition or temperature. In the Ainsworth district, at the Bluebell mine, in 1962 a flow of water and carbon dioxide at a temperature of 105°F and a rate of 400 gpm under a pressure of 325 pounds per square inch forced abandonment of the lowest level of the mine.

Northern Hydrogeological Region

This region includes in general all of Canada north of the southern limit of discontinuous permafrost. Throughout this vast region hydrogeological data are scarce; precipitation is low, less than 10 inches in the north and less than 20 inches in the south. The region is underlain by a wide variety of crystalline rocks and folded or unfolded sedimentary rocks. Surficial materials cover much of the area and reach their greatest thicknesses in the valley bottoms along Mackenzie River and in the Yukon. In the mountain belts west of Mackenzie River and on Ellesmere Island relief is many thousand feet but elsewhere it is commonly only a few hundred feet and seldom exceeds 1,000 feet.

The approximate southern limits of continuous and discontinuous permafrost are indicated on Figure XIII-7. In the continuous zone permafrost occurs everywhere beneath ground surface and is generally hundreds of feet thick. This zone grades into the discontinuous zone where permafrost exists in combination with areas of unfrozen material, and may be only a few feet thick. Permafrost also exists at high altitudes in southern Labrador–Ungava and in the Cordilleran Region some distance south of the limits shown in the figure.

The area contains three climatic regions. The southern part is sub-arctic, that is having cool short summers of 2 to 3 months with a mean temperature above 32°F. North of this is the Tundra where the mean temperature of the



LEGEND

Clay and silt | Sand | Gravel | Limestone | Permafrost

Water percolating throughout the year . . . ⟶

Water percolating only during the warm season . . . . . ⟶

Depth of winter freezing (bottom of active layer) . . — — —

OCCURRENCE OF GROUNDWATER IN PERMAFROST REGIONS

A. Groundwater above the permafrost (suprapermafrost water)
B. Groundwater within the permafrost (intrapermafrost water)
C. Groundwater below the permafrost (subpermafrost water)
- C_1. Water in solution channels (karst water)
- C_2. Water along a fault fissure (fissure water)
- C_3. Water in a porous layer in bedrock (aquifer)
- C_4. Water in bedrock joints (fissures) - (fissure water)
- C_5. Water in alluvial deposits (alluvial water)

Springs A and C will cease flowing in the winter
Spring B will probably flow the year round

GSC

FIGURE XIII-16. Occurrence of groundwater in permafrost areas.

warmest month is below 50°F but above freezing and the climate is characterized by extremely short cool summers and very long cold winters. The dividing line between these two regions, the 50°F mean temperature isotherm for the month of July, closely approximates the northern limit of tree growth and coincides in general with the southern limit of continuous permafrost. The third climatic region is the region covered by ice caps where the mean temperature for all months is below 32°F. This region is relatively small occupying mountain peaks in the Cordillera and embracing larger ice caps of the northern and eastern parts of Arctic Archipelago.

Hydrogeology

Groundwater in permafrost areas differs in its occurrence from that in warmer climates because of the frozen impermeable ground mixed with unfrozen deposits varying in degree of permeability (Fig. XIII-16). In the discontinuous permafrost zone groundwater may be obtained in unfrozen zones. In the area of continuous permafrost, development of groundwater supplies is frequently impractical and commonly impossible especially where the permafrost extends to bedrock.

The main effect of permafrost on the hydrology of an area is that it restricts the movement of groundwater. In regions of low relief this results in numerous lakes and swamps. Conversely in areas of high relief the run-off is rapid and complete due to lack of infiltration.

Groundwater in permafrost areas may occur as suprapermafrost water, intrapermafrost water, or subpermafrost water. *Suprapermafrost water* is present nearly everywhere in permafrost regions during the summer thawing season. If thawing extends deep enough it can create an appreciable reservoir of groundwater perched upon the underlying frozen material. Included with the suprapermafrost water is groundwater which occurs beneath lakes and rivers. Drilling in the Mackenzie River Delta, for example, indicated unfrozen sediments beneath the centre of a small lake to a depth of 230 feet. Some temporary water supplies are obtained during the summer from shallow wells dug into this active zone.

Intrapermafrost water exists within the thawed zones of permanently frozen ground. It commonly occurs in alluvium near rivers or abandoned river channels and in glaciofluvial material covering the floor of wide river valleys such as Pelly River valley in the Yukon. It may be connected hydraulically with water both in the suprapermafrost and subpermafrost zones.

Wells drilled in such widely separated communities as Mayo, Hay River, Fort Simpson, and Mile 1095 Alaska Highway obtain quantities of water from coarse-grained materials lying between frozen materials.

Mineral content of intrapermafrost water depends upon the conditions of formation of the materials in which it occurs. Waters which rise to the surface and are connected to the zone of active flow usually have a low mineral content whereas those that are connected downward by bedrock structure to a deep seated aquifer may contain highly mineralized water.

Subpermafrost water occurs beneath large areas of permafrost. Examples of this are the flowing wells which occur along the Alaska Highway between Haines Junction and the Alaska Boundary. The highway follows Shakwak Trench, a great valley whose floor is covered with extensive deposits of coarse-grained, glaciofluvial sand and gravel. The water in these flowing wells is reported to be coming from beneath permafrost. Another example was encountered during mining operations in Eldorado Creek valley near Dawson when a 221-foot shaft was sunk to bedrock through frozen muck, sand, and gravel. Water encountered in the bedrock flowed at an approximate rate of 1,000 gpm. The mineral content of subpermafrost water is frequently high.

Mines in the northern region occur mainly in the area of discontinuous permafrost and although none is extremely wet, quantities of water varying from 40,000 gpd to 470,000 gpd are pumped from individual mines. At Yellowknife large flows were encountered at a depth of 2,300 feet. Some of this water is highly mineralized, which indicates that water is present in fractured rocks at depth even in permafrost areas. The possibility of economical development is however extremely remote.

Numerous springs are known in the northern region but little is known regarding the relation to permafrost. The most common host rocks of the larger springs are limestones and dolomites and consequently the waters are high in salts of calcium and magnesium. Many of the larger springs have a high sulphate content suggesting solution of gypsum or possibly oxidation of pyritiferous rocks. Chloride waters are known but the source of the chloride is unknown.

Thermal-springs occur along and west of the valley of Mackenzie River. The temperatures of these springs are not high, 70° to 80°F, but are appreciably above the mean annual temperature of the frozen ground through which they must flow. The maximum temperature reported is 118°F at Mile 497 Alaska Highway.

No Canadian communities situated in the area of continuous permafrost obtain a permanent water supply from groundwater but some do in the area of discontinuous permafrost. Some of these are situated beside a large stream or lake and are obtaining their water from wells in unfrozen material close to the surface water. Dawson and Hay River both have water supplies of this type. Fort Smith and Whitehorse obtain water from well fields in addition to supplies from the river. These communities use groundwater to warm the river water during cold periods in order to prevent freezing in the pipes.

Many small communities along Alaska Highway obtain groundwater from permeable materials near surface water but unfortunately there is little information concerning the occurrence of permafrost in these areas except that it has been encountered close to ground surface in test pits put down close by.

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CH. III

Frontispiece, G. Hunter

CH. IV

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3 C. K. Bell, 117969
4 L. J. Kornik, 113491
5 J. F. Henderson, 84169
6 P. F. Hoffman
7 EMR/A5619-36
8 J. A. Fraser, 124123
9 EMR/A14233-138
10 L. P. Tremblay, 200479
11 P. F. Hoffman
12 P. F. Hoffman
13 J. D. Aitken, 132860
14 EMR/A15649-62
15 F. C. Taylor, 63-2-8
16 EMR/A15830-21
17 M. J. Frarey, 5-5-64
18 M. J. Frarey, 5-9-64
19 R. F. Emslie, 120598
20 H. R. Wynne-Edwards, 3-16-57
21 F. C. Taylor, 5-6-63
22 K. L. Currie, 112643H
23 B. V. Sanford, 200841B

CH. V

Frontispiece, S. M. Roscoe

Plate 1 G. A. Gross, 6-8-58
2 G. A. Gross, 130833
3 E. R. Rose

CH. VI

Frontispiece, D. G. Kelley, 200354G

Plate 1 W. H. Poole, 4-6-67
2 R. D. Hutchinson, 5-6-50
3 L. M. Cumming, 123853
4 W. H. Poole, 5-3-67
5 W. H. Poole, 6-10-67
6 D. M. Baird
7 D. M. Baird
8 H. Williams, 131324
9 B. V. Sanford, 200934
10 F. D. Anderson, 3-4-65
11 B. V. Sanford, 200874
12 B. V. Sanford, 200354E

CH. VII

Frontispiece, G. Hunter

Plate 1 B. V. Sanford, 200835
2 D. M. Baird
3 E. R. Rose
A, 200842G
B, 200842A
4 H. H. Bostock, 114274
5 D. M. Baird
6 Canadian Rock Salt Co. Ltd.
7 A. E. Wooton
8 P. A. Hacquebard

CH. VIII

Frontispiece, J. E. Reesor, 8-6-58

Plate 1 R. A. Price, 2-3-62
2 R. T. Bell, 145960
3 W. H. Fritz, 200875
4 EMR/T8-50L
5 W. H. Poole, 107545
6 R. A. Price (personal)
7 J. E. Reesor, 10-3-63
8 J. E. Reesor, 7-8-59
9 J. A. Roddick, 119768
10 J. O. Wheeler, 123503
11 W. H. Poole, 107538
12 R. A. Price, 6-7-65
13 S. L. Blusson, 123792
14 N. C. Ollerenshaw, 122947
15 E. W. Mountjoy
16 R. A. Price, 6-5-67
17 B.C. Government
18 EMR/T6-117R
19 N. C. Ollerenshaw, 122956
20 RCAF/GSC99472

CH. IX

Frontispiece, Granby Mining Co. Ltd.

Plate 1 G. Hunter
2 International Minerals and Chemical Corp. (Canada) Ltd.
3 British American Oil Co. Ltd.
4 H. R. Belyea
5 Shell Canada Ltd.

CH. X

Frontispiece, E. T. Tozer, 113083B

Plate 1 EMR/T441R-208
2 EMR/T419R-18
3 R. Thorsteinsson, 111950
4 EMR/T419R-36
5 E. T. Tozer, 113086I
6 E. T. Tozer, 113083A
7 EMR/T489L-46
8 EMR/T428C-161

CH. XI

Frontispiece, Hypotype, GSC No. 13995

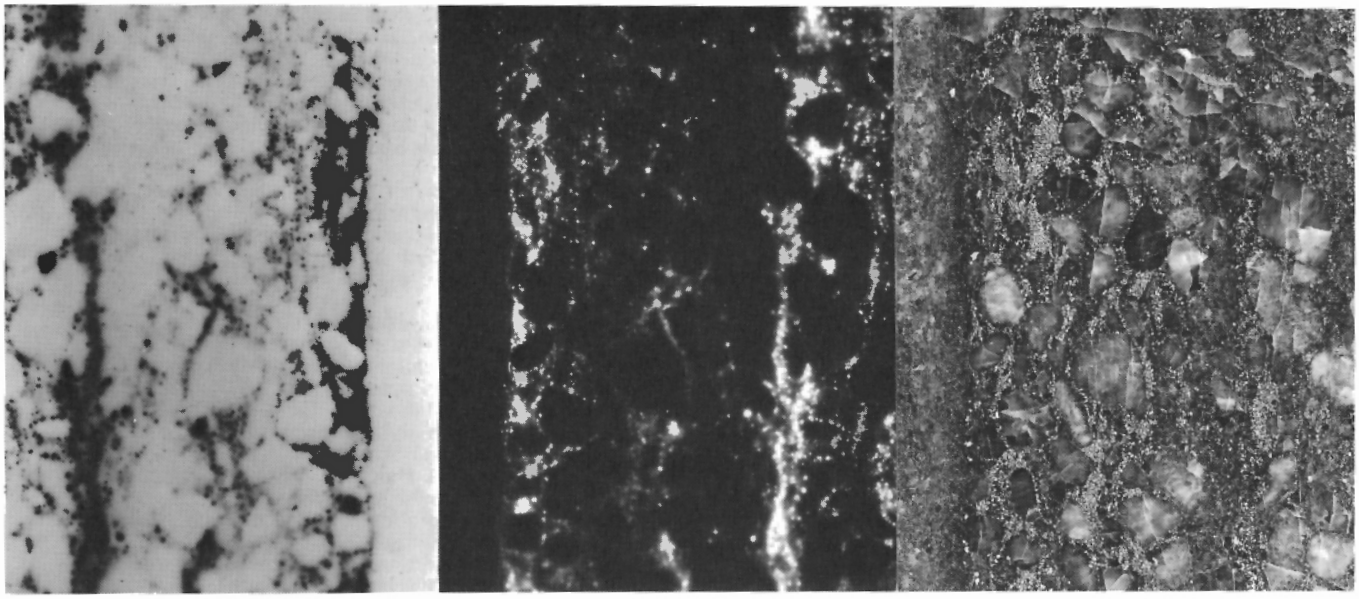
CH. XII

Frontispiece, EMR/T404L-56

Plate 1 I. M. Stevenson, 1-2-56
2 N. R. Gadd, GSC/K5085
3 EMR/A15489-161
4 A. M. Stalker, 127752, 53, 54
5 R. W. Klassen, 124558
6 J. G. Fyles, 200798
7 EMR/A18863-75
8 EMR/A1141-86
9 EMR/A13479-248, 249
10 D. St. Onge, 113453
11 EMR/A19379-154
12 A. M. Stalker, 3-2-55
13 EMR/T327R-193
14 EMR/T152C-107
15 EMR/A12966-315, 316
16 EMR/A16302-29
17 EMR/A15728-63, 64

CH. XIII

Frontispiece, J. S. Scott, 118628, 118661, 3-2-66



Illustrations of these pebbles that appear on pp. 152, 153, cannot be compared because of difference in size of sections. They are repeated here in equal-sized sections for comparative purposes.

GUIDE TO ILLUSTRATIONS

CHAPTER I

Figure 1. Geological regions of Canada 3
Table 1. Area of principal geological and geographic elements of Canada 4

CHAPTER II

Frontispiece. Raised beaches on Hudson Bay Lowland . 10
Plate 1. Fiords and glaciers, Davis Highlands, Baffin Island 14
Plate 2. Alberta Plain and Missouri Coteau 21
Plate 3. Unglaciated part of Yukon Plateau and Yukon River near Dawson, Yukon Territory 22
Plate 4. Old Crow Plain, northern Cordilleran Region, Yukon 24
Figure 1. Physiographic regions of Canada 12

CHAPTER III

Frontispiece. Uranium-fueled electrical generating plant, Douglas Point, Lake Huron, Ontario 32
Figure 1. Value of mineral production 32
Figure 2. Indexes of physical volume of mineral production and of total industrial production 34
Figure 3. Exports from Canada 1950–68 34
Figure 4. Total and per capita consumption of steel and energy in 1966 35
Figure 5. Canadian mineral production and consumption as a per cent of world total, 1965 38
Figure 6. Export market of Canadian minerals in raw and semi-processed form 41
Table 1. Mineral production in Canada, selected years, 1890–1968 32
Table 2. Mineral production related to gross national product selected years, 1930–68 ... 33
Table 3. Net values of production, 1950 and 1967, all industries 33
Table 4. Measured and indicated reserves of some Canadian minerals, 1956 and 1967 39
Table 5. Mineral exports from Canada, by principal commodity 40
Table 6. Exports of principal minerals in relation to production, 1967 40
Table 7. Value of mineral production in Canada by geological regions, 1967 42
Table 8. Value of mineral production in Canada by provinces, 1950–1968 42

CHAPTER IV

Frontispiece. McDonald fault scarp bordering East Arm Fold Belt of Churchill Province, Northwest Territories 44
Plate 1. Partly migmatized Archean greywacke, Superior Province, Povungnituk, Quebec .. 58
Plate 2. Magnetic anomalies, Kapuskasing Subprovince, Superior Province, Ontario 65
Plate 3. Garnetiferous greywacke of Archean Cross Lake Group, Cross Lake, Manitoba 69
Plate 4. Magnetic anomalies, vicinity of boundary between Churchill and Superior Provinces, Manitoba 71
Plate 5. Pillowed lavas of Archean Yellowknife Group, Slave Province, Yellowknife, Northwest Territories 72
Plate 6. Graded beds of greywacke and shale of Archean Yellowknife Group, Yellowknife, Northwest Territories 73
Plate 7. Granite stock intrusive into Archean Yellowknife metagreywacke, Slave Province, Northwest Territories 74
Plate 8. Stromatolites in Aphebian Epworth Group, Bear Province, Northwest Territories 79
Plate 9. Vertical airphoto of diabase dykes of Mackenzie swarm, Bear Province, Northwest Territories 83
Plate 10. Unconformity between late Aphebian Martin Formation and gneisses of the Archean Tazin Group, Churchill Province, Uranium City, Saskatchewan 93
Plate 11. Algal stromatolites in Aphebian Great Slave Group, East Arm Fold Belt, Northwest Territories 95
Plate 12. Thin, graded beds of greywacke and argillite of Aphebian Great Slave Group, Great Slave Lake, Northwest Territories 95
Plate 13. Banded metasedimentary gneiss, possibly Archean, northern Churchill Province, Northwest Territories 99

Plate 14. Aphebian Nastapoka Group unconformably overlying Richmond Gulf Group, Belcher Subprovince, Quebec 105
Plate 15. Orbicular diorite sill, in Aphebian Nastapoka Group, Richmond Gulf, Quebec 106
Plate 16. Gently south-plunging anticline in Aphebian Belcher Group, Belcher Islands, Northwest Territories 107
Plate 17. Sandstone dyke in Espanola siltstone, Southern Province, Espanola, Ontario ... 113
Plate 18. Rafted pebble in Gowganda varved sediments, Aphebian Cobalt Group, Espanola, Ontario 114
Plate 19. Layered leuco-troctolite of the Michikamau anorthosite intrusion, Michikamau Lake, Labrador 119
Plate 20. Grenvillian monzonite intrusive into pyroxene gneiss, Westport, Ontario 127
Plate 21. Contact between diabase dyke and granite, Superior Province, Quebec 133
Plate 22. Clearwater Lake shock metamorphic structure, Clearwater Lake, Quebec 136
Plate 23. Upper Ordovician Bad Cache Rapids Group unconformably on Precambrian granitic gneiss, Churchill River, Manitoba .. 139
Figure 1. Structural provinces and subprovinces of the Canadian Shield 46
Figure 2. Structural trends in the Canadian Shield .. 48
Figure 3. Histogram of dates and ages of orogenies of the Canadian Shield 49
Figure 4. Diagrammatic section of an Archean volcanic-sedimentary complex 55
Figure 5. Thicknesses of Archean sequences in Superior Province 56
Figure 6. Geology of the Timmins–Malartic area, Ontario and Quebec 61
Figure 7. Aeromagnetic expression of the volcanic sequence in the Timmins–Kirkland Lake area, Ontario 63
Figure 8. Geology of a part of Kapuskasing Subprovince, Superior Province, Ontario 64
Figure 9. Location and ages of carbonatites in the Kapuskasing Subprovince, Superior Province, Ontario 65
Figure 10. Geology of part of Cross Lake Subprovince, Superior Province, Ontario 68
Figure 11. Boundary between the Superior and Churchill Provinces, Manitoba 70
Figure 12. Geology of the Yellowknife area, Slave Province, Northwest Territories 75
Figure 13. Typical cross-section of the Muskox Intrusion 82
Figure 14. Possible extent of Aphebian geosynclines and Hudsonian orogens in Churchill and Bear Provinces 86
Figure 15. Helikian and Hadrynian rocks in Churchill and Bear Provinces 88
Figure 16. The McDonald fault system south of Great Slave Lake, Northwest Territories 88
Figure 17. Paleocurrent orientations, Great Slave and Et-Then Groups, East Arm Subprovince, Northwest Territories 96
Figure 18. Paleocurrent orientations, Dubawnt Group, north-central Churchill Province, Northwest Territories 98
Figure 19. Schematic cross-section of the Labrador Fold Belt at successive stages in its development 103
Figure 20. Distribution of the Huronian Supergroup, Southern Province, Ontario 111
Figure 21. Major structures in the Huronian Supergroup, Southern Province, Ontario 116
Figure 22. Labrador Shelf of the Atlantic Continental Shelf 120
Figure 23. Geology of Surprise Lake region, Quebec 122
Figure 24. Mineralogical and textural changes in Archean rocks across Grenville Front, Surprise Lake area, Quebec 123
Figure 25. Tectonic steps in the evolution of Grenville Province, Mont-Laurier–Kempt Lake map-areas, Quebec 124
Figure 26. Structure-section through Seal Lake Fold Belt, Grenville Province, Labrador 126
Figure 27. Geology and metamorphic zones of part of Grenville Province in southeastern Ontario 128
Figure 28. Geology of part of the Westport map-area, Ontario 129
Figure 29. Inferred relationships in the metamorphic interior of an orogenic belt 130
Figure 30. Diabase dyke swarms of the Canadian Shield 132
Figure 31. Ratio of alkali and silica content of diabase dykes from the Canadian Shield 133
Figure 32. Classification of basaltic dyke swarms 133
Figure 33. Sulphur isotope ratios of sulphides from diabase dykes of the Canadian Shield .. 134
Figure 34. Diabase dyke swarms of Canada, Greenland, and Scotland 134
Figure 35. Sites of shock metamorphic structures in the Canadian Shield 135
Figure 36. Configuration of Precambrian rocks beneath Hudson Platform 137
Figure 37. Cross-section of Phanerozoic rocks of southern Hudson Platform 138
Figure 38. Facies and thickness of Upper Ordovician rocks of southern Hudson Platform 140
Figure 39. Facies and thickness of Silurian (early Niagaran) rocks of southern Hudson Platform 141
Figure 40. Facies and thickness of Silurian (late Niagaran(?)) rocks of southern Hudson Platform 141
Figure 41. Facies and thickness of Upper Silurian rocks of southern Hudson Platform 142
Figure 42. Facies and thickness of Lower Devonian rocks of southern Hudson Platform 143
Table 1. Precambrian time-stratigraphic classification 51
Table 2. Average chemical composition of rocks from northeastern Superior Province.... 59
Table 3. Alternative correlations of Archean rocks, Abitibi Subprovince 62
Table 4. Comparison of stratigraphic nomenclature of Huronian rocks, Southern Province ... 110

CHAPTER V

Frontispiece. Uraniferous conglomerate, Blind River, Ontario 152, 793
Plate 1. Algoma-type, Archean iron-formation, Michipicoten district, Ontario 171
Plate 2. Ruth Lake mine, Labrador, Newfoundland 175
Plate 3. Photomicrographs of polished surfaces of titanium ores 184
Figure 1. Archean protocontinents 157
Figure 2. Representative gold occurrences in the Superior Protocontinent 158
Figure 3. Representative nickel and iron occurrences in the Superior Protocontinent 160
Figure 4. Representative copper–zinc and zinc–lead occurrences in the Superior Protocontinent 161
Figure 5. Representative mineral occurrences in relation to Archean sedimentary and volcanic rocks in northern Canadian Shield 162
Figure 6. Distribution of uranium mines in the Canadian Shield 164
Figure 7. Distribution of uranium in the Matinenda Formation, Elliot Lake, Ontario 165
Figure 8. Schematic stratigraphic section through main uranium-bearing part of Quirke Lake syncline 166
Figure 9. Generalized geology and location of uranium mines, Lake Athabasca, Saskatchewan 167
Figure 10. Iron deposits and mines of the southern Canadian Shield 169

Figure 11. Diagram showing major types of chemically precipitated iron-formation with their sedimentary facies and depositional environment ... 170
Figure 12. Iron-formations of the Labrador Geosyncline ... 173
Figure 13. French mine, Schefferville-Knob Lake area, Quebec ... 174
Figure 14. Helen iron mine, Algoma, Ontario ... 176
Figure 15. Steeprock Lake area, Ontario ... 177
Figure 16. Nickel deposits of the Sudbury district, Ontario ... 179
Figure 17. Vertical section through the Strathcona deposit, Sudbury basin, Ontario ... 180
Figure 18. Geology of the Thompson nickel belt, Manitoba ... 181
Figure 19. Distribution of ore at the Thompson mine, Manitoba ... 182
Figure 20. Copper and zinc mines of the Abitibi Belt, Ontario and Quebec ... 186
Figure 21. Main copper and zinc producing part of Noranda area, Quebec ... 191
Figure 22. Features common to many sulphide deposits in Noranda area, Quebec ... 192
Figure 23. Isometric projection of orebodies at Horne mine, Noranda Mines Limited ... 192
Figure 24. Mines and prospects of Matagami area, Quebec ... 193
Figure 25. Generalized geology and copper–gold mines of Chibougamau region, Quebec ... 194
Figure 26. Main Campbell Chibougamau–Merrill Island copper–gold deposit, Quebec ... 195
Figure 27. Flin Flon mine, Manitoba ... 197
Figure 28. Gold deposits of the Porcupine camp, Ontario ... 202
Figure 29. Gold deposits of Kirkland Lake–Larder Lake camp, Ontario ... 204
Figure 30. Gold deposits of northwestern Quebec ... 205
Figure 31. Gold deposits of Little Long Lac camp, Ontario ... 206
Figure 32. Gold deposits of Red Lake camp, Ontario 206
Figure 33. Section showing generalized geological features of Cobalt area, Ontario ... 210
Figure 34. Silverfields mine, Cobalt, Ontario ... 211
Figure 35. Port Radium mine, Port Radium, Northwest Territories ... 214
Figure 36. Regional zoning of lithium and beryllium occurrences, Preissac–Lacorne area, Quebec 216
Figure 37. Diagrammatic composite vertical section of Chemalloy Minerals Limited pegmatite ... 218

Table 1. Production of the ten main mineral commodities from the Canadian Shield in 1966 152
Table 2. Representative mineral occurrences in Superior Protocontinent ... 159
Table 3. Representative mineral occurrences in Slave and Hudson Protocontinents ... 160
Table 4. Percentages of representative mineral occurrences in Archean protocontinents ... 162
Table 5. Classification of iron deposits ... 172
Table 6. Composition of iron-formation types ... 173
Table 7. Estimated distribution of copper and zinc production in 1966 ... 186
Table 8. Copper and zinc mines in Superior Province ... 187–189
Table 9. Copper and zinc mines in Churchill Province ... 190
Table 10. Analyses of rocks from Cobalt and South Lorrain, Ontario ... 213

CHAPTER VI

Frontispiece. Gypsum in the Codroy Group, St. George's Basin, Newfoundland ... 228
Plate 1. Cambrian Cow Head Group at Humber Arm klippe, Cow Head, Newfoundland ... 238
Plate 2. Hadrynian overlain with angular unconformity by Lower Cambrian, Brigus South Point, Newfoundland ... 239
Plate 3. Limestone breccia with interbedded shale and limestone of the Lower Ordovician Green Point Formation, Green Point, Newfoundland ... 248
Plate 4. Bedding planes of Lower Ordovician Green Point Formation, Green Point, Newfoundland ... 249
Plate 5. Contorted Ordovician shale near base of Humber Arm klippe, Port au Port, Newfoundland ... 253
Plate 6. Upper Ordovician English Head Formation, Anticosti Island, Quebec ... 258
Plate 7. Bioherm in Upper Ordovician, Ellis Bay limestone, Anticosti Island, Quebec ... 259
Plate 8. Silurian red sandstone overlain by agglomerate, Botwood Group, Little Fogo Islands, Newfoundland ... 266
Plate 9. Niagara Falls, Ontario ... 271
Plate 10. Recumbent Acadian folds in Ordovician Baie d'Espoir Group, Baie d'Espoir, Newfoundland ... 280
Plate 11. Limestone concretion in Upper Devonian Kettle Point Formation, southern Lake Huron, Ontario ... 285
Plate 12. Erosional unconformity between the Horton and Windsor Groups, Antigonish–Mabou Basin, Nova Scotia ... 286

Figure 1. Principal geological elements of Southeastern Canada ... 231
Figure 2. Late Hadrynian sedimentation, volcanism, and tectonism with features of the middle Hadrynian Avalonian Orogeny and the pre-Hadrynian basement ... 232
Figure 3. Cross-section of Precambrian and Cambrian rocks of Avalon Platform, eastern Newfoundland ... 233
Figure 4. Latest Hadrynian and/or earliest Cambrian sedimentation and tectonism ... 233
Figure 5. Lower Cambrian sedimentation and tectonism ... 236
Figure 6. Middle and Upper Cambrian sedimentation, volcanism, and tectonism ... 237
Figure 7. Cross-section of Paleozoic formations of the St. Lawrence Platform from Michigan Basin to Anticosti Basin ... 241
Figure 8. Cross-section of Cambrian and Ordovician formations in southwestern Ontario and northern New York ... 242
Figure 9. Lower Ordovician sedimentation, volcanism, and tectonism ... 243
Figure 10. Early Middle Ordovician sedimentation, volcanism, and tectonism ... 244
Figure 11. Late Middle Ordovician sedimentation, volcanism, and tectonism, with some features of Taconian Orogeny ... 245
Figure 12. Upper Ordovician sedimentation, volcanism, and tectonism, with some features of Taconian Orogeny ... 246
Figure 13. Lower Silurian sedimentation, volcanism, and tectonism ... 260
Figure 14. Middle Silurian sedimentation, volcanism, and tectonism ... 261
Figure 15. Upper Silurian sedimentation, volcanism, and tectonism ... 262
Figure 16. Cross-section of Lower and Middle Silurian strata, Algonquin Arch to northwestern Allegheny Trough ... 269
Figure 17. Cross-section of Silurian strata, eastern flank of Michigan Basin, southwestern Ontario ... 270
Figure 18. Early Lower Devonian sedimentation, volcanism, and tectonism ... 273
Figure 19. Late Lower and early Middle Devonian sedimentation, volcanism, and tectonism ... 274
Figure 20. Late Middle and Upper Devonian sedimentation, volcanism, and tectonism and features of Acadian Orogeny ... 278
Figure 21. Cross-section through Devonian rocks ... 284

Figure 22. Carboniferous, mainly Tournaisian, sedimentation, volcanism, and tectonism 288
Figure 23. Carboniferous, mainly Viséan, sedimentation, volcanism, and tectonism 290
Figure 24. Extent of marine deposition of subzones of Windsor Group 291
Figure 25. Carboniferous, Viséan to early Westphalian, sedimentation, volcanism, and tectonism 292
Figure 26. Carboniferous, mainly early Westphalian, sedimentation and tectonism 293
Figure 27. Carboniferous, early Westphalian, sedimentation and tectonism 294
Figure 28. Carboniferous and Permian, late Westphalian to early Permian, sedimentation and tectonism 295
Figure 29. Diagrammatic restored section of Carboniferous rocks, New Brunswick and Nova Scotia 296
Figure 30. Mesozoic and Tertiary sedimentation and volcanism, with features of late Triassic Palisades Disturbance 297
Figure 31. Structure-sections, Atlantic Continental Shelf 299
Figure 32. Earthquake epicentres of eastern Canada . 300
Figure 33. Fence diagram through crust, Appalachian Orogen 301
Table 1. Carboniferous strata in the Maritime Provinces 286
Table 2. Carboniferous stratigraphy in the Maritime Provinces 287
Table 3. Subdivisions of the Horton and Anguille Groups 289

CHAPTER VII

Frontispiece. Gaspé Copper Mines, Limited, Murdochville, Quebec 306
Plate 1. Petrolia oil field, Ontario, about 1885 .. 309
Plate 2. First Maritime Mining Corporation Limited copper–zinc–lead mine at Tilt Cove, Newfoundland 327
Plate 3. Photomicrographs of Ordovician iron deposits at Wabana, Newfoundland 333
Plate 4. Magnet Cove Barium Corporation, Walton, Nova Scotia 340
Plate 5. Banded fluorite vein, Director mine, Newfoundland Fluorspar Limited, St. Lawrence, Newfoundland 342
Plate 6. Salt mine of Canadian Rock Salt Company Limited at Ojibway, Ontario 344
Plate 7. Drilling barge 'Timesaver II' on location in eastern Lake Ontario 354
Plate 8. Photomicrographs of Carboniferous coal . 359
Figure 1. Mineral provinces of Southeastern Canada 308
Figure 2. An idealized metallogenic model of the Appalachian Orogen 312
Figure 3. Principal deposits of copper, zinc, and lead in the Appalachian Mineral Province 313
Figure 4. Principal deposits of asbestos, chromium, and nickel in the Appalachian Mineral Province 314
Figure 5. Principal deposits of gold, molybdenum, antimony, bismuth, tin, and tungsten in the Appalachian Mineral Province 315
Figure 6. Generalized cross-section, Needle Mountain and Copper Mountain deposits, Quebec 318
Figure 7. Relationship of copper and lead–zinc deposits to Devonian granite in Shickshock Mountains, Gaspé 319
Figure 8. Principal base-metal deposit, Bathurst–Newcastle district, New Brunswick 322
Figure 9. Bathurst sulphide deposits, New Brunswick 323
Figure 10. Geology and ore zones, Heath Steele mine, New Brunswick 324
Figure 11. Buchans copper–zinc–lead deposits, Newfoundland 326
Figure 12. Geology of the Walton–Cheverie area, Nova Scotia 328
Figure 13. Magnet Cove barite–sulphide mine, Walton, Nova Scotia 329
Figure 14. Geology of Conception Bay, Newfoundland 331
Figure 15. Wabana iron mines, Newfoundland 332
Figure 16. Ore trends, Wabana iron mines, Newfoundland 334
Figure 17. Asbestos deposits at Thetford Mines, Quebec 337
Figure 18. Cross-section through the Bell–Johnson deposit, Thetford Mines, Quebec 338
Figure 19. British Canadian orebody, Black Lake, Quebec 339
Figure 20. Geology of the St. Lawrence fluorspar area 341
Figure 21. Salt mines and brine fields in the Silurian Salina Formation of southwestern Ontario 343
Figure 22. Upper Cambrian oil and gas fields of southwestern Ontario 345
Figure 23. Upper Cambrian, Clearville oil field, Ontario 346
Figure 24. Middle Ordovician oil and gas fields of southwestern Ontario 347
Figure 25. Principal oil and gas fields in the Silurian, Salina, and Guelph Formations of southwestern Ontario 348
Figure 26. Facies of the Silurian Guelph Formation, southwestern Ontario 349
Figure 27. The Tilbury and D'Clute reefs in the Silurian Guelph Formation, Ontario 350
Figure 28. The Silurian Dawn 156 reef, southwestern Ontario 351
Figure 29. Gas fields in the Silurian Clinton and Cataract Groups 352
Figure 30. Devonian oil fields of southwestern Ontario 353
Figure 31. Cross-section through Stony Creek oil and gas field, Moncton Basin, New Brunswick 355
Figure 32. Location, production, and recoverable reserves of the coalfields of Southeastern Canada 357
Figure 33. Lycospora distribution at Joggins and Springhill, Nova Scotia 358
Figure 34. Flood-plain environment of coal deposition 360
Figure 35. Limnic environment of coal deposition ... 361
Table 1. Mineral production of Southeastern Canada in 1966 ..... 310, 311
Table 2. Sulphide deposits of the Bathurst area, New Brunswick 321

CHAPTER VIII

Frontispiece. Valhalla Range, southern Selkirk Mountains, British Columbia 366
Plate 1. Angular unconformity between Upper Cambrian dolomites and Helikian phyllite on Hart River, northern Yukon Territory . 374
Plate 2. Angular unconformity between the Ordovician and Helikian, northern Rocky Mountains, British Columbia 380
Plate 3. Bedding surface with algal mounds in Upper Cambrian Mistaya Formation, Chaba Creek, southern Rocky Mountains, Alberta 384
Plate 4. Whittaker and Iverson Ranges, southern Mackenzie Mountains, Northwest Territories 390
Plate 5. Photomicrograph of subgreywacke from the Devono-Mississippian Sylvester Group, northern Cassiar Mountains 402
Plate 6. Upper Devonian succession along Peechee reef front, Flathead Range, southern Rocky Mountains 405
Plate 7. Veined gneiss of Shuswap Metamorphic Complex, central Monashee Mountains, British Columbia 425
Plate 8. Fold-mullion structures in Shuswap Metamorphic Complex, southern Selkirk Mountains, British Columbia 426
Plate 9. Agmatite, Coast Plutonic Complex, British Columbia 427

Plate 10. Conglomerate of Lower Jurassic Laberge Group with well-rounded granite cobbles, southwest of Whitehorse, Yukon Territory 440
Plate 11. Photomicrograph of Lower Jurassic microcline-bearing greywacke from northern Cassiar Mountains, British Columbia 441
Plate 12. Folds with steeply west-dipping axial plane cleavage, southern Rocky Mountains, British Columbia 451
Plate 13. Late Columbian (Cretaceous) granitic pluton, Selwyn Mountains, Yukon Territory . 453
Plate 14. Ripple-marks on several bedding surfaces of Blackstone Formation, Foothills of Alberta 462
Plate 15. Westerly overturned folds in Miette Range, southern Rocky Mountains, Alberta 470
Plate 16. Normal fault in Paleozoic carbonates, Cirrus Mountain, southern Rocky Mountains, Alberta 471
Plate 17. Rocky Mountain Trench, near the junction of Fraser and Torpy Rivers 471
Plate 18. Late Tertiary thrusts in Kluane Ranges, St. Elias Mountains, Yukon Territory 475
Plate 19. Sandstone channel-filling in mudstones of Tertiary Paskapoo Formation, Burnt Timber Creek, Foothills of Alberta 479
Plate 20. Cinder cone on post-glacial basalt of north flank of Mount Edziza volcanic complex, British Columbia 480

Figure 1. Principal geological elements of Western Canada 368
Figure 2. Pre-Paleozoic paleogeology of Western Canada, and Helikian and Hadrynian sedimentation and tectonism 370
Figure 3. Correlation of the Helikian (Purcell) rocks of southern Rocky Mountains and eastern Purcell Mountains, Alberta and British Columbia 372
Figure 4. Lower Cambrian sedimentation and tectonism 376
Figure 5. Middle Cambrian sedimentation and tectonism 377
Figure 6. Upper Cambrian sedimentation and tectonism 379
Figure 7. Schematic restored stratigraphic section of Cambrian and Ordovician rocks, southern Rocky Mountains and Interior Plains of Alberta and British Columbia 382
Figure 8. Composite stratigraphic column of Middle Cambrian to Middle Ordovician rocks of southern Rocky Mountains 383
Figure 9. Lower and Middle Ordovician sedimentation and tectonism 386
Figure 10. Upper Ordovician and Silurian sedimentation and tectonism 387
Figure 11. Diagrammatic restored sections of lower Paleozoic rocks between Cassiar Platform, Cordilleran Geosyncline, and northwestern Canadian Shield 389
Figure 12. Stratigraphic section of Cambrian, Ordovician, and Silurian rocks of Lloydminster Embayment and Williston Basin 392
Figure 13. Lower Devonian sedimentation and tectonism, and some features of the pre-Devonian geology and physiography 393
Figure 14. Middle Devonian, Eifelian, sedimentation and tectonism 395
Figure 15. Middle Devonian, Givetian, sedimentation and tectonism 396
Figure 16. Upper Devonian, Frasnian, sedimentation and tectonism, and some features of the Ellesmerian Orogeny 397
Figure 17. Upper Devonian, Famennian, sedimentation and tectonism 398
Figure 18. Stratigraphic section of Devonian rocks between White Mountains and southern Rocky Mountains 400
Figure 19. Diagrammatic restored section of Devonian and Mississippian rocks between Cassiar Mountains and Hay River 401
Figure 20. Stratigraphic section of Devonian rocks between the southern Rocky Mountains and Manitoba 403
Figure 21. Main depositional stages in the formation of the Upper Devonian, Frasnian, Miette reef complex 404
Figure 22. Early Mississippian (Kinderhookian) sedimentation and tectonism 412
Figure 23. Carboniferous (post-Kinderhookian) sedimentation and tectonism 413
Figure 24. Permian sedimentation and tectonism 414
Figure 25. Cross-section of Carboniferous and Permian rocks of Peace River Embayment .. 419
Figure 26. Stratigraphic section through Mississippian rocks of Williston Basin in Saskatchewan 421
Figure 27. Distribution of ultramafic intrusions in the Cordilleran Orogen 422
Figure 28. Distribution of granitic plutons and metamorphic complexes of Cordilleran Orogen 422
Figure 29. Upper Triassic, Karnian, sedimentation and tectonism, and some features of the Tahltanian Orogeny 429
Figure 30. Upper Triassic, Norian, sedimentation and tectonism 430
Figure 31. Diagrammatic restored sections across Whitehorse Trough, Yukon Territory, during the Mesozoic era 432
Figure 32. Diagrammatic restored sections across Whitehorse, Nechako, Quesnel, and Rossland Troughs of the Cordilleran Geosyncline from Upper Triassic to Middle Jurassic times 433
Figure 33. Lower and early Middle Jurassic sedimentation and tectonism, and some features of the Inklinian Orogeny 437
Figure 34. Late Middle and Upper Jurassic sedimentation and tectonism and some features of the Nassian Orogeny 439
Figure 35. Early Lower Cretaceous sedimentation and tectonism, and some features of the early phase of the Columbian Orogeny and of pre-Albian geology 446
Figure 36. Lower Cretaceous, Albian, sedimentation and tectonism, and some features of the middle phase of the Columbian Orogeny 447
Figure 37. Early Upper Cretaceous sedimentation and tectonism, and some features of the late phase of the Columbian Orogeny 449
Figure 38. Structure-sections through southern Cordilleran Orogen ..... Facing p. 450
Figure 39. Structure-sections through northern Cordilleran Orogen ..... Facing p. 450
Figure 40. Structure-section through southern Ogilvie Mountains 452
Figure 41. Restored longitudinal stratigraphic section of Mesozoic and Tertiary rocks along the Foothills of Alberta and British Columbia 456
Figure 42. Stratigraphic section of Mesozoic and Tertiary rocks from British Columbia to Manitoba 457
Figure 43. Stratigraphic section of Mesozoic and Tertiary rocks from northeastern British Columbia to Saskatchewan 458
Figure 44. Late Upper Cretaceous and early Tertiary sedimentation and tectonism, and some features of the Laramide Orogeny 465
Figure 45. Tertiary, post-Eocene, sedimentation and tectonism 466
Figure 46. Livingstone thrust, southern Foothills of Alberta 467
Figure 47. Limits of formations above McConnell thrust plane, southern Rocky Mountains .. 468
Figure 48. Flexural-slip folds in southern Foothills of Alberta 469

Figure 49. Cross-section through Lewis thrust sheet, southern Rocky Mountains 472
Figure 50. Geology in the vicinity of Bonnet Plume Basin, northern Yukon Territory 474
Figure 51. Hypothetical cross-section through the crust beneath Queen Charlotte Islands ... 476
Figure 52. Magnetic anomalies southwest of Vancouver Island 477
Figure 53. Mount Edziza volcanic complex, northwestern British Columbia 480
Table 1. Summary of events in evolution of Valhalla Complex 426

CHAPTER IX

Frontispiece. Ironsides pit of Phoenix copper mine, British Columbia 490
Plate 1. Open pit and mill of Endako molybdenum mine, British Columbia 514
Plate 2. Mining potash underground at Esterhazy, Saskatchewan 520
Plate 3. Sulphur extraction and natural gas processing plant, Pincher Creek, Alberta 521
Plate 4. Photomicrographs of typical Devonian carbonate rocks 529
Plate 5. Drilling rig, Shell Waterton #1, Alberta .. 541
Figure 1. Mineral provinces of Western Canada 494
Figure 2. Lead-zinc deposits of Western Canada ... 496
Figure 3. Geology in the vicinity of Pine Point lead–zinc deposits, District of Mackenzie 497
Figure 4. Cross-section through Sullivan orebody, British Columbia 505
Figure 5. Diagrammatic cross-section through Jersey lead–zinc deposit, Salmo, British Columbia 506
Figure 6. Geology of Keno and Galena Hills, Yukon 507
Figure 7. Lode gold deposits of the Cordilleran Region 508
Figure 8. Bralorne–Pioneer gold mine, British Columbia 509
Figure 9. Copper deposits of the Cordilleran Region 511
Figure 10. Geology in the vicinity of Phoenix copper deposits, British Columbia 512
Figure 11. Geology and cross-section of Galore Creek copper deposits, British Columbia 513
Figure 12. Cross-section through Canada Tungsten mine, Flat River, District of Mackenzie .. 516
Figure 13. Occurrences of potash in the Prairie evaporite basin, Saskatchewan 519
Figure 14. Coal-bearing areas of Western Canada .. 523
Figure 15. Devonian oil and gas fields of Western Canada 526
Figure 16. Cross-section through Middle Devonian Clarke Lake field, British Columbia 527
Figure 17. Cross-section through Upper Devonian, Swan Hills reef of south Judy Creek field, Alberta 530
Figure 18. Distribution of major facies in upper Leduc Formation, Redwater reef complex, Alberta 531
Figure 19. Composite cross-section through the upper Leduc Formation, Redwater reef complex, Alberta 531
Figure 20. Bonnie Glen–Wizard Lake reef trend, Alberta 532
Figure 21. Mississippian and Permian oil and gas fields of Western Canada 533
Figure 22. Mississippian oil and gas fields, Sylvan Lake–Minnehik trend, Alberta 534
Figure 23. Cross-section through the Mississippian from Brazeau River field to Minnehik field, Alberta 535
Figure 24. Mesozoic oil and gas fields of Western Canada 536
Figure 25. Structural contours on top of the Triassic system near Fort St. John, British Columbia 537
Figure 26. Isopachs and facies of the Triassic Pardonet and Baldonnel Formations near Fort St. John, British Columbia 538
Figure 27. Isopachs and facies of Triassic Halfway Formation near Fort St. John, British Columbia 538
Figure 28. Oil sands of northern Alberta 539
Figure 29. Schematic cross-section showing oil sands deposits and associated formations, Alberta 539
Figure 30. Structure-section through the central part of the Turner Valley oil and gas field, Alberta 540
Figure 31. Structure-section through Savanna Creek gas field, Alberta 542
Figure 32. Structure-section through Waterton and Pincher Creek gas fields, Alberta 543
Table 1. Mineral production of Western Canada 492–493
Table 2. Main lode metal mines and industrial mineral properties of Western Canada ... 499–503
Table 3. Proved reserves of crude oil in Western Canada 525
Table 4. Proved reserves of natural gas in Western Canada 525

CHAPTER X

Frontispiece. Triassic Blaa Mountain shales with gabbro sills on Raanes Peninsula, east side of Eureka Sound, Ellesmere Island 548
Plate 1. Lower Paleozoic strata overlying Precambrian crystalline rocks, south coast of Devon Island 554
Plate 2. Upper Devonian rocks, and late Cenozoic fault, Blue Hills, southwestern Melville Island 565
Plate 3. Upper Devonian Frasnian Gyrfalcon Bluff, northern Banks Island 566
Plate 4. Parry Islands Fold Belt, northwest Melville Island 567
Plate 5. Lower Devonian Eids Formation unconformably overlain by Permian Trold Fiord Formation, north of Trold Fiord, Ellesmere Island 573
Plate 6. Mesozoic rocks north of Blaa Mountain, Ellesmere Island 574
Plate 7. Eurekan folds with diapiric anticlines, western Axel Heiberg Island, south of Strand Fiord 585
Plate 8. Isachsen dome, Ellef Ringnes Island 587
Figure 1. Geological provinces of the Arctic Archipelago 550
Figure 2. Columnar sections, Proterozoic and Cambrian rocks, central-eastern Ellesmere Island 553
Figure 3. Distribution of Upper Ordovician to Lower Devonian rocks in Arctic Archipelago 556
Figure 4. Diagrammatic restored sections of Upper Ordovician to Lower Devonian rocks of Arctic Archipelago 558
Figure 5. Distribution of Middle Devonian rocks in Arctic Archipelago 562
Figure 6. Diagrammatic restored section of Middle and Upper Devonian rocks of Arctic Archipelago 563
Figure 7. Outcrop of Middle Devonian, Vendom Fiord, Eids, and Goose Fiord Formations, Ellesmere Island 564
Figure 8. Distribution and restored section of Pennsylvanian and Permian rocks in Arctic Archipelago 570
Figure 9. Distribution of Triassic rocks in Arctic Archipelago 576
Figure 10. Restored sections, Triassic formations of Arctic Archipelago 577
Figure 11. Distribution of Jurassic and Cretaceous rocks in Arctic Archipelago 580
Figure 12. Restored sections, Jurassic and lowermost Cretaceous rocks of Arctic Archipelago .. 581
Figure 13. Restored sections, Cretaceous rocks of Arctic Archipelago 582

CHAPTER XI

Frontispiece. *Eurypterus lacustris* Harlen — an arthropod from the Upper Silurian Bertie Formation 592

Plates 1 to 33. Illustrations of fossils

Figure 1. Distribution of Ordovician and Silurian faunas 603

Figure 2. Inferred maximum extent of Lower Cretaceous (Berriasian to Barremian) seas in western and arctic Canada 652

Figure 3. Inferred maximum extent of Early Middle Albian and mid-Upper Cretaceous seas in western and arctic Canada 653

Table 1. Ranges of Cambrian trilobite genera in the North American and Atlantic faunal provinces 594

Table 2. Ordovician faunal successions 602

Table 3. Silurian faunal successions 604

Table 4. Distribution and succession of Devonian faunas 615

Table 5. Distribution and range of Carboniferous and Permian invertebrate fossils 624

Table 6. Range and distribution of Triassic pelagic faunas 634

Table 7. Distribution and succession of Jurassic guide-fossils 642

Table 8. Succession and distribution of Upper and Lower Cretaceous faunas 650

Table 9. Known occurrences of taxonomically identifiable plant fossils in Canada 668

CHAPTER XII

Frontispiece. Viking ice cap and glaciers, Ellesmere Island, Northwest Territories 676

Plate 1. Karst topography on gypsum at site of interglacial deposits, Milford, Nova Scotia 681

Plate 2. Section of postglacial sediments, Gentilly Till, and St. Pierre sediments on Bécancour River, north of Aston Junction, Quebec .. 682

Plate 3. Deep meltwater channel between unglaciated Cypress Hills and Wisconsin end moraine complex, southwestern Saskatchewan 690

Plate 4. Section of non-glacial and glacial sediments on Oldman River at Brocket, Alberta 692

Plate 5. Section of glacial and interglacial deposits on Minnedosa River, north of Minnedosa, Manitoba 694

Plate 6. High-level terraces with Beaufort-type sediments overlying Cretaceous bedrock, eastern Axel Heiberg Island, Northwest Territories 699

Plate 7. Ribbed moraine in eastern Newfoundland . 707

Plate 8. Fluted ribbed moraine south of Ungava Bay, Quebec 736

Plate 9. Ice-marginal and subglacial meltwater channels, Mushalagan River, Quebec 737

Plate 10. Hummocky disintegration moraine near Cypress Hills, Saskatchewan 738

Plate 11. Hummocky disintegration moraine east of Lake Johnstone, southern Saskatchewan .. 739

Plate 12. Blocks of pebbly quartzite of erratics train from Athabasca Valley glacier, Claresholm, Alberta 740

Plate 13. Drumlin belt, Stephansson Island, Northwest Territories 743

Plate 14. Esker complex west of Ennadai Lake, Northwest Territories 744

Plate 15. Fluted and kettled ice-contact deposits with raised beaches, Ennadai Lake, Northwest Territories 746

Plate 16. Barnes Ice Cap with recent marginal and end moraines, Baffin Island 749

Plate 17. Klutlan Glacier, St. Elias Mountains, Yukon Territory 753

Figure 1. Time-stratigraphic relationships of Late Pleistocene in east-central Canada 678

Figure 2. Location of buried organic deposits in Nova Scotia 679

Figure 3. Location of buried organic deposits in St. Lawrence Lowlands 682

Figure 4. Schematic diagram showing stratigraphic relationships of Pleistocene deposits at Scarborough Bluffs, Toronto, Ontario 683

Figure 5. Schematic diagram showing stratigraphic relationships of Pleistocene deposits in Don Valley brickyard, Toronto, Ontario 684

Figure 6. Wisconsin and Illinoian drift in railway-cut, Woodbridge, Ontario 685

Figure 7. Generalized profile through Pleistocene deposits near Port Talbot, Ontario 686

Figure 8. Location of buried organic deposits in Hudson Bay Lowland 687

Figure 9. Inferred preglacial drainage systems of the Interior Plains 691

Figure 10. Location of buried organic deposits in the Interior Plains 693

Figure 11. Location of buried organic deposits of the mainland Arctic Coastal Plain 696

Figure 12. Occurrences of 'old' organic materials on Baffin Island 697

Figure 13. Location of buried organic deposits and 'old' shell occurrences on the arctic islands 700

Figure 14. Location of buried organic deposits in the Cordilleran Region: (a) Southern British Columbia; (b) Glaciated parts of Yukon Territory 703

Figure 15. Stages in the deglaciation of Wisconsin ice .....Facing p. 706

Figure 16. Glacial lake phases during the recession of Wisconsin ice from central Canada; parts a to x 714–725

Table 1. Composite section of Pleistocene deposits, St. Lawrence Lowlands 683

Table 2. Composite section of Pleistocene deposits, Fraser Lowland 704

Table 3. Production and value of sand and gravel . 757

CHAPTER XIII

Frontispiece. Artesian groundwater flow Coldstream ranch, Vernon, B.C. 766

Figure 1. Hydrogeological regions of Canada 769

Figure 2. Hydrologic cycle 770

Figure 3. Water-table conditions 770

Figure 4. Artesian conditions 771

Figure 5. Diurnal and semidiurnal fluctuations 772

Figure 6. Seasonal and instantaneous fluctuations .. 773

Figure 7. Climatic regions of Canada, permafrost regions and annual run-off in southern Canada 774

Figure 8. Bedrock hydrostratigraphic units in New Brunswick, Nova Scotia, and Prince Edward Island 775

Figure 9. Climatic data for stations in St. Lawrence Lowlands 777

Figure 10. Average yield of bedrock wells in Ontario part of St. Lawrence Lowlands Hydrogeological Region 778

Figure 11. Average yield of wells in surficial deposits in Ontario part of St. Lawrence Lowlands Hydrogeological Region 779

Figure 12. Bedrock hydrogeological divisions and buried valleys of Interior Plains Region .. 782

Figure 13. Hydrochemical zonation of shallow bedrock aquifers, Interior Plains Hydrogeological Region 784

Figure 14. Hydrochemical zonation of low-flow river 785

Figure 15. Major groundwater features, Cordilleran Hydrogeological Region 787

Figure 16. Occurrence of groundwater in permafrost areas 789

Table 1. Canadian water supplies 768

PAGE

Abitibi Belt 50, 55, 121, 185, 191
River 688, 689, 734
Formation 617
Subprovince 60
Upland 15–17
Abitibi–Wawa belt 157–159
Abloviak River 735
Acadian Bay ice lobe 709, 710
Orogeny 229, 251, 253, 257, 262, 278–283, 313
Ace–Verna (2) mine, Sask. 163, 167
Ackley-type granite 276, 282, 335
Acknowledgments 6
Actinolite, Ont. 221
Active Formation 383, 391, 403
Acton mine, Que. 320
Acton Vale 320
Adamant batholith 415
Adams Creek 690
(60a) mine, Ont. 175
Adanac Member 444
Adeline Island Formation 126
Adeyton Group 240
Adirondack anorthosite 127
Highlands 124, 127
Mountains 17, 242, 277, 713
Admiralty Formation 604
Inlet 747, 749
Lake 727
Advocate Harbour 710
(4) mine, Que. 338
Aftonian Interglacial Interval 677
Agassiz, B.C. 786
Agnew Lake 163
Agnico mine, Ont. 210
Agnostus Cove Formation 240
Agutua Moraine 731–733
Ahr Lake 102
Aida Formation 372
Aiken Lake 415, 504
Aillik, Labrador 127, 130
Group 127
Ainsworth, B.C. 788
Ajax quartzite 391
Aklavik, N.W.T. 473
Arch 438, 444, 455
Range 444
Alaska 18, 25, 26, 378, 381, 388, 400, 418, 423, 424, 427, 435, 444, 455, 474, 649
Highway 518
Alaskan Island 27
Albanel Formation 59
Lake 174
Albany River 689, 732
Albemarle Group 269
Albert county, N.B. 342
Formation 287, 288, 355
mine (53a), Que. 320
mines, N.B. 336, 355
Alberta Arch 378, 385, 388, 394, 403, 408
Group 461

PAGE

Alberta Plain 20, 21, 459
Plateau 20
Ridge 394
Trough 375, 378, 380, 384, 385, 388, 391, 394, 399, 411, 415, 419, 420, 431, 438, 440, 444, 445, 448, 784
Albertan till 692
Aldridge Formation 371, 495, 504
Alexo (54) mine, Ont. 182
Algonquin Arch 228, 230, 235, 242, 243, 245, 255, 256, 263, 268–272, 283, 345, 346
Alice Arm 514
Alida-Tilston beds 534
Alkaline complexes 65, 76, 122, 128, 129, 186, 298, 311, 335
Allan mine, Sask 520
Allard Lake 126, 156, 183
anorthosite 126
Allegheny Basin 143, 144, 345, 346, 351
Trough 228, 230, 244, 245, 255–257, 259, 260, 262, 263, 268, 269, 272, 277, 283
Allen Bay Formation 557, 558, 561, 605
Alsek River 479
Altyn Formation 371
Amabel Formation 142, 270, 272, 344, 605
Amable du Fond River 729
Amadjuak Lake 748, 749
Amaranth (17) mine, Man. 519
Amherst, N.S. 772
Harbour, Que. 711
Island 711
Amherstburg (119a) brine field, Ont. 342
Formation 144, 277, 344
Amisk belt 157–159
Group 90–92, 196
Amos, Quebec 192
Amulet (18a) mine, Que. 192
Amund Ringnes Island 751
Amundsen Gulf 19
lobe 742–744
Anderdon Formation 277
Anderson Cove Formation 267
Formation 277
Lake 198
Plain 19, 455
River 742
Anglo-Rouyn (6) mines, Sask. 198
Anguille Group 287
Animikie Group 108, 118, 213
Annapolis Formation 298
Lowland 28
Anticosti Basin 228–230, 235, 240–243, 245, 247, 250, 257, 259, 260, 263, 271, 278, 345, 354
Island 28, 228, 229, 242, 250, 257, 260, 261, 271, 281, 296, 298, 354, 602, 605, 776, 778
Antigonish Highlands 28, 247, 253, 268, 709, 710
Upland 288, 289, 294
Antigonish–Mabou Basin 287–292, 295, 296
Antimony deposits 315, 316
Antler Formation 415

PAGE

Antoinette Formation ... 571
Apatite deposits ... 220
Aphebian ... 51, 52, 59, 74–124, 369
  mineral deposits ... 155, 156, 159–162
Appalachian Glacier Complex ... 706, 708–712
  Hydrogeological Region ... 772–776
  Mineral Province ... 311–314
  Mountains ... 28, 29
  Orogen (Geosyncline) ... 228–240, 243–255, 257–268, 272–296, 300
  Region ... 17, 27–29
Appekunny Formation ... 371
Archean ... 51, 54–76, 85, 88–93, 96, 100, 101, 104, 119–124, 156–159, 369
Archer Fiord ... 560
Archibald Formation ... 442
Arcs Member ... 403, 410
Arctic Archipelago ... 548–590
  Bay ... 555
    Formation ... 101
  Coastal Plain (Continental Shelf) ... 17–19, 366, 461, 478, 548, 551, 552, 584, 587, 588
  Lowlands ... 17–19, 584
  Ocean ... 17–19
  Red River ... 20, 702
Arctomys Formation ... 385, 451
Ardtrea phase lake ... 728
Argentia, Nfld. ... 330
Argonaut (46) mine, B.C. ... 515
Arisaig, N.S. ... 331, 605, 617
  Group ... 247, 268, 276
Arizona ... 623
Arkona shale ... 283
Armagh Group ... 236, 237
Arnica Formation ... 401, 406
Arnold River Formation ... 246
Arsenic deposits ... 217
Artillery Lake ... 746
Arvida, Que. ... 219
Asbestos deposits ... 36, 37, 156, 221, 336–338, 518
  Hill, Que. ... 221
Ascot Formation ... 246, 252
Ashcroft, B.C. ... 442
Ashern Formation ... 143, 406
Ashuanipi Lake ... 737
Ashville Formation ... 460
  Group ... 460, 464
Asitka Group ... 416
Askin Group ... 399, 402
Aspy Bay ... 709
Assean Lake Group ... 88
Asselin Formation ... 263
Assiniboia, Sask. ... 21
Assiniboine River Valley ... 732
Assinica Lake ... 218
Assistance Formation ... 573, 574
Aston Formation ... 100
  Junction, Que. ... 682
Astroblem ... 117, 135–137, 178
Atan Group ... 378, 424
Athabasca district, Sask. ... 163, 167
  Formation ... 59, 94, 137, 167

PAGE

Athabasca Lake ... 47
  Oil Sands ... 37, 41, 459, 460, 539
  Plain ... 13
  Plate ... 47, 53
  River ... 20, 403, 409, 419, 436, 444, 461, 463, 479, 498, 539, 740–742, 754
Athole Point Formation ... 101
Atikokan, Ont. ... 66
Atlantic Coastal Plain ... 230
  Continental Shelf ... 28, 29, 120, 145, 230, 297, 298, 309, 311, 313, 315, 345
  Uplands of Newfoundland ... 28
    Nova Scotia ... 28
Atlin, B.C. ... 517
  Horst ... 423, 431, 434, 438, 440, 442, 443, 452
  Intrusions ... 423
  Lake ... 417
Atomic Energy Control Board ... 163
Attawapiskat Formation ... 142, 605
  River ... 689
Attikamagen Formation ... 102
  Lake ... 198
Atwater–Lapointe deposit, Que. ... 185
Aurum (31) mine, B.C. ... 509
Australia Creek ... 703
Avalon ice cap ... 706
  Isthmus ... 708
  Peninsula ... 29, 234, 240, 279, 280, 282, 334, 706, 708
  Platform ... 230, 235, 239, 240, 243–245, 248, 249, 252, 253, 255, 261, 262, 265, 267, 268, 272, 276, 277
Avalonian Orogeny ... 230, 231, 233
Awantijish Formation ... 263, 605
Awingak Formation ... 575, 581–583
Axel Heiberg Island ... 551, 552, 560, 561, 568, 569, 572, 574, 575, 579, 581–586, 588, 589, 698, 699, 701, 751, 752
  Mountains ... 17

Bache Peninsula ... 14, 552, 555, 557
Back Lowland ... 13
  River ... 13, 671, 745, 746
Backbone Ranges ... 24, 388
Bad Cache Rapids Group ... 138, 139, 604
  Heart Formation ... 463
  Vermillion Lake ... 67, 185
Badger deposit, Ont. ... 214
  Newfoundland ... 319
Baffin Bay ... 29, 575
  Coastal Lowland ... 14
  Island ... 14, 19, 29, 87, 88, 100, 153, 156, 159, 177, 548, 554, 557, 582, 589, 697, 698, 747–750
  Upland ... 14
Baie d'Espoir ... 280, 329
  Group ... 239, 255
Baié Johan Beetz ... 221
Baie St. Paul ... 328
  Verte ... 37, 251, 309, 318, 327, 336, 338, 708
  Group ... 247, 338
Baillarge Formation ... 557, 558

PAGE

Baillie Island ... 696
Baker Lake ... 97, 99, 159
Baker Talc (50, 51) mines, Que. ... 340
Bakken Formation ... 410, 419, 534
Baldonnel Formation ... 436, 438, 537
Bamberton, B.C. ... 520
Bancroft, Ont. ... 36, 124, 128, 156, 163, 164, 199, 221
(107, 108) district, Ont. ... 168
Banff, Alta. ... 444, 455, 623, 788
Formation ... 405, 419, 522, 535
Banks Island ... 18, 84, 551, 552, 561, 566, 574, 575, 582, 583, 587, 616, 698–700, 742
Bannockburn township, Ont. ... 221
Banquereau Bank ... 29, 299
Bar River Formation ... 114
Barachois Group ... 292, 293
Baril Member ... 420
Barite deposits ... 221, 338
Barkley Sound ... 786
Barn Mountains ... 373, 399, 418
Barnat (32a) mine, Que. ... 204, 205
Barnes Ice Cap ... 697, 698, 748, 749
Barney's River ... 710
Barron River ... 728, 729
Barrow Dome ... 625
Bashaw (86) oil field, Alta. ... 532
Basler Lake ... 77, 78, 80
Bass Islands Formation ... 143, 271, 277
Bastien Ridge Formation ... 583
Bathurst, N.B. ... 37, 309, 316, 321–324, 328, 335, 711, 712
fault ... 88
Hills ... 13
Inlet ... 13, 76, 84, 100, 159, 199, 200, 744, 747
(6) iron mine, N.B. ... 333
Island ... 18, 550, 551, 559–567, 572, 584, 587, 589, 616, 701, 750
Formation ... 550, 559, 560, 616
Plate ... 45, 52
Subprovince ... 76
Bathurst–Newcastle area (1 to 5), N.B. ... 36, 320–322
Battery Point Formation ... 274, 275, 276
Battle Formation ... 464, 521
River ... 691, 740
Baumann Fiord ... 558, 568, 585, 586
Formation ... 557
Bay d'Espoir ... 280, 329
Lowland ... 28
du Nord Group ... 276
Fiord ... 564, 571, 572
Formation ... 557
of Chaleur ... 265, 276, 283, 318, 711, 712
Exploits ... 255, 316
Fundy ... 28, 284, 297, 298, 356, 709, 712, 776
Islands ... 316, 708
complex ... 316, 317, 336, 338
surface ... 708
St. George's Delta ... 707
St. Lawrence ... 679
Bayham gas field, Ont. ... 351
Bayonne batholith ... 450

PAGE

Bear Cove ... 329
Point Formation ... 552
Province ... 45, 77–84, 155
Rock Formation ... 401, 614
Bearpaw Formation ... 464, 521
Bear–Slave Upland ... 13
Beattie Peaks Formation ... 456
Beattie–Duquesne mine, Que. ... 217
Beatton River (13a) oil field, B.C. ... 537
Beauceville Formation ... 254
Beaufort Formation ... 18, 584, 587, 588, 698–701
Sea ... 19, 742, 755
Beauharnois, Que. ... 36
Axis ... 776, 780
Formation ... 249
Beaver Bight Formation ... 255
mine, Ont. ... 214
Mines Formation ... 456–459
Beaverfoot Formation ... 391, 405, 605
Beaverhill Lake Formation ... 408, 528
Group ... 404
Platform ... 409
Beaverlodge, Sask. ... 36, 156, 163, 167
Bécancour, Que. ... 779
River ... 338
Till ... 682
Beck mine, Ont. ... 214
Becscie Formation ... 271, 605
Bedford shale ... 284
Beechey Island ... 751
Beechhill Cove Formation ... 268
Formation ... 605
Beekmantown Group ... 344, 354
Beg (6) gas field, B.C. ... 537
Béland–Upper York River highlands ... 712
Belcher Channel ... 550
Formation ... 571–573
Fold Belt ... 47, 52, 86–88
Group ... 106, 107, 119
Islands ... 15, 105–107, 154, 155, 161, 174
Bell Arch ... 138
Island ... 309
Group ... 249
River anorthosite ... 193
Shale ... 283
Bell–Johnson deposit, Que. ... 338
Bella Bella, B.C. ... 460
Bella Coola, B.C. ... 427, 441
Belle Bay Formation ... 267, 335
Plaine (13) mine, Sask. ... 520
Belledune smelter, N.B. ... 309
Belleoram granite ... 283
Belleterre, Que. ... 180
Belleville beach ... 727
Belleville–Fort Anne phase lake ... 727
Bellevue, Alta. ... 524
Bellingham Basin ... 476
Belloy Formation ... 419, 537
Bellshill Lake (76a) oil field ... 539
Belly River Formation ... 461, 463, 464, 540, 783
Group ... 463
Benacadie, N.S. ... 680

PAGE

Bending Lake iron range ... 175
Bennett Formation ... 245
Group ... 237
Bentonite deposits ... 521
Berea sandstone ... 284
Bering Sea ... 755
Berland River oil field, Alta. ... 533
Bernard Island ... 700
Bernic Lake ... 220
Bernier–Berlinguet Peninsula ... 749
Bertie Akron Formation ... 605
Formation ... 277
Beryllium deposits ... 217
Besa River Formation ... 403, 410, 418, 471, 615, 625
Bessie Lake Formation ... 126
Best Bet deposit, N.W.T. ... 217
Bethlehem (58) mine, B.C. ... 318, 512
Betts Cove ... 325
(8) mine, Nfld. ... 327
Bevcon (37) mine, Que. ... 205, 206
Bicroft (108a) mine, Ont. ... 168
Bide Arm Formation ... 250
Big Beaverhouse, Ont. ... 732
Moraine ... 733
Bend, B.C. ... 451, 504, 756
Bull (4) mine, B.C. ... 504
Chief–Little Chief (10), B.C. ... 510
Ledge deposit, B.C. ... 506
Salmon Complex ... 416, 424
Snowy Group ... 420
Spruce Lake ... 76
Valley Formation ... 410
(91a) oil field, Alta. ... 532
Bignell deposit, Que. ... 183
Billhook Formation ... 443
Billings Formation ... 259, 260, 354
Bindloss (128) gas field, Alta. ... 539
Bingham deposit, Utah ... 318
Biochronology ... 592
Birch Creek Schist ... 424
Island ... 518
Lake, Ont. ... 67
(9) deposit, Sask. ... 198
Birchlake Member ... 783
Birchtree Lake orebody, Man. ... 181
Bird Fiord Formation ... 550, 561, 564–566, 616
(Oiseau) River ... 183
Birdbear Formation ... 409, 410
Bishopric, Sask. ... 519
Bismuth deposits ... 217, 316
Bisset area, Man. ... 207
Bitumen deposits ... 539, 589
Bjorne Formation ... 576, 578, 579, 582, 583, 589
Peninsula ... 572, 574
Blaa Mountain ... 574, 583, 586
Formation ... 576, 578, 583, 636
Black Cape Formation ... 265
Creek ... 345
salt ... 407
Donald mine, Ont. ... 222
Lake ... 337, 339

PAGE

Black River Group ... 255, 257, 344, 346, 354
Shale Brook Formation ... 240
Blackley Member ... 559, 561, 564
Blackstone Formation ... 461, 462
Blairmore ... 444
Group ... 456–458, 460, 462, 463
Blake River Group ... 60–62
Blind Fiord Formation ... 576, 578, 583
River ... 109, 112, 115, 153, 156, 161–165
Blomidon Formation ... 298
Blood Reserve Formation ... 464
Blow-me-Down Brook Formation ... 247
Blue Fiord Formation ... 561, 563, 564, 566, 589, 616
Mountain ... 128, 222
Formation ... 259
nepheline syenite ... 131, 222
Pinion Formation ... 234
Bluebell (91) mine, B.C. ... 506, 788
Blueberry (9) oil and gas field, B.C. ... 537
Bluesky Formation ... 459
Bog iron deposits ... 517
Bois Blanc Formation ... 144, 277, 777
Boisdale granite ... 253
Boissevain Formation ... 464, 479
Bonanza Group ... 435, 442, 443, 454, 636
Bonaparte Lake ... 442
Bonaventure Formation ... 283, 292
Bonavista Bay ... 240, 283, 708
Formation ... 239
Bone Creek–Instow (19) oil field, Sask. ... 538
Bonne Bay ... 242, 317, 708
Bonnet Plume Basin ... 452, 473, 474
Formation ... 461, 473
Bonnie Glen (60a) oil field, Alta. ... 533
Boothia Arch ... 549
Peninsula ... 100, 554, 558, 663, 671, 743, 744
Plain ... 19
Plateau ... 13, 19
Uplift ... 548–550, 555, 557, 559–561, 565, 567, 572, 584, 587, 589
Borden, P.E.I. ... 710, 711
Island ... 18, 575, 578, 579, 583, 698, 699, 751
Formation ... 575, 579
Peninsula ... 698, 747, 748
Plate ... 47, 53
Subprovince ... 101
Boreal Forest Region ... 680
Borradaile Member ... 459
Borsato Formation ... 405
Borup Fiord Formation ... 571
Boss Mountain (33) mine, B.C. ... 514
Point Formation ... 293, 772
Bothwell oil field, Ont. ... 307, 345
Botwood Group ... 251, 266, 267, 276, 282
Bouger Gravity Anomaly Map of Canada (1257A) ... 5
Boulder Creek Member ... 458
Bouleaux Formation ... 264, 605
Boundary Intrusions ... 198
Lake Member ... 436, 537
oil field (17), B.C. ... 537
South gas field (5), Alta. ... 537
Mountain anticlinorium ... 246

PAGE

Bourinot Group ... 240, 268, 325
Bourlamaque, Que. ... 191
batholith ... 205, 206
Boutellier non-glacial interval ... 705
Bouzan (3a) mine, Que. ... 195
Bouzan-Cadillac break ... 204
Bow Island Formation ... 460, 539, 540
River ... 403, 435, 444, 498, 524
Valley ... 419
Bowser, B.C. ... 788
Basin ... 438, 440, 442, 443, 448, 454, 654
Group ... 442, 443, 454
Box (5) mine, Sask. ... 208
Boyne Mountains ... 464
Bradore Formation ... 240, 241
Bradtville Till ... 686, 687
Braeburn gas field, Alta. ... 537
Bralorne–Pioneer mine (36), B.C. ... 507, 509
Bras d'Or Lake ... 680, 709
Brazeau Formation ... 464, 540
Brébeuf, Que. ... 222
Brent Crater ... 137
Brew Group ... 453
Brewster Limestone Member ... 436
Brian Boru Formation ... 454
Bridge River, B.C. ... 434, 516, 518
Bridgewater, N.S. ... 300
Brigus Formation ... 240
South Head ... 334
Brisco mine (101), B.C. ... 518
Bristol–Masham area, Que. ... 128
Britannia Formation ... 512
mine (56), B.C. ... 504, 510, 512, 518
British Canadian mine (61a), Que. ... 337
Empire Ranges ... 17
Mountains ... 18, 21, 23, 25, 366, 367, 373, 399, 400, 411, 418, 435, 444, 551
Broadview Formation ... 406, 415
Brock (9) gas field, Sask. ... 540
Inlier ... 45, 53
Island ... 18, 575, 583, 751
Plain ... 81
Subprovince ... 77, 80
Brocket Till ... 692
Brockville, Ont. ... 727
Brodeur Peninsula ... 19, 697, 698, 747–750
Brokenback Hill Formation ... 454
Brompton Formation ... 246
Brooklyn limestone ... 510
Brooks Geanticline ... 400, 438, 448, 450, 455
Peninsula ... 27
Range area ... 23, 25
Broughton Soapstone and Quarry (63a), Que. ... 340
township, Que. ... 336
Broulan (53a) mine, Ont. ... 203
Browns Bank ... 29
Mountain Group ... 247, 255, 258, 331
Bruce county, Ont. ... 347
Formation ... 112, 115
Group ... 156
Lake ... 155, 175
Mine, Ont. ... 109, 112–115

PAGE

Bruce Mountains ... 698
Peninsula ... 27, 269, 271
Bruneau mine, Que. ... 195
Brunswick No. 6 (3a) mine, N.B. ... 320, 324
No. 12 (3b) mine, N.B. ... 320, 324
Brynnor (52) mine, B.C. ... 515
Bubbles (5a) gas field, Alta. ... 537
Buchan Gulf ... 30
Buchanan Lake ... 581
Buchans, Nfld. ... 309
Group ... 325
(13) mines, Nfld. ... 267, 325
Bucke Formation ... 257
Buckingham, Que. ... 221
Buckinghorse Formation ... 458
Buckles mine, Ont. ... 166
Buffalo Ankerite (50b) mine, Ont. ... 203
Lake Till ... 692
Bug Creek Formation ... 444
Bulkley River ... 522
Bull Arm Formation ... 234, 330
Bullhead Group ... 458, 524
Bunde Fiord ... 579, 586
Burdett oil field, Alta. ... 538
Burgeo, Nfld. ... 708
Burgess Mountain Formation ... 614
Buried organic deposits ... 679–705
Burin Peninsula ... 234, 282, 283, 335, 708
Burleigh gneiss ... 131
Burlington Geanticline ... 244, 252, 257, 261, 262, 266
granite ... 325
granodiorite ... 252, 267
Peninsula ... 230, 237, 267, 327
Burnais Basin ... 394
Formation ... 405, 519
Burnett Inlet ... 557
Burnt Hill mine, N.B. ... 335
Jam Brook Formation ... 264, 605
Timber Creek ... 479
Burrard Formation ... 460, 476
Burtons Head Group ... 255
Burwash placer field, Y.T. ... 517
Buttle Lake ... 417
Byam Martin Island ... 701
Bylot Island ... 14, 749

Cabano Formation ... 263, 605
Cabot Group ... 234
Head Formation ... 269, 271, 605
Strait ... 706, 713
Cache Creek Group ... 415–417, 423, 431, 434, 442, 443, 454, 518
Cadillac area, Que. ... 60
break ... 204, 205
fault ... 60, 63
Group ... 61, 63
Cadmium deposits ... 217, 316
Cadomin, Alta. ... 520, 524
Formation ... 457–459
Cadotte sandstone ... 540
Cadwallader shear zone ... 509
Caesium deposits ... 217

PAGE

Cairn Formation ... 403–405, 409
Caldwell Group ... 237, 246
  syenite ... 53
Caledonia, Ont. ... 780
  (114) mine, Ont. ... 342
Caledonian mountain chain ... 229
  Upland ... 293
Calgary, Alta. ... 754
California ... 649
Calmar siltstone ... 410
Calumet Member ... 408
Cambrian ... 235–243, 375–385, 552–555, 593–600
  Lake ... 104
Cameron Bay Group ... 77–80
  Hills ... 20
  Islands ... 561, 575
  River ... 73
Campbell Chibougamau (5b) mine, Que. ... 195
Campbell Lake ... 688
  (9b) mine, Ont. ... 206
  phase lake ... 733
Campbell–Sitidgi Lakes ... 742
Camrose Formation ... 410, 533
Camsell Formation ... 399
  River deposits, N.W.T. ... 215
Canada Basin ... 552
  Bay ... 241, 250
  Head Group ... 248
  Tungsten (1) mine, N.W.T. ... 516
Canadian Charleston mine ... 757
  Dyno (108b) mine, Ont. ... 168
  Malartic (32c) mine, Que. ... 204, 205
  Shield ... 43–226
    Hydrogeological Region ... 780, 781
    mineral production ... 152
Candego mine, Que. ... 328
Can–Met mine, Ont. ... 166
Canmore coalfield, Alta. ... 524
Cannes de Roches Formation ... 292
Cannings Cove Formation ... 234
Cannonball Formation ... 479
Canoe Brook Formation ... 240
Canol Formation ... 400, 401, 528
  Road ... 518
Canon Diablo meteorite ... 134
Canrobert Formation ... 557
  Hills ... 559, 572
Canso Group ... 286, 289–291, 329, 626, 663
Cantley mine, Que. ... 220, 222
Cantuar Formation ... 460
Canyon Creek Formation ... 385
  Fiord ... 560, 561, 563, 572, 573, 578, 582, 585
    Formation ... 571, 572
  Ranges ... 24
Cap-des-Rosiers Est ... 712
Cape Aston ... 698
  Bathurst ... 18, 20, 696, 742, 743
  Bon Ami Formation ... 273, 274, 317
  Breton Development Corporation ... 37
    Highlands ... 28, 708
  Chidley ... 29
  Christian ... 698

PAGE

Cape Clay Formation ... 557
  Columbia Group ... 552, 560
  Crauford Formation ... 557, 558
  Crozier anticline ... 551
  de Bray Member ... 564
  George ... 29, 258
  Henrietta Maria Arch ... 137, 140, 142
  Ingersoll Formation ... 555
  Jones ... 713
  Kent Formation ... 555
  la Hune ... 276
  Leiper Formation ... 555
  Levvel ... 586
  M'Clure ... 552
  Parry ... 20
  Phillips Formation ... 550, 557, 560, 563, 601, 616
  Rawson Formation ... 555, 558, 561, 563, 578, 606
  St. John Group ... 251, 267
  St. Mary ... 268
  Smith ... 16
    Fold Belt ... 47, 52, 53, 86, 87, 104, 105, 107, 121, 156, 174, 182, 198
    Subprovince ... 104
  Spencer ... 296
  Split Trough ... 296, 297
  Tormentine ... 711, 712
  Wolstenholme ... 736
  Wood Formation ... 555
Capital expenditures ... 34, 38
Capreol, Ont. ... 175
Carbon Creek ... 654
  basin ... 524
  synclinorium ... 455
Carboniferous ... 284–296, 356–362, 411–421, 568–572, 622–632, 663–670
Carcajou Ridge ... 615
Cardiff, Ont. ... 221
  complex ... 168
Cardium Formation ... 462, 463, 540
Carey–Canadian (64) Mines Limited, Que. ... 337
Cariboo (30) district, B.C. ... 517
  Gold Quartz (32) mine, B.C. ... 507, 509
  Group ... 415, 509, 517
  Lake ... 435
  Mountains ... 20, 375, 381, 406, 411, 415, 450, 472
  River ... 435
Caribooan Orogeny ... 368, 406, 411, 415, 424
Carleton Township, Que. ... 316
Carlsbad Formation ... 260, 354
Carmacks, Y.T. ... 443, 478
  coalfield, Y.T. ... 522
  Group ... 478
Carmanah Formation ... 479
Carnarvon Member ... 420
Caron Lake fault ... 125
Carswell Formation ... 94, 137
  structure ... 137
Cartridge Mountains ... 20
Cascade Complex ... 428
  Fold Belt ... 368, 427, 428, 467
  Mountains ... 26, 394, 417, 428, 431, 434, 441, 443, 453, 468, 476, 477, 702, 754–756

PAGE

Cascapedia River ... 318
Casey quartz monzonite ... 515
Cass Fiord Formation ... 555
Cassiar, B.C. ... 518
  Asbestos (5) mine, B.C. ... 518
  batholith ... 450, 451, 463, 518
  Mountains ... 21, 25, 375, 378, 380, 382, 389, 391, 399, 402, 416, 417, 424, 434, 451, 461, 754–756
  Platform ... 375, 378, 389, 394, 402
Castle River ... 479
Castor coalfield, Alta. ... 524
Cat Lake ... 45
Cat Lake – Winnipeg River district, Man. ... 217
Catamaran fault ... 257
Cataract Group ... 142, 268, 271, 353
Catfish Creek Till ... 687
  mine, B.C. ... 512
Cathedral Formation ... 384, 385, 495, 518
  Mountain ... 614
Catskill Delta ... 272
Caution Creek Formation ... 139
Cecilia Lake ... 604
Cedar Barn volcanic member ... 264
  Lake Formation ... 143, 605
  Point phase lake ... 728
Cedared Formation ... 405
Celestite deposits ... 221
Cenozoic ... 296–301, 464–481, 584–588, 663–674
Central Ellesmere Fold Belt ... 551, 566, 568, 589
  Manitoba (20) mine, Man. ... 207
  Montana Uplift ... 394
  Patricia (4) mine, Ont. ... 207
  Plateau ... 708
  Plateau and Mountain area ... 25
  shoals ... 15
  St. Lawrence Lowland ... 27, 29
Centre anticline ... 557
  Peak anticline ... 469
Cessford (124a) gas field, Alta. ... 539, 540
Chaba Creek ... 384
Chakonipau Lake ... 104
Chaleur Bay ... 265, 276, 283, 318, 711, 712
  Uplands ... 29, 711
Chaleurs Bay Group ... 247, 264, 324
Challenger Group ... 560
Chamberlains Brook Formation ... 240
Chambly, Que. ... 780
Champlain Sea ... 671, 726–729, 777, 779
Chancellor Formation ... 385
Change Islands ... 266
Chantrey Group ... 100
  Inlet ... 745, 747
Chapel Cove ... 316
  Island Formation ... 234
Chapleau, Ont. ... 66
Chaplin Lake, Sask. ... 519
Chapperon Group ... 417
Charles Cove ... 336
  Formation ... 420
Charlie Lake Formation ... 436, 537

PAGE

Charlotte County, N.B. ... 617
  Group ... 247, 251, 265, 276
Charny Formation ... 236
Chasm Creek Formation ... 139
Chateau Richer, Que. ... 222
Chatham Sag ... 283, 284
Chatham–Grenville stock ... 122, 128, 130
Chaudière River ... 307, 330, 713, 714
Chaumont Formation ... 602
Chazy Group ... 250, 344
Cheakamus Formation ... 460
Cheam Lake, B.C. ... 522
Chedabucto Bay ... 297, 298, 709
  Lake ... 605
    Formation ... 388
Cheddar granite ... 168
Cheecham Hills ... 20
Chelmsford Formation ... 117
Chemahawin Member ... 143
Chemalloy Minerals mine, Man. ... 217, 219, 220
Chemical composition, Canadian Shield ... 59
Cheno River Formation ... 59
Cherabusco, N.Y. ... 727
Chesterfield Inlet ... 13
Cheverie Formation ... 287, 328
Chibougamau, Que. ... 36, 109, 121, 153, 155, 185, 194, 195, 737
  Group ... 59
Chicago, Ill. ... 730
Chicotte Formation ... 272
Chilko Lake ... 431
Chilliwack, B.C. ... 522
  batholith ... 476, 477
  Group ... 417, 434
Chimney Arm Formation ... 250
  Corner coalfield, N.S. ... 356
Chinchaga Formation ... 406, 614
Chischa Formation ... 372
Chisel Lake (8) mine, Man. ... 198, 200
Chitina Valley ... 435
Chitistone limestone ... 435
Christie Street sand pits ... 685
Christina Member ... 408
Christopher Formation ... 582–584, 665
  Island Formation ... 99
Chromite deposits ... 183, 336
Chuckanut Formation ... 477
Chukotat Group ... 104, 105
Chungo Mountain ... 463
Churchbridge, Sask. ... 694
Churchill, Man. ... 140
  Province ... 46, 85–107, 155
  River Group ... 139, 140, 605
Chushina Formation ... 391, 603
Cinder cone ... 23
Cirrus Mountain ... 472
Citadel Formation ... 254
Clam Bank Formation ... 272, 278, 282
Claremont Formation ... 293
Clarenville Formation ... 249

PAGE

Clarke Lake (3) gas field, B.C. 527
Range 371
'Classical' Wisconsin Glaciation 677, 686, 702, 703, 705–756
Clausen Formation 418
Clay deposits 521
Clear Creek, Y.T. 504
placer field, Y.T. 517
Hills, iron deposits, Alta. 498
gas field 537
Lake outlier 260
Clearville oil field, Ont. 346
Clearwater Formation 458, 459, 539
Lake 138, 139
Crater 136
River 746
Valley 788
Clements–Markham Inlet 752
Clemville Formation 264, 605
Climax deposit, Colorado 318
Clinton Creek deposit, Y.T. 518
Group 269, 353, 776
Lake 746
Cloud Mountains Formation 241
Rapids Formation 241
Cloudmaker Formation 391
Clyde, N.W.T. 698
Clyde phase lake 748, 750
Coal deposits 145, 356–362, 523, 524, 589
River 478
Coalspur, Alta. 524
Coast Geanticline 369, 411, 423, 424, 428, 431, 435, 438, 440, 441, 443, 448, 453, 455, 460, 474
Mountains 21–26, 417, 427, 431, 432, 441–443, 453, 454, 460, 467, 476, 479, 481, 704, 754–756, 786
Plutonic Complex 367, 368, 417, 423, 424, 427, 440, 441, 443, 454, 460, 467, 512, 516
Coats Island 15, 137, 138, 140
Cobalt deposits 182, 183, 208–212
(96–98) District, Ont. 210
Group 108, 112, 115, 210–223
Lake fault 210
Plain 13, 16
Plate 45, 47, 52
Subprovince 155
Cobequid Bay 28, 710
fault 710, 712
Mountains 28, 268, 276, 277, 292, 293, 332, 617, 709, 710
Upland 288, 292–296
Coboconk Formation 256
Cobourg Formation 256, 259
Cochenour (9c) mine, Ont. 207
Cochrane, Ont. 733
ice advance 732, 734
recession 734
Till 734
Cockburn Moraine System 748
phase lake 748, 750
Codroy Group 289, 292, 342

PAGE

Coffin Island 711
Colchester oil field, Ont. 346
Cold Lake, Alta. 460
salt 406
Coldbrook Group 233, 234, 253, 279
Coleman, Alta. 457, 524
Formation 210
Coleraine, Que. 336
Coleville sandstone 419
Coleville–Smiley (6) oil field, Sask. 534, 540
Collingwood Formation 259, 260, 346, 602
Colorado Group 459, 460, 461, 463, 540
Colpoy Bay Member 270
Columbia Highlands 26
Icefield 25
Lakes 382
Mountains 26, 467, 479, 756, 786
Plateau 754
River 451, 504, 518, 745, 755
Columbian Orogeny 367, 369, 439, 448–453, 455, 464, 468, 470, 518
plutonic rocks 422–428
Zwischengebirge 368, 450, 452, 460, 467
Colville Hills 20, 418
Trough 438, 444
Committee Bay 19, 745, 747
Commotion Formation 458, 459
Comox, B.C. 788
coalfield, B.C. 522
Group 460
Compton Formation 273
Con (9a) mine, N.W.T. 207
Rycon (9c) mine, N.W.T. 207
Conception Bay 249, 279, 280, 331, 334, 706
Group 232, 234
Coniagas (8) mine, Que. 200
Coniaurum (51) mine, Ont. 203
Connaught varved sediments 734
Connecting Point Group 234, 330
Connor Lake 419
Conrad (147) oil field, Alta. 538
Consolidated Discovery (6) mine, N.W.T. 208
(Hardrock) (31a) mine, Ont. 206
Rambler (6) mines, Nfld. 327
Construction materials 36, 37
Contact Lake 215
Rapids Formation 406
Continental Arch 394
Ranges 25
Contwoyto Lake 72, 76, 745
Cooking Lake Formation 409, 531
platform 528
Cooks Brook Formation 238, 247
Copes Bay 586
Formation 557
Copper Cliff Formation 109
deposits 185–189, 317–327, 498, 510, 589
Mountain 495
granite 318
(65) mine, B.C. 512
(72) mine, Que. 316–318, 335
stock 423, 440

PAGE

Copper Rand (3b) mine, Que. ... 195
smelters and refineries ... 37
Coppercorp (68) mine, Ont. ... 196
Copperfields deposits, Ont. ... 186
Coppermine Arch ... 378
deposits, N.W.T. ... 199
Group ... 76, 77, 81, 82, 88, 199
Hills ... 14
Homocline ... 45, 53
River ... 81, 161, 186
Subprovince ... 77, 80, 369
Coquitlam, B.C. ... 704
Coral Rapids, Ont. ... 145, 734
Corbin, B.C. ... 524
Cordilleran Glacier Complex ... 726, 741, 753–756
Hydrogeological Region ... 786–788
Orogen (Geosyncline) ... 366–383, 385–406, 411–420, 423–435
Region ... 13, 17–20, 464
Cordova Bay ... 704
(112) mine, Ont. ... 208
Corner Brook ... 250, 252
Cornwall anticline ... 550
Island ... 575, 579, 583, 585
Cornwallis Fold Belt ... 549, 550
Group ... 550, 557–559, 567, 586, 604
Island ... 550, 551, 555, 557, 558, 561, 562, 584, 587–589
Cornwallis–Annapolis Valley ... 709
Coronation Gulf ... 14, 80, 81, 745
Hills ... 14
(10) mine, Sask. ... 198
Corral Creek Formation ... 375
Corundum deposits ... 221
Coryell Intrusions ... 476, 478, 518
Costigan Member ... 404
Cottonbelt deposits, B.C. ... 506
Couchiching Group ... 66, 67
Coulombe deposit, Que. ... 183
Covey Hill outlet ... 727
Cow Head Group ... 238, 248, 250
Peninsula ... 248
Cowichan Bay, B.C. ... 788
Lake ... 435
Coxheath Hills granite ... 253
Craigmont mine (57), B.C. ... 510
property, Ont. ... 221
Craignish Formation ... 287
Cranbourne Group ... 263
Cranbrook, B.C. ... 380, 518, 519
Formation ... 380, 518
Cree Lake Moraine ... 745
Creighton granite ... 116, 117
mine, Ont. ... 154
Crescent Lake Formation ... 258
Moraine ... 732
Mountains ... 733
Creston Formation ... 371, 380
Cretaceous ... 145, 297–299, 445–464, 574, 575, 582–583, 649–670
Croker Island Complex ... 116
Cross Lake ... 45, 57, 69, 165

PAGE

Cross Lake Group ... 57, 69
Subprovince ... 68, 70, 71
Point ... 518
River ... 518
Crossfield dolomite ... 534
(107) oil field, Alta. ... 540
Croteau Lake Group ... 120, 126
Crowsnest coalfield, B.C. ... 522, 524
Formation ... 457, 460, 462
Lake ... 520
Pass ... 25, 444, 457, 524, 625, 754
Cruiser Formation ... 459
Crystal Peak Formation ... 416
Cultus Formation ... 434, 442
Cumberland Basin ... 284, 287, 288, 290, 292–296
Group ... 287, 292–295, 356
Lowland ... 28
Peninsula ... 29
Sound ... 15, 747, 748
Cummings Member ... 459
Cunningham Formation ... 415
Limestone ... 375, 381
Cupra (56a) mine, Que. ... 320
Cuprus (5) mine, Man. ... 198
Custer gneiss ... 428, 477
Cutler granite ... 108, 116, 117
Cypress Hills ... 20, 464, 479, 481, 690, 693, 784
Formation ... 481

Daiber Group ... 436
Dakota ... 408
Dalhousie Formation ... 275, 276
Sandstone ... 456
Daly (5) oil field, Man. ... 534
Dana Inlet facies ... 477
Daniels Harbour ... 329
Dantzig Group ... 234
Darnley Bay ... 654, 741, 747
Peninsula ... 743
Dashwood drift ... 704
Davidson halite ... 409
Davis Highland ... 10, 13, 14
Inlet ... 120
Region ... 14, 16, 17
Shelf ... 29
Strait ... 29, 747
Dawn (156) gas field, Ont. ... 348
reef ... 351
Dawson, Y.T. ... 22, 424, 517, 518, 522, 654, 790
Bay Formation ... 144, 270, 271, 408, 614, 615
Creek, B.C. ... 537, 539
River ... 790
Strait ... 654
D'Clute gas field, Ont. ... 348, 350
Deadmans Cove Member ... 267
Deadwood Formation ... 384, 391, 522
Dear Lake ... 237
Dease Creek, B.C. ... 517
River ... 756
Deas Thompson Point ... 81
Debolt Formation ... 418, 420, 537

PAGE

Decew Member ... 270
Deer Bay Formation ... 581–583
Horn mine, Ont. ... 210
Park Pond Formation ... 232, 234
DeGeer moraines ... 736
Degerbols Formation ... 573, 574
Delaware limestone ... 617
Del Bonita Hills ... 693
Delnite (50c) mine, Ont. ... 203
Delorme Formation ... 388, 399, 606, 614
Deloro (111a) mine, Ont. ... 208
Denali fault system ... 369, 476
Denault Formation ... 102
Denison (74a) mine, Ont. ... 165, 166, 781
Department of Energy, Mines and Resources ... 4
Derry mine, Que. ... 221
Deschaillons, Que. ... 682
Deschambault Formation ... 256
Destor fault ... 63
Detroit River Group ... 617, 777
Deville Member ... 459
Devon Ice Cap ... 747
Island ... 14, 19, 30, 101, 552, 555–559, 564, 568, 584, 587, 603, 751
Formation ... 559
Devonian ... 143–145, 272–284, 392–410, 557–568, 614–622, 663–670
Dewdney Creek Group ... 442, 453, 454
Dezadeash area, Y.T. ... 478
Group ... 455
Diapiric intrusions ... 18, 583, 586, 589
Diatomite deposits ... 521
Diatreme breccia ... 259, 272, 277, 278, 298
Dibble Creek fault ... 380
Dickenson (9d) mine, Ont. ... 206
Dickstone deposit, Sask. ... 198
Digby, N.S. ... 710
Digges Islands ... 15
Dina Member ... 459
Dip-needle surveys ... 185
Director (15b) mine, Nfld. ... 340
Disappointment Bay Formation ... 550, 560–562, 564
Discovery Point ... 588
Dismal Lakes ... 14, 199
Disraeli Fiord ... 752
Dixon Entrance ... 27
Dodger mine, B.C. ... 516
Dog Lake Moraine ... 731, 732
River ... 732
Dogtooth Range ... 375, 380, 381
Doig Formation ... 436
Dollard (21) oil field, Sask. ... 538
Dolly Varden (12a) mine, B.C. ... 506
Dolomite deposits ... 344, 520
Dome (50e) mine, Ont. ... 203
Don Formation ... 681, 684, 685, 689
Donjek Glacier ... 756
Range ... 435
River ... 756
Donnacona, Que. ... 682
Don Valley brickyard ... 684, 685
Dore Lake fault ... 195

PAGE

Dorion beach ... 730
Doten Cove Formation ... 234
Double Mer Formation ... 127
Doublet Group ... 102
Doucers Formation ... 250, 267
Douro Formation ... 559, 606
Dover oil field, Ont. ... 346, 347
Dowling Member ... 463
Driftwood Bay Formation ... 562, 563
Drumheller badlands ... 464
coalfield, Alta. ... 524
front ... 409
Member ... 464
Drummond pit, N.B. ... 333
Drummondville Moraine ... 713
Dryden, Ont. ... 218
Dubawnt Group ... 59, 76, 88, 94, 98
Lake ... 159, 160, 745
River ... 746
Duck Hawk Bluff ... 700
Mountain ... 693
Ducktown deposit, Tennessee ... 317
Dufault Lake granodiorite ... 191
Duhamel (85a) oil field, Alta. ... 528
Duke River thrust ... 475
Duluth gabbro ... 109
Duncan Formation ... 109
Lake deposit, B.C. ... 506
Dundas shale ... 344
Dundee Formation ... 144, 277, 283, 344, 351, 353
Dunvegan Formation ... 459, 461, 463, 540
Dunwich Till ... 687
Duperow Formation ... 409
Duthie (21) mine, B.C. ... 506
Duvernay, Alta. ... 519
Formation ... 410
Dyer Bay Formation ... 269, 271, 605
Dyno mine, Ont. ... 164

Eager Trough ... 375, 378, 380, 383
Eagle granodiorite ... 454
Plain ... 23, 455, 473, 542
Formation ... 461
River ... 755
Eagle-Finlayson Moraine ... 731
Earth Physics Branch ... 4, 5
Earthquakes ... 300, 474, 477
East Arm Fold Belt ... 47, 52–54, 85, 94
Formation ... 242, 605
Hills ... 11, 13
Subprovince ... 87, 94
Braintree–West Hawk Lake district, Man. ... 218
Branch Pond Formation ... 257
Cape ... 571
Jersey orebody, B.C. ... 514
Kootenay Orogeny ... 368, 371, 373, 374, 424, 450
Malartic (32d) mine, Que. ... 204, 205
Pacific Rise ... 476
River ... 307
Waite orebody, Que. ... 191

PAGE

Eastend Formation 464, 783
(21b) oil field, Sask. 538
Eastern Columbides 450–452, 461
Cordilleran Orogen 467–473
Metals deposit, Que. 317
Mineral Province 491, 494–498
Quebec Uplands 29
System 21, 23
Eastmain Lowland 16
Eastview Formation 259, 260
Eau Claire deposit, Ont. 222
Formation 242
Echo Bay Group 77–80, 168, 215
(2) mine, N.W.T. 199, 215
Echoing River 733
Echooka Member 418
Economic growth 2, 33
Edgecliff Member 144, 277
Edmonton, Alta. 463, 479, 519, 695, 738, 740
Formation 461, 464, 478, 521, 524, 540, 783
Eglinton Fiord 14
Graben 551
Island 583, 699
Eids Fiord 572
Formation 560, 561, 563
Ekwan River Formation 142, 144
Eldon Formation 384, 385
Eldorado Creek 790
(3) mine, N.W.T. 163, 168, 199, 215
Mining and Refining Company 3
Eleanor River Formation 557
Electric oil field, Ont. 346
Elise Formation 442
Elk conglomerate 444
Lake 209
Point Basin 394, 406, 407
Group 406, 409, 519, 527, 614
River 435, 452
Elko Formation 405
Elkton Member 420, 535
Elkwater Drift 693
Ella Bay 552
Ellef Ringnes Island 18, 552, 579–584, 586–588, 698, 699, 751
Ellenburg, N.Y. 727
phase lake 727
Ellerslie Member 459
Ellesmere Fold Belt 551, 566–568, 589
Island 14, 17, 19, 101, 551–589, 698–701, 751–753, 788
Ellesmerian Orogeny 367, 394, 399, 411, 418, 551, 566–569
Ellice River 13
Elliot Cove Group 240, 249
Lake, Ont. 36, 39, 109, 112, 115, 152, 154, 163, 781
(72-75) district, Ont. 164–167
Group 108, 109
Ellis Bay Formation 261, 271
Elm Point Formation 144, 520, 615
Elmira, Ont. 778
Elmtree Group 325

PAGE

Elsie Mountain 112
Elsonian Orogeny 49, 119–125, 127–131
Emerald Bank 29
Basin 29
(80a) mine, B.C. 516
Emma Fiord Formation 569, 571, 663
Empetrum Formation 460
Empire (40) mine, B.C. 515
Endako Group 478, 479
(28) mine, B.C. 318, 514
quartz monzonite 515
Enderby, B.C. 788
Englee Formation 250
English Head Formation 261
River Belt 45, 53, 54
Subprovince 57, 67, 68
Englishmans Group 416
Ennadai belt 100, 155, 156, 159
Lake 96, 744, 746
Enniskillen township, Ont. 307
Enrage Formation 293
Entrance conglomerate 464, 479
Entry Island 711
Epworth Group 76–80
Eqalulik Group 88, 101
Ernestina Lake Formation 406
Escarpment Member 409
Escuminac, Que. 617
Formation 283, 663
Eskimo Lakes 696, 742
Espanola, Ont. 109, 112, 113, 115
Formation 112
Esperanza Inlet 443
Essex county, Ont. 345, 346
Estella (106) mine, B.C. 495
Esterhazy (14) mine, Sask. 520
Estevan coalfield, Sask. 524
Coastal Plain 27
Etanda Formation 418
Etherington Formation 419, 420
Et-Then Group 94, 96
Eureka Sound 579, 583, 751
Formation 568, 583–586, 589
Upland 28
Eurekan Orogeny 551, 568, 569, 575, 584–586, 588
Eustis–Capelton mines (53b), Que. 320
Evans Creek Stade 754
Evie Member 407
Exmouth Island 575
Expedition Fiord 583
Exploits Group 247, 255, 258
Exports 34, 35, 41
Exshaw, Alta. 520
Formation 404, 410, 419, 522

Fabricius Fiord Formation 101
Fairholme Group 403, 405
Falher Member 459
Falkland (62) mine, B.C. 518
Falsen Island 557
Fantasque Formation 418

PAGE

Faraday (107a) mine, Ont. ... 164, 168
granite ... 168
Farr Formation ... 257, 260
Favel Formation ... 464
Federal mine, Que. ... 329
Feeney mine, B.C. ... 516
Feldspar deposits ... 221
Fenelon Falls outlet ... 728
Fennell Formation ... 431
Ferguson Lake ... 155
Fergusson fault ... 509
Group ... 417, 509
Fernie, B.C. ... 444, 455, 521
Basin ... 444, 456
Formation ... 444, 445, 455, 522, 641
Field, B.C. ... 385, 495
Findlay Arch ... 283, 345, 348
Finlayson Lake ... 67, 424
Fire Lake Group ... 454
Firebag Member ... 408
First (Upper) White Specks ... 463, 464
Firth River ... 25, 435
Fish Hook Bay prospect, Sask. ... 167
Scale sand ... 459–464
Fisher Branch Formation ... 605
Fisset Brook Formation ... 287
Flagler Fiord ... 557
Flat Bay Brook anorthosite ... 333
(14) mine, Nfld. ... 342
River ... 516
Sound ... 583
Flathead fault ... 470, 481
Range ... 405
River ... 405, 470, 480, 524
Flatwater Pond ... 267
Flavian Lake granite ... 191
Fleming Formation ... 102
Flemish Cap ... 29
Flett Formation ... 418
Fleur de Lys, Nfld. ... 318, 329, 335
Group ... 237, 239, 247, 252, 279, 280
Fleurant Formation ... 283
Flexspar deposit, Sask. ... 198
Flin Flon, Sask. ... 90–92, 153, 155, 185, 196, 198, 208, 217, 220
belt ... 90, 155
(3) mine, Sask. ... 190, 198
Flin Flon–Herb Lake belt ... 208
Flinkote Mines Limited (62b), Que. ... 337
Flume Formation ... 404
Fluorite deposits ... 221, 340
Fogo Island ... 266
Foothills ... 20, 24, 367, 404, 435–438, 444, 445, 455–463, 467–472, 478, 479
Fording River fault ... 470
Foremost Formation ... 463, 464, 524
(149a) gas field, Alta. ... 540
Foremost–Cypress Hills area ... 693
Forillon Peninsula ... 273, 274
Formosa reef ... 277

PAGE

Fort Anne phase ... 713, 727
Augustus Formation ... 459
Covington ice ... 727
Erie, N.Y. ... 727
Fitzgerald, Alta. ... 519
Fraser, B.C. ... 514
McMurray, Alta. ... 408
Nelson, B.C. ... 389
Arch ... 394
Formation ... 463
River ... 20, 741
Norman, N.W.T. ... 461, 473, 478
Qu'Appelle, Sask. ... 694
Selkirk, Y.T. ... 703, 705
Simpson ... 408, 410
Formation ... 402, 408–410, 742, 790
Smith, N.W.T. ... 790
St. James, B.C. ... 434
St. John (19) gas field, B.C. ... 537, 734
Group ... 458
Vermilion Member ... 408
Whyte, Man. ... 520
William, Ont. ... 729
Forteau Formation ... 241
Fortin Formation ... 273, 275
Fortune Bay, Nfld. ... 239, 240, 267, 335, 708
Forty-four (18a) mine, Man. ... 207
Fosheim Peninsula ... 578–584
Fossil Brook Formation ... 240
Hill Formation ... 142, 269, 271
Fossmill Outlet ... 728, 729
Fosterton oil field (15b), Sask. ... 538
Fourchu Group ... 233, 234
Fox Creek ... 537
Lake Belt ... 69, 70, 90
Foxe Basin ... 138, 549, 555, 698, 747–750
Peninsula ... 15, 748
Plain ... 12, 14, 19
Foxe–Baffin Glacier Complex ... 706, 747
Francis Lake ... 424
François Lake ... 515, 518
Franklin Bay ... 743
Mountain Formation ... 381
Mountains ... 24, 367, 373, 374, 378, 381, 388, 389, 401, 406, 418, 455, 461, 472, 528
Franklinian Eugeosyncline ... 555
Geosyncline ... 17, 385, 548, 584
Miogeosyncline ... 555, 589
Orogen ... 568
Fraser Basin ... 26
Canyon ... 26, 518
Delta ... 27, 460, 476
fault system ... 423, 476
Glaciation ... 703–705, 754, 755
Lowland ... 26, 704, 786–788
River ... 26, 443, 472, 477, 479, 490, 517, 735, 754, 756, 788
Trough ... 411
Valley ... 453, 518, 521, 754, 756
Fraserdale, Ont. ... 733, 734
Arch ... 138, 140, 143
Frenchman Formation ... 464, 479

PAGE

Frenchman's Cap Dome 425
Frederickhouse varved sediments 734
Fredericton, N.B. 255, 265, 335, 336, 712, 775
  Trough 261, 265, 268, 272, 276
French River 125, 128
    Formation 268
    granite 131
Frobisher, Sask. 785
  Bay 15, 747, 748
  Upland 15
Frobisher–Alida beds 534
Frontenac Arch 130, 228, 235, 242, 243
  Axis 776
  Group 273, 298
  mine, Ont. 200
  phase lake 727
Frood Formation 109
Fundy Epieugeosyncline 272, 283–285, 295
Funeral Formation 401, 402
Furnace deposit, Que. 183
Fury and Hecla Strait 101

Gabarus Bay 335
Gagnon, Que. 125, 128
Gagonville, Que. 174
Galena Hill 506
Gallery Formation 554
Galore Creek batholith 440
    mine, B.C. 512, 514
Galton Range 371
Gambier Group 454, 512
Gananoque syenite 131
Galets Formation 257, 260
Galetta, Ont. 200
Gander Bay 336
  Lake 280
    Group 239, 247, 251, 255, 258, 267, 279
    serpentine belt 251, 338
Garbutt Formation 459
Garnet deposits 222
Garrington (101) oil field, Alta. 540
Gascons Formation 264, 605
Gaspé, Que. 36
  Bay 29
  Copper Mines (72), Que. 315–318
  Limestone 273, 274, 328, 617
  Sandstone 273–276, 328, 329, 617, 663
  Trough 244, 261–265, 272–276, 345, 354
Gasport Member 270
Gataga River 403
  Formation 372
Gates Formation 458, 459
Gateway Formation 371, 405
Gatineau area, Que. 221
Geco (36) mine, Ont. 193, 194, 200
General Electric deposit, Que. 183
Genex deposit, Ont. 191
Gentilly Till 683
Geographic subdivisions 4
Geological Map of Canada (1250A) 5
  organizations 4, 5
  research 2

PAGE

Geological subdivisions 3, 4
  Survey of Canada Branch 4
Geophysical surveys 53, 54, 64–66, 68, 70, 71, 154, 185, 300, 476, 477, 586
George Bay 710
  Formation 372
  McLeod mine, Ont. 174
  Plateau 15
  River Basin 735
    Group 232, 235, 240, 253, 279
Georgia Basin 369, 448, 460, 475, 786
  Lowlands 26, 786, 788
Georgian Bay 27, 125, 259, 728, 730, 778
Geotectonic correlation charts ..... 5, 6, 592; in folio
Germanium deposits 517
Gething Formation 458, 459, 524
Giant (8) mine, N.W.T. 207
Gilbert River 330
Gilby (95) gas field, Alta. 537
Giles Cove Formation 267
Gillis Brook 331
Gilmour Lake outlier 256
Gilwood Formation 408, 527
Glacial Great Lakes 726
  Lake Agassiz 21, 731–733, 740, 741
    Algonquin 727, 728
    Anteves 734
    Arkona 726
    Barlow-Ojibway 733, 734, 780
    Chicago 726
    Duluth 729
    Edmonton 740, 741
    Falher 741
    Frontenac 727
    Grassmere 726
    Iroquois 727
    Kazan 746
    Kelvin 732
    Lundy 727
    McConnell 741, 742
    McLean 735, 736
    Nakina 733
    Naskaupi 735, 736
    Ojibway 733
    Peace 741
    Penticton 755
    Regina 740
    Rycroft 741
    Tyrrell 741, 746
    Vermont 713, 727
    Warren 726
    Whittlesey 726
  Map of Canada (1253A) 6
Glacier Creek stock 450
  Gulch deposit, B.C. 514
  Peak 755
Gladmar, Sask. 519
Gladstone Formation 456–458
Glen Falls Formation 235, 239, 240
  Lake mine, Ont. 210
Glenbrooke Formation 263
Glencoe Brook Formation 268

PAGE

Glenelg Formation 84
Glenlyon area, Y.T. 424
Glenogle Formation 391, 601
Glenphyll oil field, Alta. 538
Gloucester Formation 260
Goat Island Member 270
Gobles oil field, Ont. 346
Goderich, Ont. 307, 341, 342
(116) mine, Ont. 342
God's Lake 69
River 689
Gog Group 380, 381
Golata Formation 418, 663
Gold deposits 200–208, 330, 498, 507, 517
Eagle (9e) mine, Ont. 207
Golden, B.C. 518
Grove granite 253
Spike (59b) oil field, Alta. 528
Goldenville Formation 240, 249
Goldson Formation 251, 265
Good Friday Bay 582
Hope Group 375
Goodrich Formation 459
Goodsir Group 385, 400
Inlet 701
Goose Cove 316
Formation 238, 248, 251, 252, 316
Fiord Formation 561, 563
River 695
Gordon Lake Formation 114
(13) mine, Ont. 182
Gossage Formation 399–401, 606
Gouin Peninsula 195
Reservoir 128
Goulburn Group 76, 88, 199
Gowganda, Ont. 182, 183, 208, 209, 211
Formation 59, 112, 115
Gowland Mountain 333
Graham Island, B.C. 522, 583
Graminia siltstone 410
Grand Banks 29, 298, 706
Falls 737
Lake 28, 279
Brook Group 237, 238
coal basin 307
Group 279
Le Pierre Formation 267
Manan Island 298
Mira South 331
Rapids Formation 459, 539
River 342, 344, 726
Grande Grève Formation 274, 275, 317
Granduc mine, B.C. 510
Grandview Hills 20
Granisle (22) mine, B.C. 512
Granite wash 527
Grant Lake 78
Grantland Mountains 17
Grantmire Formation 287
Grantsburg sublobe 739
Granville Lake 91
Graphite deposits 222

PAGE

Grass Lake orebody, Man. 181
Gravelbourg Formation 445
Gravity 53, 64, 66, 68, 70, 185, 586
Great Bay de l'Eau Formation 283
Bear Cape 574
Lake 77, 80–82, 153, 154, 159, 163, 168, 186, 209, 215, 388, 418, 741, 742
Plain 20
Island Group 97
Northern Inlier 47, 231
Peninsula 238, 250, 313, 707
Plain of the Koukdjuak 19
Slave Group 88, 94, 95
Lake 13, 36, 47, 73, 85, 92–94, 153, 159, 198, 388, 405–410, 495, 741, 742, 746
Plain 20
Whale River 200
Greely Fiord 573, 578
Green Bank 29
Beds 444, 445
Creek syncline 469
Head Group 232, 234, 253
Mountains 29
Point Formation 247, 248
Greenberry limestone 415
Greenfield, N.S. 300
Greenland 29, 121, 134, 551, 555, 575, 752
Ice Sheet 750
Greenwood, B.C. 36, 417, 431
Gregherd, Sask. 694
Gregory River, Nfld. 316, 317
Grenville, Que. 128
Front 47, 52–54, 104, 108, 121–123, 125, 127, 130, 131
Group 53, 121, 124, 125, 127, 219–223
Province 47, 121–131, 156
Grenvillian Orogeny 49, 121–125, 127–131, 371
Grey Beds 436, 438, 444, 445, 635
River 336
Greyhawk (107b) mine, Ont. 164, 168
Griffin Cove Formation 273, 274
Griffith mine, Ont. 175
Grimsby Formation 269, 351, 605
Grindstone Island 711
Grinnell Formation 371
Peninsula 550, 564, 565, 568, 572, 751
Griper Bay Formation 550, 561, 564–566, 663
Grizzly Bear Hills 20
Grosmont dolomite 408, 409
Gross domestic product (GDP) 34
national product (GNP) 2, 33
Grosse Isle 711
Grotto Member 403, 410
Groundhog coalfield, B.C. 443, 522
Groundwater 766–791
Guelph Formation 142, 143, 270, 344, 348–351, 605
Guichon batholith 440, 442, 443, 454, 512, 514
Guigues Formation 257
Guilmette Lake 729
Gulf of Alaska 369, 474
Boothia 19
Maine 297, 299

PAGE

Gulf of Mexico 448, 479, 575, 654
St. Lawrence 28, 29, 311, 315, 345, 780
Gull Island Formation 255
Lake (18) oil field, Sask. 538
Pond 319
River Formation 256, 601
Gullbridge (12) mine, Nfld. 316, 317, 319, 325
Gun River Formation 271, 605
Gunflint Formation 118, 213
iron range, Minnesota 174
Gunnar (21) mine, Man. 156, 163, 167, 207
Gunton Member 392
Gurney (7) mine, Man. 208
Guyet Formation 415
Gypsum deposits 342, 518, 588
Gypsumville (15) mine, Man. 519

Hadley Bay 549
Hadrynian 51, 53, 81, 84, 88, 127, 231–235, 373–375
Hagersville (115) mine, Ont. 342
Haggart placer field, Y.T. 517
Haida Formation 454, 460
Haines Junction, Y.T. 790
Haldimand county, Ont. 342
gas field, Ont. 351, 354
Haley deposit, Ont. 219
Halfway Formation 436, 438, 537
River 403, 436
Haliburton, Ont. 124, 128
Highlands 124, 130, 131
Halifax, N.S. 300
county, N.S. 336
Formation 240, 249, 268, 316
Hall Formation 442
Lake fault 450
Peninsula 15, 29
Upland 15
Hallnor (53b) mine, Ont. 203
Halton county, Ont. 346
Hambone Lake orebody, Man. 181
Hamill Group 380, 381
Hamilton, Ont. 727
Group 144, 283, 284, 344, 617
Plateau 17
River 737
Upland 17
Hand Hills 481, 690
Hanford Brook Formation 240
Hanson Lake 198, 200
Member 463
Harbour Main Group 231, 232, 234, 239, 340
Harcourt Group 240
Hare Bay 316, 338, 708
Complex 336
klippe 238, 248, 250–252
Fiord 578
Formation 571, 572
Indian Formation 401, 528, 615
Harmattan (104) oil field, Alta. 535
Harmon Member 459

PAGE

Harp Lake 126
River 735
Harricanaw River 688
Harrison Lake 443, 454
porphyries 443
Harrogate Formation 405
Hart Ranges 25
River 374
Hartman Moraine 731, 732
Harvey Group 424
Hill mine, Que. 320
Hasaga (10a) mine, Ont. 207
Hasler Creek 524
Formation 459
Hassel Formation 583, 665
Hastings Cove Formation 240
Group 53, 125
Highland 124, 130, 131
Lowland 130, 131
(46) oil field, Sask. 534
Hat Creek, B.C. 522
Hatton gas field (137), Sask. 540
Havelock, Ont. 208
Haven Member 463
Havre St. Pierre, Que. 163
Hawke Bay Formation 241
Hay River 616, 741, 790
Formation 408, 409
Hayes River 689, 734
Group 68, 69
Hazelton, B.C. 454, 515, 522
Group 441–443, 522
Mountains 25
Hazen Plateau 585, 752
H.B. (83) mine, B.C. 506
Headland Group 325, 327
Headless Formation 402
Hearne Channel 220
Hearst, Ont. 733
Heart Hill 693
Peaks Formation 479
Heath Steele (5) mine, N.B. 324
Hebron Fiord 121
Hecate Basin 369, 448, 460
Lowland 26
Strait 22, 27
Hecla Bay Formation 561, 564–566
Hector Formation 375
Hedley, B.C. 434
Camp, B.C. 510
Mascot (66a) mine, B.C. 510
Hedman mine, Ont. 221
Heiberg Formation 575, 579, 582–584, 633, 636, 665
Helen (37) mine, Ont. 170, 174
Helena Island 572
Helikian 51, 53, 80–84, 87, 88, 93, 94, 96–99, 104, 105, 124–131, 231–232, 371–373
Helium 522
Helm Formation 460
Henderson and Conley mines (110a), Que. 223
Lake 786
orebody, Que. 195

PAGE

Henderson placer field, Y.T. ... 517
Henry Creek Formation ... 372
Kater Peninsula ... 14
Heppel Formation ... 275
Herald Member ... 392
Herb Lake ... 218
Herschel Island ... 18, 696
H.G. Young (8) mine, Ont. ... 206
Hibbs Hole Formation ... 234
Hidden Creek (15c) mine, B.C. ... 510
Hand fault ... 504
High Lake ... 199
Prairie, Alta. ... 741
Highland Bell (71) mine, B.C. ... 504
Front Moraine ... 713
Highwood River ... 444, 463, 524
Hiho mine, Ont. ... 210
Hillcrest Member ... 444
Hillier Formation ... 602
Hillsborough, N.S. ... 680
Bay ... 710
Formation ... 287
(9) mine, N.B. ... 342
Hilton (42) mine, Que. ... 178
Hinton, Alta. ... 524
History ... 36–37, 153–154, 307–309, 490
Hodder deposit, Nfld. ... 317, 318
Hodgewater Group ... 234, 235
Hogameen Range ... 26
Holden Lake ... 746
Hollandia mine, Ont. ... 200
Hollebeke Formation ... 404
Hollinger (49b) mine, Que. ... 202, 203, 220, 780
Holyrood granite ... 232–234, 239, 279, 340
horst ... 234
Homathko River ... 26
Home Bay ... 30
Hondo evaporites ... 410
Honna Formation ... 454, 460, 654
Honorat Group ... 254, 264, 265
Hope, B.C. ... 477, 518
Hopedale, Labrador ... 120
Group ... 291, 292
Horn Plateau ... 20, 741
Formation ... 616
River Formation ... 402, 407, 615
Hornby Bay ... 80
Group ... 76, 77, 80–82
Horne (18e) mine, Que. ... 190–192
Hornepayne, Ont. ... 733
Horseranch Group ... 375, 424
Horsethief Creek Group ... 375, 380, 382, 406
Horton Bluff Formation ... 287, 328
Group ... 287–289, 292, 295, 296, 329, 626, 663
Plain ... 19
River ... 19
Horwood Bay ... 266
Hotailuh batholith ... 440
Hottah Lake ... 215
Hough Lake Group ... 108, 112
Howells River ... 737
Howey (10b) mine, Ont. ... 201, 207

PAGE

Hoyle Group ... 202
(52a) mine, Ont. ... 203
Hozameen Group ... 428
Hudson Bay ... 15
Basin ... 137–140
Lowland ... 12, 15, 16, 107, 137, 139, 140, 617, 667
Mountain ... 514
Platform ... 137–145, 262, 272, 385
Region ... 15
River ... 726, 727, 736
Strait ... 15, 29, 104, 105, 698
Hudsonian Orogeny ... 49, 70, 75–77, 79, 85–97, 100–109, 115–124, 130–131, 369
Hughes Range ... 371, 380, 405
Hulcross Member ... 458
Hull Formation ... 256
Humber Arm Group ... 247, 317, 354
klippe ... 238, 242, 244, 247, 250–252, 257
Hume Formation ... 401, 406, 614, 615
Hummingbird oil field, Sask. ... 524
Hungry Hollow limestone ... 283
Hunt mine, Ont. ... 219
Hunter Creek ... 702
Hunter V and Double Standard mine (18), B.C. ... 506
Hunting Formation ... 100
Huntingdon Formation ... 477, 521
(52) mine, Que. ... 316–318
Hurley Formation ... 434, 509
Huron Ice Lobe ... 726
Huronian Supergroup ... 59, 108, 109, 112, 117, 123
Hurwitz Group ... 96, 97, 100
Hutshi Group ... 452, 461
Hydrogeological regions ... 771–790
Hydrologic cycle ... 767
Hyland Plateau ... 25
Hyper–Dubawnt Lake ... 746

Ibbet Bay Formation ... 559, 601
Ice River alkaline complex ... 394, 400, 495
Icefield Glaciation ... 702, 705
Ranges ... 26
Icon (1) mine, Que. ... 195
Idaho ... 504
Ile-à-la-Crosse Lowland ... 20
Illinoian Glaciation ... 677, 681, 705
Illinois River ... 726
Ilmenite deposits ... 183
Iltyd Range ... 388
Imina Group ... 560
Imperial Formation ... 399–401, 418, 435, 663
Vermilion Ridge No. 1 well ... 381
Imports ... 41
Index Formation ... 383, 406
Indian Cove ... 328
Head anorthosite ... 333, 335
deposit ... 185, 333, 335
Range Complex ... 231, 240, 241
Islands Group ... 266, 267
Mountain Lake ... 73, 199
Peninsula ... 219

PAGE

Indian Point Formation ... 265
River ... 703
Indicated (probable) ore ... 39
Indin Lake ... 75
Industrial mineral deposits ... 220, 336, 518
Inferred (possible) ore ... 39
Ingenika Group ... 381, 425
Inglefield Land ... 555, 564
Inhabitants River ... 680
Inklin Formation ... 440, 441
Inklinian Orogeny ... 369, 424, 428, 438–440
Innisfail, Alta. ... 522
Innuitian Orogen ... 547–590
Region ... 17, 18
Insular Fold Belt ... 369, 467, 475
Geanticline ... 369, 438, 454, 460
Mountains ... 26
Trough ... 411, 431, 432, 435, 438, 442, 443
Interior Plains ... 19–20, 25, 388, 481
Hydrogeological Region ... 781–786
Plateau ... 25, 416, 434, 454, 467, 476–480, 754, 756
Platform ... 366, 384, 385, 388–392, 406–410, 418–421, 435–438, 444, 445, 455–464, 478, 481
System ... 21, 22, 25
Interlake Group ... 142, 392
Intrapermafrost ... 790
Intrepid Inlet ... 578, 588
Inuvik, N.W.T. ... 741
Invermere, B.C. ... 518
Inverness coalfield, N.S. ... 357, 359
Iona orebody, B.C. ... 514
Ipperwash limestone ... 283
Irene Bay Formation ... 375, 406, 557
Ireton Formation ... 410
Iridium deposits ... 217
Irishtown Formation ... 238
Iron deposits ... 169–177, 330–333, 498, 515, 589
Irondequoit Member ... 270
Iron-formation, Algoma type ... 155, 170, 174–177
Superior type ... 161, 170–174
Iron Mike (37) mine, B.C. ... 515
Mountain, Michigan ... 108
Isaac Formation ... 375
Isachsen Formation ... 572, 575, 582–584, 589, 665
Peninsula ... 588
Ishbel Group ... 420
Iskut River ... 26, 416, 481
Island belt ... 212
Lake Group ... 69
Mountain (31) mine, B.C. ... 509
Islay Member ... 459
Isle Royale ... 729
Isogun River ... 741
Isortoq River ... 749, 784
Isotopic Age Map of Canada (1256A) ... 5
Iverson Range ... 390
Ives mine, Que. ... 317
Ivishak Member ... 435
Ivry deposit, Que. ... 183

Jáchymov, Czechoslovakia ... 212, 215

PAGE

Jackass Mountain Group ... 453
Jacksons Arm Formation ... 267
Jacksonville (7) mine, N.B. ... 334
Jacobsville Formation ... 235
Jade deposits ... 518
Jaeger Formation ... 579
James Bay ... 15, 16, 58, 66, 107, 137, 139
Region ... 16, 17
Jamieson (46) mine, Ont. ... 191
Jardun (69) mine, Que. ... 200
Jarvis mine, Ont. ... 214
Jasper, Alta. ... 421, 445, 470, 519, 603, 625
Jean-Marie Member ... 408, 410
Jedney (5b) gas field, B.C. ... 537
Jedway (26) mine, B.C. ... 515
Jeffrey (58) mine, Que. ... 338
Jennings area, B.C. ... 424
Jens Munk Island ... 555
Jersey (80b) mine, B.C. ... 505, 506, 514
Joarcam–Joseph Lake (66) oil field, Alta. ... 539
Joffre (94) oil field, Alta. ... 539
Joggins coalfield, N.S. ... 356, 358
moraine ... 710
Johnson alkaline complex ... 311
Canyon Formation ... 420
Joli Fou Formation ... 459, 460
Jones Creek Formation ... 605
Sound ... 19, 751
Jones–Lancaster Plateaux ... 749
Jordon Falls ... 335
Mountain ... 333
Joutel (11a) mine, Que. ... 193
Jowett Formation ... 406
Joy Bay ... 736
Juan de Fuca ridge ... 476
Strait ... 27
Jubilee, N.S. ... 329
Formation ... 382, 385
Judy Creek (19) oil field, Alta. ... 528
Julies Harbour Group ... 255
Jumping Pound (110a) gas field, Alta. ... 522, 542
Jupiter Formation ... 272, 605
Jurassic ... 296, 298, 438–445, 574, 579–582, 641–648, 663–670

K-Ar dates ... 49, 50
Kagawong Formation ... 140, 259
Kahochella Formation ... 95
Kaiashk Interlobate Moraine ... 731
Kakabeka Formation ... 118
Kakisa Formation ... 408, 410
Kaladar granite ... 131
Kaladar–Dalhousie synclinorium ... 131
Kam-Kotia (47) mine, Ont. ... 191
Kamloops, B.C. ... 516, 704
Group ... 477, 478
Kamouraska Formation ... 246
Kananaskis, Alta. ... 520
Formation ... 420
River ... 524
Kandik Basin ... 455

PAGE

Kane Basin 555, 568
Kanguk Formation 575, 583, 584, 665
  Peninsula 583
Kaniapiskau Plateau 11, 16
  Supergroup 60, 102, 104, 108, 119, 123, 125
Kaninuriak Lake 747
Kansan glaciation 677, 681
Kanuyak Formation 76
Kaolin deposits 222
Kapico iron range 175
Kapuskasing, Ont. 66, 220
  Belt 45, 52–54
  Subprovince 64
Karmutsen Group 435
Kashabowie Lake 66
Kashba Lake 746
Kaskapau Formation 463, 498
Kaskawulsh Glacier 756
  Group 402
Kaslo Group 423, 434
Katherine Group 381
Kaumajet Mountains 121, 697, 734, 735
Kay Point 696, 742
Kaybob (29) gas fields, Alta. 539
  South gas field, Alta. 537
Kaza Group 374, 375, 406
Kazan Formation 98
  Region 13, 16
  River 744, 746
  Upland 13
Kechika Group 391, 601–604
  Ranges 382, 402
  River 403
  Trough 378, 391, 394, 403, 411
Kee Scarp Formation 401, 527, 542, 614
Keele Range 24, 388, 418, 444, 455, 473
  River 373
  Trough 455
Keewatin Ice Divide 745–747
  sector 705, 706, 738
Keg River Formation 407, 527
Keld Member 464
Kelligrews Brook 334
Kellogg, Idaho 495
Kelvin Glen Group 240, 268
Kemano, B.C. 427
Kenbridge deposit, Ont. 180
Keno Hill 451, 504, 506, 515
Kenogami River 689
    Formation 143, 144
Kenora, Ont. 180, 731
Kenoran Orogeny 49, 57–76
Kent county, Ont. 348
  Peninsula 747
Kentville Formation 268, 277, 605
Kerr Addison (59f) mine, Ont. 204
Kerrowgare Formation 268
Kettle Point Formation 145, 283, 284, 478, 617
Ketza River 504
Kewagama Group 60, 62
Keweenawan Group 109, 118, 199
  Peninsula, Michigan 235

PAGE

Keymet (1) deposit, N.B. 324, 325
Kibbey Formation 420
Kicking Horse (98) mine, B.C. 495, 505
Kidd Creek (48) mine, Ont. 153, 190, 191, 209
Kiglapait intrusion 119, 121
  Mountains 697, 734, 735
Kilian Formation 84, 589
Kilmar (48) mine, Que. 219
Kilrush, Ont. 728
  sill 729
Kiltsuit River 427
Kimberley, B.C. 515
  fault 504
Kimberlite dyke 298
Kindle Group 403, 418
King Point 696
  Salmon Formation 432
    thrust 440, 442
  William Island 743, 744
Kingak shale 444
King–Bell–Johnson–Beaver mine (61d), Que. 337
Kingdon (100) mine, Ont. 200
Kingsport, N.S. 775
Kingston, Ont. 200, 220, 221, 223, 777
Kingsvale Group 453, 454
Kingsville gas field, Ont. 307, 345, 348
Kipalu Formation 106
Kipp, Alta. 695
Kippens Formation 241
Kirkfield Formation 256
  outlet 728
Kirkland Lake, Ont. 36, 57, 62, 63, 109, 154, 155, 175, 186, 201, 298
    fault 204
  Minerals (61c) mine, Ont. 204
  phase lake 728
Kirkland Lake – Larder Lake Camp, Ont. 200, 203, 206
Kishenehn Formation 470, 480
Kiskatinaw Formation 418, 537
Kispiox, B.C. 522
Kisseynew Complex 90–92, 196, 198
Kitchener Formation 371
Kitimat Range 26
Kitsilano Formation 477
Kitson Formation 559
Kittys Brook moraine 707
Klaza Glaciation 702, 703
Klinaklini River 26
Klondike, Y.T. 36, 481, 702
  district (3), Y.T. 517
  Plateau 25
  River 22
  Schist 424, 517
Kluane Glaciation 702, 705, 756
  Lake 23, 654, 702, 705, 706
  Range 26, 435, 478, 479
  River 756
Knee Lake 69
Kneehills Tuff 464
Knob Hill Formation 510
  Lake 156, 221
    Group 102

PAGE

Knob Hill–Ironsides orebody, B.C. ... 510
Knoydart Formation ... 276
Kognak Lake ... 97
Koksoak River ... 17, 154
Kongsberg, Norway ... 212
Kootenay Arc ... 406, 434, 450, 504
  Formation ... 444, 445, 455, 522
  Inlet ... 477
  King mine, B.C. ... 495
  Lakes ... 23, 375, 381, 415, 442, 506, 704
Korah phase lake ... 729
Koraluk River ... 735
Koroksoak Glaciation ... 697
Kotaneelee anticline ... 542
  Formation ... 463
Kotcho Formation ... 410
  Lake (2) gas field, B.C. ... 527
Koukdjuak River ... 19
Kowash iron range ... 175
Krist Group ... 202
Kunga Formation ... 435, 442
Kuskanax batholith ... 440
Kwataboahegan Formation ... 144
  River ... 689
Kwejinne Lake ... 77
Kyak Formation ... 399, 418
Kyanite deposits ... 222

Laberge Group ... 440, 441, 443, 452
La Biche River ... 459
  Formation ... 463
Labrador Fold Belt ... 47, 52, 53, 86, 87, 107, 121, 123, 125, 126, 297
  Geosyncline ... 60, 101, 102, 155, 156, 160, 171, 173, 198
  Group ... 238, 240
  Highland ... 10, 13 15
  Hills ... 11, 13, 15, 16
  Sea ... 29, 747
  sector ... 705, 706, 712
  Shelf ... 29, 120, 121
  Subprovince ... 101
Labuma till ... 692
La Butte Formation ... 615
Lac Croche Formation ... 274
  des Baies Member ... 263
  des Mille Lacs ... 66
  Expanse ... 218
  Jeannine (74) mine, Que. ... 39, 174
  la Biche ... 741
  la Croix ... 218
  la Martre ... 742
  le Fer ... 737
  Pin Rouge ... 185
  Raymond Formation ... 263
  Seul ... 68
    moraine ... 731, 733
  Tio (73) mine, Que. ... 105, 177, 183, 184
Lachine Formation ... 260
Lachute, Que. ... 726
Lacnor mine, Ont. ... 166
Lacorne (28) mine, Que. ... 217–219

PAGE

Ladner Group ... 443
Laforce Member ... 264, 605
Laguna (11) mine, Man. ... 208
La Have Bank ... 29
  Basin ... 29
  River ... 709
Laib Formation ... 383
Lake Albanel ... 130
  Asbestos (61e) mine, Que. ... 337
  Athabasca ... 88, 92, 94, 153, 156, 159, 208, 745, 746
  Aylmer Group ... 263
  Branch Formation ... 275
  Champlain ... 281
    Valley ... 727
  Chibougamau ... 59
  Chippewa ... 730
  Couchiching ... 728
  De Montigny ... 204
  Deschaillons varved clay ... 683
  Dubuisson ... 781
  Dufault (18g) mine, Que. ... 191
  Erie ... 686, 727
  George ... 316
  Hazen ... 753
  Hough ... 729
  Houghton ... 730
  Iroquois ... 728
  Labelle ... 222
  Manicouagan outlier ... 257
  Manitoba ... 733
  Manitouwadge ... 154
  Maumee ... 726
  Melville ... 17, 127
    Plate ... 47, 53
  Memphremagog ... 263, 281, 317
  Michikamau ... 735
  Minong ... 729, 730
  Mistassini ... 16, 59, 108, 109, 156, 159, 161, 195
  Muskoka ... 131
  Nipigon ... 16, 118, 175
  Nipissing ... 129, 138, 156, 730
  of the Woods ... 175, 201, 731
  Panache ... 113, 114, 123
  Plateau ... 16, 713
  St. Clair ... 277
  St. John ... 138, 194, 257, 260, 601, 602
    outlier ... 140, 256, 260
  St. Joseph ... 67, 155
    iron range ... 175
  Scarborough ... 684, 685
  Shore (61b) mine, Ont. ... 204
  Simcoe ... 27
    Lowland ... 728
  Stanley ... 729, 730
  Superior ... 731, 732
  Témiscouata ... 254
  Timagami ... 112, 131
  Timiskaming ... 138, 200, 257, 260, 602, 733, 734
    outlier ... 12, 140, 142, 256, 260, 262, 271
  Traverse ... 732
  Winnipeg ... 20, 67, 139, 733, 739

PAGE

Lake Winnipegosis 733, 785
La Manche lead mine, Nfld. 307
vein 329
Lamaque (34b) mine, Que. 205
Lambton county, Ont. 307, 345, 348, 351
Lancaster Formation 292
Plateau 19
Sound 19, 30, 747
Landry Formation 401, 402
Langis mine, Ont. 210
Lanigan, Sask. 694
mine, Sask. 520
Lansdowne House, Ont. 732
Lantsville, B.C. 788
Lapa Cadillac (26b) mine, Que. 205
La Poile Bay 267, 276, 279, 329
Group 267, 276
Laporte Group 104
Laprise Creek (4) gas field, B.C. 537
Laramide Orogeny 367, 368, 424, 448, 452, 461, 464–476, 481
Larch Plateau 16
River 16
Lardeau area, B.C. 625
Group 391, 406
Trough 375, 378, 382, 383, 385, 388, 391, 394, 406, 434
Larder Lake, Ont. 62, 154, 200, 201–204, 733
break 203, 204
fault 62, 63, 201
La Ronge, Sask. 155, 198
belt 157–159
Larrabee mine, B.C. 518
Last Mountain Lake 783
Laurentian Channel 29, 300, 706, 713
Highlands 17, 682, 713
Region 17
Laurentide Ice Sheet 705, 706, 711, 747, 750
Lauzon Formation 236
Laval Formation 250, 256
La Vieille Formation 264, 605
Lawn, Nfld. 335
Lea Park Formation 463
Lead deposits 156, 200, 319–330, 495–498, 504–506, 589
smelters and refineries 38
Leadville deposit, N.S. 329
Leaf River 11, 16
Leaside Till 685
Ledbetter slate 391
Leduc Formation 408, 409, 522, 528, 531
Leitch (28a) mine, Ont. 206
Leitches Creek 680
Leitchville–Shaunavon (20) oil field, Sask. 538
Leon Lake (206) oil field, Sask. 538
Lepine Formation 459
Leray Formation 256
Lesser Slave Lake 741
Les Vieilles-Forges, Que. 682
Lethbridge, Alta. 524, 695, 781
Letitia Lake Group 126
Level Mountain 25

PAGE

Level Formation 479
Lévis, Que. 238
Formation 245
Lewes River Group 432, 434, 636
Lewis thrust 468, 470
Li Fiord 579
Liard Arch 394, 468
Basin 411
Formation 436, 438
Plain 20, 25, 754
Plateau 20, 24, 367, 418, 473, 542, 565, 701, 754
River 24–26, 418, 436, 463, 478, 518, 524, 705, 741
Trough 411, 415, 418, 420, 431, 438, 444, 448
Lightning Creek 517
Lillooet, B.C. 431, 477–517, 518
Group 453
Lime Creek (16) mine, B.C. 514
Limestone deposits 342–344, 520
Lindbergh, Alta. 519
Lions Head Member 270
Lisburn Formation 418
Lithium deposits 217, 218
Little Abitibi River 688, 734
Bay (9) deposit, Nfld. 325, 327
Bear Formation 455
Lake 478
Buffalo Formation 615
Cornwallis Island 589
Current 728
Deer Pond deposit, Nfld. 327
Falls Formation 243
Fogo Islands 266
Ice Age 756
Long Lac 200, 201
Long Lac Camp, Ont. 206, 207
Pic River 730
Sachigo Lake 732
Smoky River 740
Whale River 200
Liverpool Bay 742
Livingstone Creek Formation 112
Formation 419, 520
placer field, Y.T. 517
Range 469
thrust 467
Llama Member 436
Lloydminster, Alta. 460
Embayment 378, 384, 385, 392
(71) oil field, Alta.–Sask. 539
shale 463
Lockhart River 746
Lockport Group 344, 605
Logan sills 53, 109, 118
Logan's Line 27, 228, 254, 272, 279, 281
thrust 252, 311
Logjam Creek 504
London, Ont. 681, 777
Londonderry (3) mine, N.S. 307, 332
Lone Land Formation 373, 374

PAGE

Long Island ... 710
Point ... 726
Formation ... 252, 257, 272, 278
Range Complex ... 231, 240
Mountains ... 313
Plateau ... 706
Rapids Formation ... 144, 617
Reach Formation ... 605
Longarm Formation ... 454, 455
Lookout Butte (143) gas field, Alta. ... 542
Loomis Member ... 420, 520
Loon Lake ... 118
River Formation ... 459
Lorrain Formation ... 59, 113, 114, 223
Lorraine (40) mine, Ont. ... 180
Lotbinière Formation ... 260
Lotsberg Formation ... 406
Love Cove Group ... 231, 239, 279
Lowville Formation ... 256, 601
Lucas Formation ... 277, 344, 351
Lucky Strike deposit, Nfld. ... 325
Lulu Island ... 520
Lumby Lake ... 67
Lundy glacial lake ... 726
Lunenburg county, N.S. ... 335, 336
Luscar, Alta. ... 524
Formation ... 457, 458, 524
Lushs Bight Group ... 247
Lussier syncline ... 405
Lyal Formation ... 385
Lyleton Formation ... 410
Lyndoch township, Ont. ... 217
Lynn Creek ... 704
Lake, Man. ... 90–92, 152–155, 180, 183, 196
belt ... 91
(1) mine, Man. ... 180
Lynx Formation ... 384

Mabou, N.S. ... 680
coalfield, N.S. ... 357, 359
Mabou–Antigonish Basin ... 293
MacAlpine Lake ... 745, 747
Macassa (61e) mine, Que. ... 204
Macasty Formation ... 260
MacCodrum Formation ... 240
Macdonald Platform ... 375, 380, 385, 391, 392, 394, 399, 411
Machin Point ... 195
MacKay Range ... 473
Mackenzie Arch ... 374, 375, 378, 381, 385, 388, 390, 392, 394, 399–401
Delta ... 18, 461, 478, 654, 695, 696, 742, 755
dyke swarm ... 75, 82, 85, 88, 96
Fold Belt ... 367, 390, 452, 464, 471, 473
King Island ... 579, 581–583, 751
Mountain Area ... 24
Mountains ... 20, 21, 24, 367, 373, 374, 378, 388, 390, 399, 401, 406, 418, 452, 455, 472, 474, 498, 524, 701, 702, 755
Plain ... 24, 455, 473, 478
River ... 13, 18, 20, 24, 473, 524, 614, 696, 741, 742, 785, 788

PAGE

Mackenzie Trough ... 378, 385, 390, 392, 394, 401, 402, 448, 455
Mackinac Channel, Michigan ... 730
Straits breccia ... 144
MacLean Brook Formation ... 240
orebody, Nfld. ... 325
MacLeod–Cockshutt (32) mine, Ont. ... 206
MacMillan Plateau ... 388, 402, 416, 417, 452
River ... 478
MacMullin Formation ... 240
MacNeil Formation ... 240, 249
Madoc, Ont. ... 154, 200, 221
Mafeking, Man. ... 520
Magdalen Islands ... 29, 289, 290, 295, 299, 333, 342
Magnesite deposits ... 380, 518
Magnesium (magnesia) deposits ... 219
Magnesium sulphate (epsom salt) deposits ... 519
Magnet Cove (8) mine, N.S. ... 327, 338–340
Magnetic anomalies ... 54, 64, 65, 70, 71, 154, 185, 476, 477
Magnetic Anomaly Map of Canada (1255A) ... 5
Magog Formation ... 252, 254, 257
Magpie Mountain ... 185
Magrath, Alta. ... 524
Maiden Point Formation ... 238, 248, 252
Mainland belt ... 212, 214
Maisonneuve township, Que. ... 154
Maitland, N.S. ... 300
River ... 341
Majeau Lake Formation ... 409, 410
Makhovik, Labrador ... 163
Makinson Inlet ... 701
Malagash salt mine, N.S. ... 342
Malartic, Que. ... 60–64, 733
Gold Fields (33a) mine, Que. ... 204, 205
Group ... 60, 62
Malbie Formation ... 274, 275
Malcolm River ... 435
Malden oil field, Ont. ... 346
Malignant Cove ... 710
Formation ... 258
Maligne Formation ... 403, 405
Malmo–Duhamel (85) oil field, Alta. ... 532
Mamainse Point ... 118
deposits, Ont. ... 196
Mandy mine, Sask. ... 196
Manetoe Formation ... 401, 402, 614
Manganese deposits ... 333
Manicouagan crater ... 136
Lake ... 12, 138
Manitoba Escarpment ... 20
Group ... 520, 785
Plain ... 20, 784
Manitou Island ... 728
Manitoulin Formation ... 268, 605
Island ... 27, 116, 230, 235, 243, 256, 259, 269–271, 602
Manitounuk Group ... 105, 107, 200
Manitouwadge, Ont. ... 36, 153, 173, 185, 193, 194, 217
Maniwaki, Que. ... 219
Mannville Group ... 459, 463, 539
Mansel Island ... 137, 138

PAGE

Manuels (17) mine, Nfld. ... 334, 340
River Formation ... 240
Maple Creek, Sask. ... 690
Mapleguard sediments ... 704
Maquereau Group ... 252, 254, 264
Marathon, Ont. ... 186
Marble Bay Formation ... 520
(48b) mine, B.C. ... 510
Canyon Formation ... 431
Marbridge (27) mine, Que. ... 181
Marcellus Formation ... 283
March Formation ... 249
Point Formation ... 242
Marine Sciences Branch ... 4
Maritime Disturbance ... 229, 272, 283, 284, 286, 295, 296, 313
Plain ... 28
Marker Lake batholith ... 451
Markham, Ont. ... 685
Markhamville deposit, N.S. ... 333
Marks moraine ... 731
Marl deposits ... 522
Marmora (111b) mine, Ont. ... 117
Marron Formation ... 477
Marsden, Sask. ... 695
Marshybank Member ... 463
Marston Member ... 420
Martin Formation ... 59, 93, 94, 167
Marymac Lake ... 737
Maryvale, N.S. ... 710
Mascarene Group ... 265, 276
Masik River ... 700
Massett Formation ... 477
Matagami, Que. ... 36, 153, 155, 185, 191, 192, 193, 217
Matapedia facies belt ... 244, 245, 254, 257, 261, 264
Group ... 254
region ... 263
Valley ... 711, 712
Matheson, Ont. ... 221
Matinenda (lower Mississagi) Formation ... 109, 165
Matonipi Lake ... 173
Mattagami Formation ... 145, 688, 689
Lake (9a) mine, Que. ... 193
River ... 688, 689
Mattawa, Ont. ... 185, 222
River ... 729, 730
Mattson Formation ... 418, 663, 665
Maude shale ... 442
Maunsel till ... 692
Maxwell Bay ... 751
Mayo, Y.T. ... 36, 424, 451, 703, 705, 755, 790
Lake ... 23
Landing, Y.T. ... 705
Mayook (108) mine, B.C. ... 519
McAdam Formation ... 268, 605
Lake Formation ... 277
McCann Hill chert ... 401
McCarthy shale ... 435
McConnell Formation ... 399
Glaciation ... 702
moraine ... 703, 755

PAGE

McConnell thrust ... 468, 470
M'Clintock Channel ... 19, 744
Group ... 560
Inlet ... 560, 569
M'Clure Strait ... 743
McDame Creek ... 36
Group ... 614
Intrusions ... 423
McDonald fault ... 54, 88, 93, 96, 417
McDougall Group ... 381
Pass ... 23
McDougall–Segur sandstone ... 457
McIntyre (49d) mine, Ont. ... 191, 202, 203
McKay Group ... 382, 385, 391, 418, 603
McKenzie (9f) mine, Ont. ... 207
McKim Formation ... 109
McLeod Brook Formation ... 249
River ... 498
McMarmac (9g) mine, Ont. ... 207
McMurray, Alta. ... 519
Formation ... 459, 539
McQuesten, Y.T. ... 424, 705
Meadowcliffe till ... 685
Meadowlake Escarpment ... 394, 406, 407
Meaford, Ont. ... 730
Meaford–Dundas Formation ... 140, 259, 344
Mealy Mountains ... 10, 13, 17
Measured (proven) ore ... 38
Mecatina Plateau ... 17, 708
Medicine Hat, Alta. ... 490, 521, 524
gas field (137), Alta. ... 540
River gas field (96), Alta. ... 537
Medina sandstone ... 269
Megantic Hills ... 29
mine, Que. ... 337
Meguma Group ... 240, 330
Trough ... 235, 240, 244, 245, 249, 253, 255, 262, 268, 272, 277
Meighen Island ... 18, 552, 587, 698, 699, 750, 751
Melville Hills ... 19
Island ... 18, 551, 555–559, 561, 564, 569, 572, 573, 575–583, 586–589, 616, 625, 699, 700, 738, 750
Group ... 399, 564–566
Peninsula ... 87, 100, 101, 747, 748, 750
Plain ... 17
Plateau ... 15
Melvillian Disturbance ... 411, 417, 418, 569, 572
Memramcook Formation ... 283, 287
Menihek Formation ... 102
Mercoal, Alta. ... 524
Mercury deposits ... 516
Merigomish Island ... 710
Merritt, B.C. ... 522, 704
Mersea–Leamington gas field, Ont. ... 348
Mesa Lake ... 72
Mesabi iron range, Minnesota ... 173
Mesozoic ... 145, 296–301, 428–464, 574–583, 663–670
Meta Incognita Peninsula ... 747
Metaline District, Washington ... 505
Formation ... 382
subtype ... 505

PAGE

Metallogenic associations 154–162, 306–315, 490–493, 498–504
Metchosin Formation 477
Methy Formation 406, 614, 615
Mexico 623
Mica deposits 222
Michel, B.C. 522, 524
Michelle Formation 399
Michigan 196, 199, 230, 235, 243, 269, 283
  Basin 143–145, 228, 230, 242–245, 255, 259–263, 268–272, 283, 284, 345, 348
Michikamau intrusion 119, 120, 127
  Lake 119, 120
Michipicoten, Ont. 153, 170, 174
  Island 16
Mictaw Group 247, 254, 264, 281
Midale beds 420, 534
  (39) oil field, Sask. 534
Midas Formation 415, 509
Mid-Atlantic Ridge 29
Middle Arm Point Formation 247
  River 330
    Group 253, 268
"Midway" dyke 510
Mid-Wisconsin interglacial interval 703
Miette Group 374, 375
  Range 470
  reef complex 404
Mildred Member 408
Milford, N.S. 680
  Group 406, 415, 434, 625
Milk River 691
    Formation 463, 464, 524, 766
Milkish Head granite 253
Millbridge, Ont. 185
Mill Creek flora 457
    Formation 457, 458, 460
Milligan Creek (13c) oil field, B.C. 537
  Lake mine, Ont. 166, 781
Milne Inlet 177
Miminiska Lakes 732
Minas Basin 28, 287–293, 295, 296, 298, 709
  Lowland 28
Mindamar (Stirling) mine (30), N.S. 325
Mineral Deposits of Canada, Map 1252A 5
  exports 36, 40, 41
  King mine (104), B.C. 518
  production 32, 40–42, 152, 306–309
  provinces 154–156, 312–315, 491
  reserves 38–40
  rights 3
Mingan, Que. 185
  Formation 250, 257
  Island 28, 228, 229, 242, 250, 257, 601, 776
Mingo River Formation 555, 603
Ming's Bight 330
Miniss Group 67
Minnedosa, Man. 694
  River 693, 694
Minnes Formation 445, 455
Minnesota 108, 119
  River 732

PAGE

Minto Arch 84, 549
  coalfield, N.B. 307, 356, 361
  Group 90–92
  Inlet Formation 84, 589
  Inlier 45, 53
  Subprovince 77, 84
Miramichi Bay 711
  Geanticline 244, 247, 257, 261, 262, 264, 265, 272, 275, 276
Misinchinka schist 425
Mispek Group 292, 296
Missi Group 90–92, 196
Missinaibi beds 688, 689
  River 688, 689
Mission Canyon Formation 420
Mississagi Formation 112, 195
Mississippi River 726, 732
  Valley 595, 623, 728
Mississippian (Lower Carboniferous) 284–291, 411–415, 418–420, 568–569, 623–632
Missouri Coteau 20, 21, 690
  River 691
Mistassini Group 59, 60, 108, 112, 195
  Hills 16
  Homocline 45, 47, 52
  Lake 173
  Subprovince 59, 87
Mistaya Formation 384, 385
Mitchell River 518
Mitsue (17) oil field, Alta. 527
Moak Lake 152
    orebody, Man. 181
Moberly Member 408
Mohawk River 726, 727
Mokka Fiord 584, 586
Molly Gibson (89) mine, B.C. 506
Molybdenum deposits 217, 219, 335, 514
Monach Formation 456
Monarch (98) mine, B.C. 495, 505
Monashee Group 416, 417
  Mountains 373, 416, 425, 461, 476
  Valley 754
Moncton, N.B. 309, 712
  Basin 284, 286–288, 290, 291, 293–295
Moneta Porcupine (49e) mine, Ont. 203
Monkman Formation 391
Monster Formation 461
  syncline 435, 452
Mont Louis, Que. 712
Montagnais Group 102
Montana 385, 394, 407, 408, 410, 461, 504
Montania 375, 380
Monteith Formation 445, 455
Monteregian Hills 27, 314, 779
  Intrusions 145, 220, 278, 296, 298, 311, 335
Montgomery Lake Group 97
Montney Formation 436, 537
Montreal, Que. 256, 260, 278, 601, 726, 766, 780
  Formation 256
Moorehouse Member 278
Moose Channel 478
    Formation 461, 478

PAGE

Moose deposit, N.W.T. ... 217
Factory, Ont. ... 688
Jaw, Sask. ... 781
Mountain Member ... 444
(84) mine, Ont. ... 175
River Basin ... 137, 139, 140, 145
Formation ... 144
Moosebar Formation ... 458
Moosonee, Ont. ... 66
Morden, Man. ... 521
Member ... 464
Moresby Island ... 515
Moretons Harbour ... 316
Morien Group ... 294
Morin anorthosite ... 125, 127
Morrison River Formation ... 234, 235, 239, 240
Morro Member ... 404
Moss mine, Que. ... 219
Mother Lode (73a) mine ... 510
Mould Bay ... 588, 750
Formation ... 552, 579, 581, 582
Moulton Hill granite ... 252
Moulton Hill–Aldermac mine (54), Que. ... 320
Mount Albert ... 316, 337
intrusion ... 251, 252, 336
Alexander Group ... 264, 275
Allan ... 444
April Formation ... 391
Assiniboine ... 385
Bayley Formation ... 571, 572
Brome ... 314
Brussilof ... 518
Burgess ... 400
Cap Formation ... 378
Churchill ... 25
Clark Formation ... 378
Copeland deposit, B.C. ... 506
Disraeli Group ... 560
Edziza ... 23, 25, 479–481
Eisenhower ... 495
Fitton granite ... 399
Forster Formation ... 405
Garibaldi ... 460
Hawk Formation ... 403, 410
Head Formation ... 419–421, 520
Hooker ... 19
Ida Group ... 417, 450
Keele ... 25
Kindle Formation ... 388, 605
Logan ... 26
Luciana ... 26
Lytton batholith ... 454
McDame ... 518
Megantic ... 314
Murray diabase sills ... 415
Musgrave Formation ... 237, 238, 279
Nelson Formation ... 518
Reed ... 39, 174
Roberts Formation ... 442
Robson ... 25, 472, 603
Royal alkaline complex ... 311
St. Elias ... 26

PAGE

Mount St. Gregory ... 317
Sedgewick granite ... 399
Shefford ... 314
Simon Formation ... 242
Steele ... 26
Tsar ... 451
Waddington ... 26
Wilson Formation ... 391
Wissick Group ... 263, 605
Wright ... 39, 128, 130, 173
Yamaska ... 314
Mountain Park, Alta. ... 524, 537
Formation ... 457, 458
Mowitch Creek ... 519
Formation ... 421
Moydart Formation ... 268
Moyie fault ... 375, 380, 394, 405, 450
gabbro sills ... 371, 504
Mugford Group ... 121
Munising Formation ... 242
Munro mine, Ont. ... 191, 221
Mural Formation ... 380
Murdoch Formation ... 102
Murdochville, Que. ... 317
Murky Formation ... 96
Murphy Creek Formation ... 254
Murray Bay ... 602
fault ... 115, 116
granite ... 116, 117
Island Formation ... 144
Musgravetown Group ... 234, 235
Mush Lake Group ... 435
Muskeg Formation ... 407, 527
Muskiki Member ... 463
Muskox disturbance ... 371
intrusion ... 53, 54, 77, 81, 82
Muskwa Formation ... 407, 408
Ranges ... 25
River ... 455
Musquodoboit River ... 298
Mutz Member ... 444
Mysterious Creek Formation ... 443
Mystery Lake orebody, Man. ... 181
Mystic Formation ... 253

Nadlo Point Formation ... 555, 603
Nahanni Formation ... 402, 403, 406
Plateau ... 399
Nahlin fault ... 423
Nahoni Range ... 24, 473
Nahwitti Lowland ... 27
Nain anorthosite ... 121
Province ... 47, 119–121
Naiset Formation ... 385
Nakina, Ont. ... 175
Moraines ... 572, 733
Nanaimo coalfield, B.C. ... 522
Group ... 460, 477, 520, 523
Lowland ... 786, 788
Nansen drift ... 703
Formation ... 571, 572, 574
Glaciation ... 703

PAGE

Nansen Sound 18, 701
Nappan (1) brine field, N.S. 342
Narrows Formation 240, 248
Nasina Group 424
Naskaupi Fold Belt 47, 53
Nass Basin 25
Nassian Orogeny 369, 424, 438, 442
Nastapoka Group 105–107
  Islands 174
  Sound 106
Nation River Formation 401
National Asbestos mines (62a), Que. 337
Natkusiak Formation 82, 84, 589
Natlins Cove Formation 267
Natural gas fields 37, 345–355, 524–544, 589
    storage reservoirs 349
    reserves 39, 345–353, 525
Navy Board Inlet 555, 747
  Island 248
Neagha Formation 270
Nebraskan glaciation 677
Nechako Plain 754, 756
  River 756
  Trough 431–434, 438, 441, 442, 452
Needle Mountain (72) mine, Que. 316–318, 335
Neepawa, Man. 519
Negus (9b) mine, N.W.T. 207
Nelson batholith 450, 461
  Front 47, 52–54
  Head 700
  Inlier 551
  Intrusions 514
  River 137, 733, 734
Nelway Formation 382, 383, 391
Neohelikian 51, 53
Nepean Formation 242, 344
Nepheline syenite deposits 222
Neroukpuk Formation 373, 399, 418, 435, 444
Nettilling Lake 749
Nettogami River 688
Neutral Hills 20
Neuville Formation 256
Nevada 603
Nevis gas field, Alta. 522
New Bay 266
  Brunswick Geanticline 236
    Highlands 28, 711
    Platform 283, 284, 290, 291, 295
  Calumet (41) mine, Ont. 199, 200, 209
  Canaan Formation 268
  Glasgow, N.S. 336, 346, 359
  Hosco (10) mine, Que. 193
  Quebec Crater 136
  Richmond township, Que. 316
  Ross, N.S. 334, 335
  World Island 255, 265, 316
  York State 124, 127, 242, 252, 277
Newcastle, N.B. 711, 712
  Creek 307
Newfoundland Basin 29
  Central Lowland 28
  Coastal Lowland 28

PAGE

Newfoundland Highlands 28, 707, 708
  Platform 283
Niagara Escarpment 27, 776, 778
  Falls 271, 686
  Peninsula 268, 270, 277, 307
  River 27, 345
Nicholson Peninsula 696
  property, Sask. 167
Nickel deposits 177–182, 316–317, 516
  Plate (66b) mine, B.C. 510
  refineries 37
Nicola Group 423, 434, 452, 454
Nicolet asbestos mines (59), Que. 337
  River Formation 260, 354
Nictaux–Torbrook area, N.S. 268
  deposits (9) 330
Nig Creek (7) gas field, B.C. 537
Nigadoo (2) mine, N.B. 315, 325
Nikanassin Formation 445, 455, 539
Nikolai greenstone 435
Nimpkish (41) mine, B.C. 515
Nine Mile Hill Formation 267
Niobium (Columbium) deposits 220, 335
Nipigon, Ont. 218
  Lake 732, 733
  Moraine 731
  Plain 13, 16
  Plate 45, 53, 206
Nipisi oil field (15), Alta. 527
Nipissing diabase 109, 115–117, 210, 212
  Great Lakes 730, 731
Nisku Formation 410, 522, 533, 544
Nitrogen deposits 522
Nizi Formation 415, 416
Nizina limestone 435
Noel Formation 434, 509
Nokomis Group 90
Nomad Member 463
Nonacho Belt 93
  Group 94
  Lake 92
Nonda Formation 391, 399
Nooksack Group 443
Nor-Acme (9) mine, Man. 208
Noranda, Que. 36, 60, 63, 109, 153–155, 191, 217, 220, 223, 733, 734
Norbestos, Que. 337
Nordegg, Alta. 524
  Member 444, 445, 537
  River 537
Nordic Belt 166
  Formation 109
  mine, Ont. 166, 781
Norfolk county 307
  gas field, Ont. 351
Normandie (60b) mine, Que. 337
Norman Range 527
  Wells, N.W.T. 490
    oil field (1), N.W.T. 527, 542
Normandville (13a) oil field, Alta. 533
Normans Cove, Nfld. 708
Normetal (12) mine, Que. 190, 191, 223

PAGE

North American Basin ... 29
Arm Mountain pluton ... 336
Bay, Ont. ... 131, 222, 223, 730
sill ... 729
Branch conglomerate ... 444, 455
Cayuga township, Ont. ... 342
Coldstream (25) mine, Ont. ... 196
Dakota ... 391, 394, 408, 410, 420, 479
Mountain ... 28, 709
Basalt ... 298
Saskatchewan River ... 435, 436, 457, 479, 740
Spirit Lake ... 57, 155
Star (6b) mine, Man. ... 198
(105a) mine, B.C. ... 504
Thompson Valley ... 788
Northern Ellesmere Fold Belt ... 551, 568, 589
Empire (29) mine, Ont. ... 206
Hydrogeological Region ... 788–790
Plateau and Mountain area ... 23, 25
Yukon Fold Complex ... 367, 464, 473
Northport, N.S. ... 710
Northumberland Strait ... 708–711
Northwest Arm Formation ... 248
Norwegian Bay ... 18, 751
Notikewin Member ... 459
Notre Dame Bay ... 238, 247, 255, 258, 262, 265–267, 281, 298, 307, 327
Mountains ... 29, 306, 712, 713
Trough ... 230, 235–239, 243–245, 247, 250–253, 255, 257, 258, 261, 262, 265, 267, 268, 272, 276, 277, 301
Nottingham Island ... 15
(47) oil field, Sask. ... 534
Nova Scotia Highlands ... 28
Platform ... 289, 291
Uplands ... 709
Nowash Lake belt ... 713
Number One (90c) mine, B.C. ... 506
Nyarling Formation ... 495

Oak Bay Formation ... 251, 265
Hill Group ... 236, 245
Ridges interlobate moraine ... 681
Oblique Creek Formation ... 424
Obonga Lake ... 183
O'Brien (25b) mine, Que. ... 204, 217
Ogama-Rockland (19) mine, Man. ... 207
Ogilvie, Y.T. ... 373, 424
Mountains ... 21, 23, 25, 374, 388, 399, 400, 418, 435, 451, 452, 461, 755
River ... 24
Ogoki Lake ... 732
River ... 732
Ohio ... 242
Oil fields ... 36, 345–355, 529–544, 589
reserves ... 39, 345–355, 525
Springs oil field, Ont. ... 307, 345, 351, 353
Ojibway (119b) mine, Ont. ... 342
Oka alkaline complex ... 122, 298
(49) mine, Que. ... 156, 220, 311, 335
Okak Bay ... 735
Okanagan Range ... 26, 702

PAGE

Okanagan River ... 754, 755
Valley ... 434, 754, 788
Oka-sur-le-Lac, Que. ... 681
Okotoks (111) gas field, Alta. ... 523, 534
Okse Bay Formation ... 560, 561, 564, 566
Old Crow Basin ... 701
Batholith ... 411
Plain ... 23–25
Range ... 25, 367, 399, 417
River ... 23, 24
Dave ultramafic dykes ... 417
Indian mine, Que. ... 219
Wives Lake ... 786
Oldman Formation ... 463, 464, 524
River ... 498, 524, 692, 695, 738
Olds (106b) gas field, Alta. ... 534
Olympia Interglaciation ... 703–705
Olympic Peninsula ... 477
Omineca, B.C. ... 517
Crystalline Belt ... 367, 368, 423, 424, 454, 460
Geanticline ... 411, 415, 423–425, 431, 434, 438, 440–444, 448, 450, 454, 463, 474, 515
Intrusions ... 454, 460
Mountains ... 25, 372, 381, 411, 415, 424, 434, 442, 454, 461
Onakawana lignite, Ont. ... 688, 689
Onaping Formation ... 117
One Tree Formation ... 443, 454
Oneida township, Ont. ... 342
Onondaga Formation ... 144, 277, 278
Onoway, Alta. ... 521
Ontaratue River ... 19
Onwater Formation ... 117
Ootsa Lake ... 454, 460
Group ... 460, 478, 515
Ooyahgah Formation ... 554
Opabin Member ... 463
Opasatika River ... 688, 689
Opemiska (6) mine, Que. ... 194, 195
Orangeville, Ont. ... 777
Orchan (9b) mine, Que. ... 193
Ordovician ... 138–140, 243–261, 385–392, 555–560, 601–613
Oregon ... 625
Orford, Que. ... 317
Oriental deposit, Nfld. ... 325
Orillia phase lake ... 728
Oriskany Formation ... 277, 344, 616
Ormiston, Sask. ... 519
Osborne Lake ... 198
Osler Formation ... 118
Osmium deposits ... 217
Ostracod Zone ... 459
O'Sullivan Member ... 459
Otish Mountains ... 16
Group ... 59, 60
Otoskwin River ... 731
Ottawa, Ont. ... 124, 130, 221, 249, 601, 777
Embayment ... 228, 235, 242, 250, 256, 260, 345, 351, 354
Islands ... 15

PAGE

Ottawa River 17, 27, 129, 130, 209, 776, 777
  Valley 728, 729
Ottawa–Quebec Lowland 228
Otter Formation 420
  Park Member 407
  Rapids 688
Ottertail Formation 385, 400
Otto Fiord 578
    Formation 571, 572, 586
Ouareau Formation 256
Outer Mountain area 26
Outpost Islands 220
Outram, Sask. 694
  Formation 391, 603
Oweegee Mountains 443
Owen Creek dolomite 391
Owen Sound, Ont. 730
Owl Capes Formation 264
Oxford Formation 249
  Group 69
  Lake 69

Pachi Group 105
Pacific Coastal Plain 367, 369, 479
  Continental Shelf 27, 367, 369, 443, 454–455, 460, 467, 476, 479
    Slope 367
  Range 26
Paddle River (34) gas field, Alta. 537
Paddy Member 459
Pakowki Formation 463, 464
Paleobotany 663–670
Paleohelikian 51, 53
Paleomagnetic surveys 81, 298
Paleozoic 138–145, 375–423, 552–557, 593–692
  outliers 138
Palisades Disturbance 296
Palladium deposits 215, 217
Palliser Formation 403, 404, 520
Palo, Sask. 519
Pamelia Formation 256
Pamour (53d) mine, Ont. 203
Panarctic Oils Ltd. 3
Pandora (25c) mine, Que. 205
Panel mine, Ont. 166
Panther River gas field, Alta. 542
Papaskwasati Formation 59
Paper shale 444
Pardee township, Ont. 180
Pardonet Formation 436, 438, 635
Paris, Ont. 342
Parkman (28) oil field, Sask. 534
Parrsboro, N.S. 710
Parry Bay Formation 76
  Channel 549, 557, 558, 562
  Islands 17, 589
    Fold Belt 551, 552, 566–568
  Plateau 18
  Sound 222
Parsnip River 756
Parson (99) mine, B.C. 518
Parson's Pond, Nfld. 352

PAGE

Pasayten Group 454
Paskapoo Formation 478, 479, 524, 783
Passage Beds 444, 445
Passamaquoddy Bay 265
Pater Formation 109
Patience Lake (12) mine, Sask. 520
Pavilion Group 431
Payette phase lake 728, 729
Paymaster (50h) mine, Ont. 203
Peace River 20, 25, 26, 436, 455, 456, 458, 463, 498, 524, 625, 635, 649, 654, 690, 691, 741, 742, 754, 756
  Arch 378, 384, 385, 391, 394, 399, 403, 408, 411, 418, 448, 456, 459, 468, 528, 537
  Basin 418, 654
  Embayment 411, 420
  Foothills 445, 463
  Formation 459, 460, 539, 540
  Plains 436, 460
  Valley 754
Peacock Hills 13
Pearl Lake porphyry 191, 203
Pearson Formation 95
Peasley Pond Formation 263
Pecors Formation 112
Peechee Member 403, 409
  reef front 405
Peejay (15c) oil field, B.C. 537
Peel Basin 378, 385, 388, 392, 394, 401, 411, 448
  Plain 20
  Plateau 20, 21, 455
  River 20, 24, 418, 435, 461, 695, 702, 742, 755
  Sound 549, 744
    Formation 550, 561
Peg Tantalum property, N.W.T. 217
Pekisko Formation 418–420, 535
Pelican Formation 460
Pelly Gneiss 424
  Mountains 25, 378, 382, 389, 391, 399, 402, 411, 416, 451, 755
  River 478, 518
Pembina Escarpment 766
  Member 464
  (36b) oil field, Alta. 463, 540
Pembroke Formation 605
Pendant d'Oreille (152) gas field, Alta. 540
Pend-d'Oreille River 490
Penetang phase lake 728
Peninsula Hill Formation 454
Pennington dyke 337
Pennsylvanian (Upper Carboniferous) .. 284–286, 291, 294, 411, 416–421, 568–572, 622–632, 663–670
Penny Highlands 697, 747
  Ice Cap 14, 747, 748
Penokean Fold Belt 47, 108, 125
  Hills 13, 16
  Orogeny 108, 109
Pense Formation 460
Percé, Que. 328
Perdrix Formation 403

PAGE

Perlite deposits ... 518
Permafrost ... 788
Permian ... 285, 294, 295, 411, 416–418, 568, 574, 622–632, 663–670
Perron (36b) mine, Que. ... 205, 206
Perry Formation ... 283
Petawawa River ... 728, 729
Peter Pond Lake ... 20
Pethei Formation ... 95
Petit Jardin Formation ... 242
Petitcodiac Group ... 294
Petitot (1) gas field, B.C. ... 527
Petroleum—*see* Oil, Natural Gas
Petrolia oil field, Ont. ... 307, 345, 351, 353
Phoenix (74) mine, B.C. ... 510
Phosphate deposits ... 521
Phroso Siltstone Member ... 436
Physiographic Regions of Canada, Map 1254A ... 6
  Subdivisions ... 4, 9–30
Pickle Crow (3) mine, Ont. ... 207
  Lake ... 201, 207
Picton county, Ont. ... 346
Pictou, N.S. ... 709
  coalfield, N.S. ... 356, 357, 359, 361
  Group ... 287, 293–295, 359, 626, 663
Pierreville, Que. ... 682, 683
Pigeon Creek sandstone ... 444
Pika Formation ... 384, 385
Pike Arm Formation ... 605
Pilley's Island (11) mine, Nfld. ... 325, 327
Pincher Creek (142) gas field, Alta. ... 522, 542, 543
Pinchi fault ... 416, 423, 452, 476, 516
  Geanticline ... 416, 423, 431, 434, 438, 440–442, 452
  Lake (23) mine, B.C. ... 516
Pine Creek (32a) gas field, Alta. ... 534
  Pass ... 403
  Point ... 504
  Point (10) mine, N.W.T. ... 495, 505
    Formation ... 405, 407, 614, 615
  River ... 455, 456
Pingoes ... 18
Pioneer (36) mine, B.C. ... 509
  Formation ... 434
Pipe Lake orebody, Man. ... 181
Pipestone Lake ... 185
  River ... 745, 746
Pirate Cove Formation ... 276, 283
Pitz Formation ... 99
Placentia, Nfld. ... 330
  Bay ... 329, 330, 706, 708
Plamondin Hill, Que. ... 733
Plateau basalts ... 23, 479
Platinum deposits ... 215, 217
Pleasant Mountain ... 335
Pleistocene ... 671–758
Plum Point non-glacial interval ... 686, 687
Pocaterra Creek Member ... 444
Point Clark ... 728
  Comfort ... 222
  Edward Formation ... 295
  Wilkins Member ... 520
Pointe-aux-Trembles Formation ... 263, 605

PAGE

Pointed Mountain anticline ... 542
Poirier (11b) mine, Que. ... 191, 192
Poker Chip shale ... 445
Polaris ultramafic body ... 423
Police Post Formation ... 555
Pond Inlet ... 30, 582, 589
  Island ... 316
Pont Rouge Formation ... 256
Pontgravé River Formation ... 260
Pontiac Group ... 60–62
Pools Cove Formation ... 283
Population ... 35, 41
Porcupine area, Ont. ... 63
  Camp, Ont. ... 200, 202, 203, 214, 220
  Hills ... 20, 739, 754
    Formation ... 479
  Mountain area ... 23
  Plateau ... 21, 23, 25, 367, 373, 374, 385, 418, 452, 473, 755
  Platform ... 385, 388, 394, 400
  Range ... 24
  River ... 418, 701, 755
  stock ... 450
Port Arthur, Ont. ... 66, 118, 180, 212
  Hills ... 16
  Homocline ... 47, 108
  Subprovince ... 118
  au Port Peninsula ... 28, 241, 242, 250, 257, 272, 278, 282, 354
  aux Basques ... 279, 708
  Harrison ... 736
  Hood coalfield, N.S. ... 356
    Formation ... 293
  Huron, Mich. ... 730
    moraine ... 726
    outlet ... 728, 731
  Nelson Formation ... 605
  Radium, N.W.T. ... 78, 168, 208, 215
  Talbot, Ont. ... 686
    non-glacial interval ... 686, 687
Portage Chute Formation ... 138
Portage orebody, Que. ... 195
Porter Road Formation ... 240
Portland Canal ... 27
Portneuf, Que. ... 199, 200
Post-Algonquin lakes ... 728–730
  Duluth lakes ... 729
  Iroquois lakes ... 713, 727
  Minong lake phases ... 730
  Nipissing lakes ... 730
Potash deposits ... 37, 519
Potential ore ... 39
Potsdam Formation ... 242, 346, 354
Pouce Coupe (6b) gas field, Alta. ... 540
Povungnituk, Que. ... 58, 133
  Group ... 104, 105
  Hills ... 16
  Ranges ... 16
  River ... 736
Powell River, B.C. ... 517
Power Glen Formation ... 605
Prairie Formation ... 407, 408, 519

PAGE

Preble Formation ... 96
Preglacial channels ... 679
Preissac (24b) mine, Que. ... 217, 219
Preissac–Lacorne batholith ... 219
district, Que. ... 217, 219
Prelate, Sask. ... 695
Ferry interval ... 695
Pre-Quadra sediments ... 704
Presqu'ile Formation ... 407, 495, 614
Preston (50i) mine, Ont. ... 203
Préval, Que. ... 712
Pride of Emory (64) mine, B.C. ... 516
Prince Albert Sound ... 700
Alfred Bay Formation ... 559
Charles Island ... 557
Edward Island ... 29, 295, 296, 299, 300, 665
George ... 434, 442, 479
of Wales Island ... 19, 549, 550, 558, 561, 616, 743, 744
Patrick Island ... 18, 551, 552, 561, 566, 575, 578, 579, 581–583, 587, 588, 616, 699, 750
Uplift ... 548, 551, 552, 569, 579, 584, 587, 588
Regent Inlet ... 19
Princess (131) gas field, Alta. ... 539
Princeton, B.C. ... 478, 521, 522
Group ... 477, 478, 521
Prong's Creek Formation ... 400
Pronto (72) mine, Ont. ... 165, 166
Prophet Formation ... 418
River ... 455
Proserpine felsite dykes ... 415
Proterozoic ... 51, 52, 59, 64–66, 74–134, 159–162, 369–375, 552
Protocontinents ... 156
Provost (79) gas field, Alta. ... 539
Pueblo (9) mine, B.C. ... 510
Pugwash (2) mine, N.S. ... 342
Pullen Island ... 696
Purcell anticlinorium ... 450, 504
Arch ... 375, 378, 380, 382, 384, 385, 388, 391, 394, 405, 438
Geanticline ... 374
Mountains ... 371, 373, 375, 380, 382, 391, 405, 415, 470, 518
System ... 371, 380
Trench ... 23, 704
Purdy mine, Ont. ... 222
Puskwaskau Formation ... 463
Putman Highland ... 19, 557
Pyrophyllite deposits ... 340

Quadeville, Ont. ... 217
Quadra non-glacial interval ... 755
sediments ... 703, 704
Qualicum, B.C. ... 788
Qu'Appelle River ... 20, 691, 740, 786
Quaternary ... 300, 301, 481, 587–588, 671–674, 676–764
Quatsino Formation ... 435
Sound ... 454
"Quartz-eye" granite ... 92
Quebec Basin ... 145, 228, 230, 235, 242, 245, 249, 250, 256, 260, 278, 298, 345, 351, 354

PAGE

Quebec City ... 27, 209, 236, 237, 245, 246, 281, 713
Formation ... 254
Geanticline ... 244, 245, 252, 254, 257, 263, 264, 272, 273
Group ... 237, 238, 242, 246, 254, 263
Lithium (29) mine ... 219
Queen Charlotte fault ... 369
Islands ... 26, 27, 369, 435, 442, 448, 454, 455, 460, 467, 476, 477, 479, 522, 654, 665, 754, 786
Lowland ... 27
Ranges ... 27
Sound ... 27
Queen Elizabeth Islands ... 18, 585, 665, 698, 750
Glacier Complex ... 706
Maud Gulf ... 100, 744, 745
Queenston Formation ... 140, 259, 260, 344, 682
Quemont (18j) mine, Que. ... 223
Quesnel, B.C. ... 442, 454, 521
Trough ... 431, 432, 434, 438, 442, 450, 452
Quetico Belt ... 45
Subprovince ... 57, 66
Quilchena, B.C. ... 521
Quirke belt ... 166
Lake Group ... 108, 112
syncline ... 165, 166
(74c) mine, Ont. ... 166
Quoddy Formation ... 605
Quyon, Que. ... 219

Raanes Peninsula ... 571, 578
Rabbit Lake gas field, N.W.T. ... 527
Mountain ... 214
Point Formation ... 552
Rackla River ... 435
Racklan Orogeny ... 367, 371, 373, 380
Radium deposits ... 163
Rae River ... 14, 81
Ragged Range ... 453
Raglan Range ... 572
Rainbow Hollow, B.C. ... 519
Member ... 407, 527
(2) oil field, Alta. ... 527
South (3) oil field, Alta. ... 527
Rainy Hollow area, Y.T. ... 478
Lake ... 66, 67, 731
Ramah Group ... 121
Rambler mine (6), Nfld. ... 325, 327
Ramparts Formation ... 401
Ramsay Lake Formation ... 112, 115
Rancheria, B.C. ... 504
Random Formation ... 234, 235, 239
Ranger Canyon chert ... 420
Rankin Inlet ... 152, 155, 178, 182
Rapdan (22) oil field, Sask. ... 538
Rapid River ... 478
Rapitan Group ... 373, 374, 498
Rat River ... 435
Ratcliffe Brook Formation ... 234
Ravenscrag Formation ... 464, 478, 479, 521, 524, 783
Rayrock (5) mine, N.W.T. ... 168

PAGE

Read Bay ... 751
Group ... 557–561, 563, 606
Red Deer, Alta. ... 691
Earth (14) oil field, Alta. ... 527
Indian Lake ... 267
Lake ... 175, 200, 201, 207
Camp (8-11) Ont. ... 206, 207
Mountain (78b) mine, B.C. ... 514
River ... 732, 739, 740, 766, 781, 783
Formation ... 139, 391, 392, 520, 524, 604, 605
Valley ... 784
Rose Formation ... 454
(20) mine, B.C. ... 515
Sucker Lake ... 220
Redknife Formation ... 408, 410
Redmond Formation ... 102
Redstone deposit, N.W.T. ... 498
River ... 374, 381
Redwall fault ... 451
Redwater (51) oil and gas field, Alta. ... 522, 528
Reed–Belanger (Chromeraine) deposit, Que. ... 336
Reeves mine, Que. ... 335
Reeves–MacDonald (79) mine, B.C. ... 506
Regina, Sask. ... 520, 781
Reid end moraine ... 702, 703
Reindeer Depot, N.W.T. ... 696
Formation ... 478
Lake ... 746
Relay Mountain Group ... 443, 453
Remote sensing devices ... 154
Rencontre Formation ... 267
Lake ... 335
Renfrew, Ont. ... 219
Rens Fiord ... 585
Complex ... 552, 560
Rensselaer Bay Formation ... 555
Resolute Bay ... 751
Restigouche River ... 29
Rexspar deposit, B.C. ... 518
Reynales Formation ... 270
Reynolds Point Formation ... 84
Rhodium deposits ... 217
Ribstone Creek Member ... 783
Rice Lake, Man. ... 67, 68, 201
Group ... 67
Richardson Islands ... 84
mine, Ont. ... 154
Mountains ... 21, 23, 366, 367, 373, 378, 388, 399, 418, 435, 444, 455, 461, 473, 474, 654, 702
River ... 14, 746
Richibucto, N.B. ... 712
Richmond Gulf ... 105, 107, 200, 736
Group ... 105
Hills ... 15, 16
Riding Mountain ... 20, 693
Formation ... 464, 783, 784
Rierdon Formation ... 445
Rigaud Mountain ... 129
Rimbey gas field, Alta. ... 522
Rimbey–Meadowbrook reef chain ... 409, 410, 531–533
Ringnes Islands ... 585, 588

PAGE

River Hebert coalfield, N.S. ... 356
John ... 294, 710
Group ... 276
Jordon deposit, B.C. ... 506
Valley property, Ont. ... 222, 223
Riversdale Group ... 292, 293, 356, 626, 663
Rivière du Loup, Que. ... 712, 713
Road River ... 601, 605, 614
Formation ... 388, 399, 418
Roaring River ... 693
Roberts Arm Group ... 258, 319, 327
Robinsons Head Drift ... 707
Robitaille Formation ... 263, 605
Rocher Déboulé Mountain ... 515
Rochester Formation ... 270, 605
Rock Candy (75) mine, B.C. ... 518
Creek Member ... 444, 445
Rockcliffe Formation ... 250, 256
Rockland Formation ... 256, 602
Rockport Quarry ... 283
Rockton oil field, Ont. ... 346
Rocky Brook–Millstream fault ... 251, 325
Mountain Area ... 24
Exogeosyncline ... 448, 455, 461, 467, 478
Group ... 418, 420, 522
Thrust Belt ... 367, 464–472
Trench ... 23, 368, 375, 380, 382, 391, 405, 425, 434, 451, 461, 470, 472, 479, 495, 519, 625, 633, 635, 754, 756, 788
Mountains ... 20, 21, 25, 26, 371, 372, 375, 378–380, 382–385, 391, 399, 400, 402–405, 418–421, 435–438, 444, 445, 450, 451, 455–472, 480, 491, 495, 701, 756
Rodney oil field, Ont. ... 351
Rogers City Formation ... 144
Rolette township, Que. ... 317
Rolling (Roaring) River ... 693
Romaine Formation ... 250, 354
Rome, N.Y. ... 727
Romney oil field, Ont. ... 346
Ronde Member ... 403
Ronning Group ... 388
Root Lake ... 218
Rope Cove Canyon ... 316
Rosaire Formation ... 245
Rosalind, Alta. ... 521
Roseray sandstone ... 460, 538
Ross Brook Formation ... 268, 605
Creek Formation ... 420
Lake ... 220
River ... 504
Rossland, B.C. ... 461, 510
Group ... 423, 442, 450
Trough ... 431, 434, 438, 441, 442
Rothermere orebody, Nfld. ... 325
Rottenstone Lake ... 198
Rougemont alkaline complex ... 311
Rove Formation ... 118, 213, 214
Ruby Range stock ... 450
Ruddock Creek, B.C. ... 506
Rundle Group ... 419, 420, 542
Russell Formation ... 260

PAGE

Ruth Formation 102
Ruthenium deposits 217
Ryans Bay 735

Sabine Bay Formation 573, 574, 665
  Peninsula 574, 584, 586, 750
Sable Bank 29
  Island 29, 299
Sackville, N.B. 772
Saddle Hills 481, 537
Saddlerochit Formation 418, 435
Saginaw Basin 726
Saglek Fiord 121
  Glaciation 697, 734
Saguenay River 602, 713
Sail Harbour Group 560
St. Alban Formation 273, 274
  Antonin moraine 713
  Bruno alkaline complex 311
  Casimar Formation 256
  Charles deposit, Que. 185
  Clare Group 474
  David's Channel 686
  Edmund Formation 269, 271, 605
  Elias Fold Belt 369, 474, 475
    Mountains 26, 402, 416, 417, 423, 435, 455, 467, 474, 475, 478, 479, 481, 654, 701, 702, 705, 754, 755, 786
    Trough 411, 431, 435, 448, 455
  Eugene (109) mine, B.C. 504
  Francis Group 263, 273, 275
  Genevieve Limestone 623
  George, N.B. 335
    Group 242, 250, 329, 354
  George's Basin 287, 289, 290, 292, 293
    Bay 333, 336, 706, 708
    coalfield, Nfld. 356
    Lowland 28, 29
    River drift 707
St. Helen's Island 278
Saint John, N.B. 232, 248, 296, 711
  anticline 269
  Bay 28
  Group 240
  Highlands 241
  River 29
    Valley 711, 712
St. Juste Group 273, 275
  Lawrence, Nfld. 283, 309, 340
    Lowlands 17, 19, 28
      Hydrogeological region 776, 777
    Mineral Province 311–340, 354–362
    Platform 228, 240–242, 249, 250, 255–257, 259–261, 268–272, 277, 278, 283, 284, 300, 309
    River 17, 27, 125, 706, 726, 776, 777
      Valley 602, 713, 726
  Léon Formation 263, 264, 273
  Louis fault 167
  Mary River Formation 464
  Mary's Bay 706
    River 730

PAGE

St. Mary's fault 167, 450
  Narcisse, Que. 726
    Till 682
  Paul Island 274, 708
  Paul's Inlet 354
  Pierre Bank 29
    les Bequets, Que. 682, 683
    sediments 682, 683, 685, 689
  Rémi deposit, Que. 222
  Rose–Chimney Corner coalfield, N.S. 356
  Stephen, N.B. 265, 775
    ultramafic body 251, 316
  Urbain (70) mine, Que. 154, 156, 183, 185
Sakami Formation 59, 104
Salina Formation .. 143, 271, 277, 342, 343, 348, 777, 780
  River Formation 381
Salisbury Island 15
Salmo, B.C. 504, 506, 516, 517
  subtype 505
Salmon Lake Formation 126
Salt deposits 341, 519
  River 519
Salter Member 420
San Andreas fault 369, 476
  Antonio (18b) mine, Man. 207
    Formation 67, 68
San Juan Islands 400, 417
Sandpile Group 391, 605
Sandstone deposits 344
Sangamon Interglacial Interval 677, 681, 689, 694, 696, 703–705
Sans Sault Group 455
Sarcee gas field, Sask. 542
Sargent Bay Formation 263, 605
Sarnia (117) brine field, Ont. 342
Saskatchewan gravels 692
  Plain 20
  River 20, 740, 741
Sassenach Formation 403, 404
Sault stage 731
  Ste. Marie, Ont. 109, 114, 186, 196, 200, 235, 729–731
Saunders, Alta. 524
Savanna Creek (113) gas field, Alta. 542
Savant Lake 67
Savik Formation 579, 581, 583
Sawtooth Formation 445, 538
  Range 582
Sayabec Formation 263, 264, 605
Scarborough Bluffs, Ont. 684
  Formation 684–686
Scatter Formation 459
Scented Grass Hills 20
Schefferville, Que. 102, 104, 297, 737
  (75) mines, Que. 171
Schei Peninsula 751
  Point Formation 575–579, 582, 635
Schist Lake (4b) mine, Man. 198
Schooler Creek Group 436
Schreiber, Ont. 195
Scoharie Formation 277
Scotian Shelf 29, 297–300

PAGE

Scotland ... 134
Scots Bay Formation ... 298
Scotsburn, N.S. ... 710
Seal Lake ... 126
  Group ... 120, 121, 126, 127
  River ... 97, 689, 697
Seboomook Formation ... 273, 275
Sechart, B.C. ... 516
Second (Lower) White Specks ... 463, 464
Seine Bay ... 185
Seismic surveys ... 54, 300
Sekwi Mountain ... 381
Selenium deposits ... 220
Selkirk anvil structure ... 380
  Group ... 480
  Mountains ... 375, 380–382, 391, 406, 415, 426, 450, 481, 498
Selwyn Basin ... 375, 385, 388, 392, 394, 400, 402, 444, 448
  Fold Belt ... 450, 452
  Mountains ... 21, 24, 25, 373, 374, 378, 381, 388, 390, 418, 452, 453, 481, 702, 755
Semiamu Glaciation ... 704
Seminary Till ... 685
Seneca Member ... 278
  Township, Ont. ... 342
Sentinel Range ... 399
Sept Iles Bay deposit, Que. ... 185
Serpent Formation ... 112
Serpentine belt ... 307
Seven Mile Island ... 688
Severn Arch ... 138, 140, 385
  River Formation ... 140, 142, 605
  Upland ... 15, 16
Seward Formation ... 102
  Member ... 409
Sextant Formation ... 144, 617, 663
  Rapids ... 145
Shabogamo gabbro ... 104
Shadow Lake Formation ... 255
Shaftesbury Formation ... 459, 460
Shakwak fault ... 369, 423, 475
  Glaciation ... 702, 705
  Trench ... 23, 26, 475, 478, 790
Shale deposits ... 345, 521
Shaler Group ... 77, 84, 88, 589
  Mountains ... 19
Shamatawa River ... 689
Sharon Creek Formation ... 391
Shatford Lakes ... 220
Shaunavon, Sask. ... 784
  Formation ... 445, 538
Shawinigan Falls, Que. ... 124
Shawville, Que. ... 178
Shediac, N.B. ... 772
Sheep River ... 524
Sheguiandah Formation ... 259
  phase lake ... 729, 730
Shepody Mountain ... 333
Sheppard Formation ... 371
Sherard Osborn Formation ... 562

PAGE

Sherbrooke, Que. ... 320
  Formation ... 257
Sherman Fall Formation ... 256, 346
  (95) mine, Ont. ... 175
Sherridon, Man. ... 90, 91, 196, 198
  Group ... 90, 91
Sherritt-Gordon (2) mine, Man. ... 196, 198
Shickshock granite ... 318
  Group ... 237, 246, 251, 252, 254
  Mountains ... 29, 245, 319, 712
Ship Point Formation ... 557
Shipiskan plateau ... 126
Shipshaw Formation ... 257
Shoal Arm Formation ... 255
Shock metamorphic structures ... 135–137
Shubenacadie, N.S. ... 298
Shublik Formation ... 435
Shulaps ultramafic body ... 423
Shunda Formation ... 418, 420
Shuniah vein, Ont. ... 214, 215
Shuswap anvil structure ... 450
  Lake ... 23, 704
  Metamorphic Complex ... 373, 375, 381, 416, 425, 426, 434, 450, 506
  subtype ... 506
Siamarnekh Formation ... 121
Siberia ... 575
Sibley Formation ... 118, 213
Sicker Group ... 417, 435, 520
Sickle Group ... 91, 92
Sidney, B.C. ... 788
Sifton Formation ... 451, 461, 470
Sigma (34d) mine, Que. ... 205
Sikanni Chief River ... 436
  Formation ... 459
Silbak Premier (10b) mine, B.C. ... 504, 506
Silica deposits ... 223
Sillery Formation ... 236
Silurian ... 140–143, 261–272, 385–392, 555–560, 601–613
Silver Cliff mine, Nfld. ... 330
  deposits ... 208–215, 330, 495, 504, 506
  Giant (100) mine, B.C. ... 518
  Islet mine, Ont. ... 210, 213
  King (87c) mine, B.C. ... 506
  Mountain mine, Ont. ... 214
  non-glacial interval ... 705
  Plains (22) mine, Man. ... 519
  Regent mine, Ont. ... 210
  Standard (18) mine, B.C. ... 504, 506
  Town mine, Ont. ... 210
Silverfields mine, Ont. ... 210, 211
Simard Formation ... 257
Simcoe, Ont. ... 307
  ice lobe ... 728
Simms Ridge Formation ... 267
Simonette (26) gas field, Alta. ... 533, 534
Simpson Range ... 416
Sims Formation ... 104, 123
  Lake ... 104
Sinwa Formation ... 434

PAGE

Sioux Lookout, Ont. ... 732
moraine ... 731, 733
Sir James Mine, Ont. ... 174
Sirois Formation ... 264
Siscoe (31a) mine, Que. ... 205
Sixtymile River placer field, Y.T. ... 517
Siyeh Formation ... 371, 372
Skaare Fiord ... 585
Skaergaard Intrusion, Greenland ... 121
Skagit metamorphism ... 428
Range ... 26
Skeena Arch ... 438, 442, 443, 452
Mountains ... 25, 756
River ... 25, 417, 427, 431, 441, 522, 654
Skidegate Formation ... 460, 654
Plateau ... 27
Skiff oil field, Alta. ... 538
Skoki Formation ... 391, 604
Skonun Formation ... 479
Skukum Group ... 478
Slater River Formation ... 455
Slave Point Formation ... 403, 408, 527, 616
Province ... 45, 71-76, 155
River ... 519, 614, 741, 745, 785
Slide Mountain Group ... 415
Slidre Fiord ... 701
Slocan, B.C. ... 504
Group ... 434
Sloko Group ... 478, 504
Smith Island ... 104
Point Formation ... 240
Smithfield, N.S. ... 329
Smoky Group ... 461, 463
Lake ... 695
River ... 435, 436, 445, 455–458, 691, 740, 741
Snake River ... 20
deposit, Y.T. ... 498
Snakes Bight Formation ... 287
Snare Group ... 77–80
Snipe Lake ... 537
Snooks Arm Group ... 247, 251, 267
Snow Lake ... 196
Snowbird Lake ... 92, 93
Snowblind Bay Formation ... 550, 561
Snowshoe Formation ... 406, 509
Soab orebody, Man. ... 181
Soapstone (steatite) deposits ... 340
Sobie Lake ... 729
Société Québecoire d'exploitation Minière ... 3
Society Cliffs Formation ... 101
Sodium sulphate (Glauber's salt) deposits ... 519
Sokoman Formation ... 102
Solbec (56b) mine, Que. ... 320
Sombre Formation ... 399
Somerset Island ... 19, 88, 100, 548, 549, 554, 557, 558, 561, 584, 587, 616, 743, 744
Sooke Formation ... 479
Soper Highland ... 19
Sophie Mountain Formation ... 461
Sop's Arm ... 330
Sops Island Member ... 267
Sorel, Que. ... 36

PAGE

Sosan Formation ... 95
Souris River ... 740, 786
Formation ... 409, 519
Valley Formation ... 419, 534
South Channel Formation ... 98
Fiord dome ... 586
Ham mine, Que. ... 315
Henrik Lake ... 746
Lorrain deposits, Ont. ... 208, 209, 211
Nahanni River ... 24, 381, 614, 625
Redstone River ... 373
Saskatchewan River ... 498, 691, 695
Sturgeon Lake (21b) gas field, Alta. ... 534
Southampton Island ... 15, 137, 140, 748, 750
Plain ... 15
Southampton, Ont. ... 780
Southeast Newfoundland Ridge ... 29
Southern Hills Formation ... 232, 234
Plateau and Mountain Area ... 25
Province ... 47, 108–119, 156
Southesk Formation ... 403, 409, 410
Southwold Till ... 687
Spain mine, Ont. ... 219
Spalding, Sask. ... 694
Spanish American mine, Ont. ... 166
Spearhill, Man. ... 520
Spences Bridge Group ... 454
Spillimacheen, B.C. ... 518
Spinney Hill ... 460
Spirit River Formation ... 459
Spitsbergen ... 575
Spray River Formation ... 420, 522
Group ... 435
Spring Valley, Sask. ... 695
Springbrook Formation ... 477
Springdale Group ... 267, 319
Springhill coalfield, N.S. ... 356–358
Springvale Member ... 277
Sproule Lake ... 220
Spuzzum Intrusions ... 428
Square Lake deposit, N.B. ... 335
Squaw Creek ... 478
Stall Lake (10) mine, Man. ... 198
Stallworthy Formation ... 560, 561
Stanbridge Formation ... 254
Stanford Range ... 405
Stanleigh mine, Ont. ... 166
Stanrock mine, Ont. ... 166
Starbird Formation ... 406
Stark Formation ... 95
Starlight Evaporitic Member ... 436
Steatite deposits ... 340
Steel Mountain deposit, Nfld. ... 185, 333, 335
Steelman (42a) oil field, Sask. ... 534
Steep Rock, Man. ... 520
iron range, Ont. ... 170, 177
(21) mine, Ont. ... 177
Steeprock Lake ... 66, 67, 153, 731, 757
Group ... 67
moraine ... 731
Stefansson Island ... 557–559

PAGE

Stellarton Gap ... 289, 291, 294, 295
Group ... 294
Stemwinder orebody, B.C. ... 510
Stephen Formation ... 383
Stephenville, Nfld. ... 776
Sterco, Alta. ... 524
Stettler Member ... 410
(89b) oil field, Alta. ... 532
Stewart Plateau ... 451
River ... 703, 705, 755
Valley ... 702, 703
Stikine Arch ... 411, 431, 432, 434, 438, 440, 443, 476, 514
Plateau ... 23, 25, 416, 417
River ... 26, 416, 417, 427, 428, 514, 705
Stobie Formation ... 109
Stoddart Group ... 418
Stone Range ... 403
Stonehouse Formation ... 276, 277, 605
Stonewall Formation ... 392, 520, 605
Stoney Creek gas field, N.B. ... 309, 346, 354
Lake ... 276
Stony Mountain Formation ... 140, 392, 520, 605
Stooping River Formation ... 144
Storelk Formation ... 420
Stoughton Member ... 392
Stowbridge deposit, Nfld. ... 336
Strait of Belle Isle ... 27, 29, 231, 240, 241, 706–708
Canso ... 291, 708, 709
Georgia ... 26, 27, 460, 703, 704, 755
Strand Fiord ... 583, 584, 586, 589
Formation ... 583
Strathcona deposit, Ont. ... 180
Fiord ... 582
Sound Formation ... 101
Strathlorne Member ... 287, 288
Structural Provinces, Canadian Shield ... 45, 46, 154
Stuart Bay Formation ... 550, 559, 561, 563, 616
River ... 701, 756
Stuhini Group ... 432, 434
Sturgeon Lake (24) oil field, Alta. ... 537
River (27) mine, Ont. ... 206
Subpermafrost ... 790
Sudbury, Ont. ... 36, 109, 112, 114, 117, 152, 176, 179, 182, 183, 220, 222, 728
Basin ... 108, 117
Contact mine, Ont. ... 210
(78 to 90) district, Ont. ... 178
gabbro ... 109, 115, 117
Irruptive ... 108, 109, 116, 117
Suffield (Ascot) (53c) mine, Que. ... 320
Sugarloaf Island ... 316, 317
Sugluk Plateau ... 16
Sukunka River ... 537
Sullivan Consolidated (31b) mine, Que. ... 205, 781
(105b) mine, B.C. ... 371, 504
River ... 451
subtype ... 504
Sully Formation ... 459
Sulphur deposits ... 223, 522, 589
Mountain Formation ... 435, 436
Point Formation ... 407, 495, 527, 614

PAGE

Sumas Stade ... 755
Summerside Formation ... 238
Sunblood Formation ... 388, 604
Sunbury shale ... 284
Sundre (103) gas and oil field, Alta. ... 535
Sunkay Member ... 462
Sunnybrae, N.S. ... 268
Sunnybrook Till ... 685
Sunridge, Ont. ... 728
Sunro (55) mine, B.C. ... 510
Sunset (73b) mine, B.C. ... 510
Superior Basin Lakes ... 230, 729
Province ... 45, 54–71, 155
Superior-type iron-formation ... 159, 170–174
Suprapermafrost ... 790
Suquash, B.C. ... 522
Surprise Creek Formation ... 139
Lake ... 122, 123
Surrey, B.C. ... 704
Survey Peak Formation ... 391, 603
Suspension Bridge Formation ... 249
Sussex, N.B. ... 772
Sustut Group ... 460, 476
Trough ... 448
Sutherland River Formation ... 559
Sutton Group ... 236, 245, 320, 636
Inlier ... 29, 47, 105, 137
Lake ... 174
Mountain anticlinorium ... 236
Mountains ... 29, 252, 713
Ridge ... 15
Svartefjeld Peninsula ... 572
Svartevaeg Formation ... 561, 569
Sverdrup Basin ... 548, 550, 552, 566, 568, 569, 571, 572, 574, 584, 589, 633, 636, 663
Lowland ... 18
Pass ... 555
Swampy Lake ... 737
Swan Hills ... 481
Member ... 409, 528
River Group ... 460
Sweetgrass Arch ... 419, 420, 438, 440, 445, 468, 784
Swift Creek ... 498
Current, Sask. ... 522
Arch ... 538
Creek ... 481
Formation ... 695
Platform ... 378, 384, 410
Formation ... 445
Swinnerton Peninsula ... 701
Swiss Range ... 586
Sydney, N.S. ... 309, 356, 680
Basin ... 287, 290–295
coalfield, N.S. ... 356, 357
phase lake ... 727, 728
Sylvania Member ... 277
Sylvanite (60b) mine, Ont. ... 204
Sylvester Group ... 402, 416, 417, 423
Trough ... 402, 411

Taber, Alta. ... 695

PAGE

Taber Formation 524
Table Head Formation 247, 250, 252, 329, 344, 601
  Island 575
Taconian Orogeny .. 229, 243–245, 247, 250–253, 257, 313
Taconic klippe 252
Tahkandit Formation 418
Tahltanian Orogeny 368, 369, 417, 424, 428, 431, 432
Tahtsa Complex 427
Taiga Ranges 24, 25, 367, 373, 381, 473
  Valley 25
Takiyuak Lake 79
Takla Group 434, 441–443, 515
Takomkane batholith 440
Taku River 26
Takwahoni Formation 440
Talbot Plateau 703
Talc deposits 223, 340, 518
Talisman (Silver) mine, N.S. 329
Tangent gas field, Alta. 537
Tanquary Fiord 571, 573, 578, 586
  Formation 571, 573
Tantalum deposits 220
Tantalus Basin 438, 442, 443, 448, 452, 454
  Formation 443, 454, 522
Tartu Inlet facies 477
Taseko Creek 517
  Lakes 23, 441
Tasu (25) mine, B.C. 515
Tathlina Arch 394, 406, 407
  Formation 408, 410
  Upland 394, 406
Tay River 435, 454
Taylor Creek Group 453
  Flat Formation 418
Tazin Group 92, 93, 167, 208
Teck-Hughes (61f) mine, Ont. 204
Tectonic Map of Canada (1251A) 5
Tegart Formation 391
Telford Formation 420
Telkwa coalfield, B.C. 522
Tellurium deposits 220
Temagami (94) mine, Ont. 196
Temiscamie Formation 59
Temiscouata Formation 273, 275
  Lake 263, 273, 274
Temiskaming, Que. 733
Ten Mile Lake 708
Tenakihi Group 372, 425
Tennycape deposit, N.S. 333
Tent Mountain 524
Terminal Range 380, 382, 399, 403
Terra Nova (5) deposit, Nfld. 318, 325
Terrace, B.C. 481, 517
Terranceville, Nfld. 617
  Formation 283
Terrebonne Formation 602
Tertiary 297, 299, 464, 476–481, 584–588, 665–668, 663–670
Teslin area, Y.T. 424
  Lake 416
  River 434, 440

PAGE

Teslin Valley 23
Tetagouche Falls 334
  Group 254, 257, 324, 333
Tetcho Formation 410
Tetreault (69) mine, Que. 199, 200, 209
Tetreauville Formation 256
Tetsa Formation 372
  River 459
Texada Island 515, 520
  (49) mine, B.C. 515
The Gully 29, 299, 300
  Pas, Man. 196
Thelon Formation 76, 94, 99
  Front 47, 53
  Plain 11, 13
  Plate 47
  River 746
  Subprovince 98
Theresa Formation 242, 243, 346
Thessalon Formation 109
Thetford Mines, Que. 29, 307, 336, 338
Thirty-one-mile Lake 222
Thistle Member 463
Thompson, Man. 36, 155, 178, 220
  Belt 70, 88, 180
  District (12, 13), Man. 180–181
  Lake Formation 102
  River 479, 704
Thorium deposits 163
Thorncliffe Formation 685
Thornloe Formation 142, 271, 605
Thor–Odin Dome 425
Thorold Formation 270, 351
Thousand Island 17
Thule Group 555
Thulean volcanicity 583
Thumb Mountain Formation 557
Thunder Bay 731
  District, Ont. 161, 208, 209, 212–215
Thuya batholith 440, 442
Tilbury gas field, Ont. 309, 348, 350
Tilston Beds 420
Tilt Cove (7) mine, Nfld. 307, 317, 325
Timagami, Ont. 175, 182
  Lake 186, 196
Timiskaming, Ont. 66
  Group 62, 63, 202, 203
Timmins, Ont. 37, 60–61, 63, 209, 734
  District, Ont. 153, 182, 185, 186, 191
Tin deposits 220, 335
Tindir Group 373, 374
Tingmisut Lake 583
Tingwick township, Que. 337
Tinney Cove Formation 76, 84
Tintina area, Y.T. 435, 444, 455
  Trench 25, 390, 399, 402, 416, 417, 424, 448, 451, 452, 454, 478, 633, 635
Tipperary Formation 391
Tisdale Group 202
Titanic iron 183
Titanium deposits 156, 183, 184, 335
Toad Formation 436

PAGE

Tobermory Formation ... 420
Toburn (60c) mine, Ont. ... 204
Toby Formation ... 375
  stock ... 415
Tochatwi Formation ... 95
Tombstone area, Y.T. ... 435, 444, 451, 452, 455
Topley Intrusions ... 440, 441, 514, 515
Torbrook, N.S. ... 307
  Formation ... 277
Toric (12b) mine, B.C. ... 504, 506
Torngat Glaciation ... 697
  Mountains ... 15, 697, 734, 735
Toronto, Ont. ... 681, 684
Torpy River ... 472
Torquay Formation ... 410
Tovell Member ... 459
Trail, B.C. ... 495, 515
Trebor mine, Ont. ... 182
Tremblay Formation ... 257
  Sound ... 101
Trembleur Intrusions ... 423
Trenton Group ... 255–257, 259, 260, 328, 344, 346, 354, 602
  phase lake ... 713, 727
Trent Valley ... 728
Treytown, Nfld. ... 283
  Pond Formation ... 241
Triassic ... 296–298, 428–438, 574–578, 633–640, 663–670
Trinité Group ... 254, 263
Trinity Bay ... 239, 249, 280, 706, 708
  Belt ... 230
  Trough ... 230, 234, 235
Triune Formation ... 391
Trois Pistoles, Que. ... 712
Trois-Rivières, Que. ... 681, 682, 726
Trold Fiord ... 560, 561, 572, 578, 582
  Formation ... 572–574
Trout Brook Formation ... 240
  Creek, Ont. ... 728
  Lake ... 731
  River Formation ... 408, 410
Truro, N.S. ... 772
Tsawwassen, B.C. ... 788
Tuchodi Formation ... 372
Tuktoyaktuk Peninsula ... 696, 742
Tulameen, B.C. ... 517
  ultramafic body ... 423
Tully Formation ... 617
Tulsequah Chief (3) mine, B.C. ... 504
Tungsten deposits ... 220, 335, 515
Tunnel Mountain Formation ... 420
Turnagain River ... 461
Turner Cliffs Formation ... 555
  Valley Formation ... 420, 535, 542
  Valley (112) oil field and gas, Alta. ... 36, 490, 542
Turtle Creek ... 333
  Mountain ... 479
    Formation ... 479
Tuya Formation ... 480
  River ... 705
Tuyas ... 23

PAGE

Twin Falls Formation ... 408, 409
Tyaughton Creek ... 633
  Formation ... 434, 438, 636
  Trough ... 442, 443, 448, 453, 454
Tyee (54) mine, B.C. ... 504, 510
Tyndall building stone ... 520
Tyrrell Sea ... 690
  deposits ... 688
Tyrwhitt Formation ... 420

Ubly River ... 726
Uchi belt ... 157, 159, 201, 207
  Lakes ... 67
Uluksan Group ... 101
Ungava Bay ... 15, 16, 101, 104, 153, 171, 198, 221, 734–738
Union Island ... 94
    Group ... 94
United Keno (7) mine, Y.T. ... 506
United States Ranges ... 17, 585, 752, 753
Unity, Sask. ... 519
Uplands of Nova Scotia ... 28
Upper Canada (59f) mine, Ont. ... 204
Uranium City, Sask. ... 92, 93
  deposits ... 155, 156, 161, 163
Uslika Formation ... 416, 450, 454
Utah ... 318, 603
Utica Formation ... 260, 354

Val Brilliant Formation ... 263, 605
Val d'Or, Que. ... 191, 221
Valders Age ... 731, 732
  ice advance ... 728
Valhalla Dome ... 425, 426
"Valley" olivine basalts ... 480
Van Hauen Formation ... 573, 574
Vanadium deposits ... 336
Vancouver, B.C. ... 369, 520
  Island ... 26, 27, 417, 435, 442, 443, 454, 460, 467, 475–477, 479, 498, 515, 518, 520, 522, 707, 755
    Ranges ... 27
Vanderhoof Strait ... 654
Vanguard Formation ... 445, 538
Vanscay mine, Sask. ... 520
Vashon Glaciation ... 755
  stade ... 754
Vauréal Formation ... 261
Vedder "greenstone" ... 400
  Mountain ... 521
Vega Siltstone Member ... 436
Velse Fiord ... 582
Vendom Fiord ... 564, 566
Verchoyle oil field, Ont. ... 346
Vermilion Pass ... 385
  River Formation ... 464, 521
Vermont ... 29, 252
Verna (2) mine, Sask. ... 167
Vernon, B.C. ... 416
Verona, Ont. ... 223
Verulum Formation ... 256
Vicary coal mine, Alta. ... 524

PAGE

Victor Bay Formation 101
Victoria, B.C. 704
and Albert Mountains 18, 585
Island 19, 84, 548, 549, 554, 557–559, 700, 743
Formation 614
Lowland 19
View Creek Group 560
Viking Formation 459, 460, 539
Viking-Kinsella (69) gas field, Alta. 539
Ville-Marie, Que. 200
Vimy Member 463
Virden (2, 3) oil field, Man. 534
Virden-Whitewater Lake limestones 534
Virgin Rocks 29
Virginnia Falls 614
Viscount mine, Sask. 520
Vittrekwa River 435, 444

Wabamun coalfield, Alta. 524
Group 410, 522, 534, 544
Wabana Group 249
(16) mine, Nfld. 309, 331, 332, 334, 336
Wabash River 726
Wabassee Group 193
Wabi Group 142, 271, 605
Wabigoon Belt 45, 157, 159
Wabiskaw Member 459
Wabowden orebody, Man. 181
Wabush Lake 39
(1) mines, Labrador 171, 173
Wabush–Mount Wright, Que. 121, 123
Wager Bay 10, 13, 100, 747
Plateau 63
Waite (17f) mine, Que. 191
Wakefield (44) mine, Que. 219
Wakeham Bay 104
Group 53, 125, 126
Wallace mine, Ont. 154
Walton, N.S. 309, 327–329, 340
(6) mine, N.S. 338
Wapella oil field (24), Sask. 538
Wapiabi Formation 461, 463, 464, 540
Lake 436
Wapussakatoo Mountains 123
Ward Hunt Island 752
Wasa, B.C. 495
Wasekwan Group 91, 92
Washington State 382, 391, 400, 428, 477
Waswanipi Lake 138
Water table 766
Waterfowl Formation 385, 451
Waterloo, N.Y. 727
Waterton Formation 371
(141) gas field, Alta. 522, 542, 543
Waterways Formation 408, 616
Watrous Formation 445, 535
Watson Lake, Y.T. 424
Group 193
Watt Mountain Formation 408
Waverly Arch 242
Wawa Belt 45
Lakes 688

PAGE

Wayne Glaciation 726
Weatherall Formation 559, 561, 563–565
Island 589
Wedge (4) mine, N.B. 324
Weedon Formation 246
(55) mine, Que. 320
Weir Formation 264
Wekusko Lake 90
Wekwemikongsing Formation 259
Welland gas field, Ont. 307, 351
Wellington Arch 549
Bay 84
Wendigo (16) mine, Ont. 201
Wernecke Mountains 24, 367, 373, 374, 378, 381, 388, 399, 452, 473
Werner Lake 67
West Bay fault 75, 207
Gore mine, N.S. 316
Malartic (26c) mine, Que. 205
Point Formation 264, 265, 605
St. Lawrence Lowland 27
Western Columbides 450, 453
Mineral Province 491, 498
System 21, 23, 26
Westmorland county, N.B. 355
Westport, Ont. 127, 129, 131
Weyburn (36b) oil field, Sask. 534
Whale Bank 29
Lowland 15
River 15, 735
Whalesback Pond mine (10), Nfld. 325, 327
Wheaton Creek 518
Whirlpool Formation 268, 344, 351, 605
Whiskey Lake Formation 126
Whistler Member 436
White Basin 287, 290, 293
Bay 28, 250, 267, 279
Channel gravels 481
Creek batholith 450
Head Formation 254, 264, 275, 605
Lake Formation 478
Mountains 29
River 418, 702, 756
Rock Formation 255, 268
Whitefish Bay 729
Falls 154
Lake 118
Whitehorse, Y.T. 440, 522, 790
Formation 435, 436, 438, 519
Trough 431, 432, 434, 438, 440, 441, 442, 452
Whitelaw gas field, B.C. 537
Whitemud Formation 464, 521
Whitesail Lake 417
Whitewater Group 108, 117, 178
mine, B.C. 504
Moraine 731
Whitsunday Bay 572, 586
Whittaker anticline 390
Formation 388, 390
Range 390
Whycocomagh, N.S. 680
Wiarton Member 270, 272

PAGE

Wigwam Creek ... 435
deposit, B.C. ... 506
Wild Bight Group ... 247, 255
Wildmint (14) oil field, B.C. ... 537
Wileman Member ... 420
Wilkens Strait ... 751
Wilkie Point Formation ... 552, 575, 579, 581
Willbob Formation ... 102
Willecho (34) mine, Ont. ... 193, 194
Willesden Green (38) oil field, Alta. ... 540
Willey oil field, Ont. ... 346
Williams Creek ... 517
Island Formation ... 144, 145, 617
Williston Basin ... 54, 140, 144, 145, 262, 385, 388, 391, 392, 411, 419–421, 430, 438, 440, 445, 479, 534, 784
Willmar (43) oil field, Sask. ... 534
Willow Creek Formation ... 464, 478, 692
Willroy (35) mine, Ont. ... 193, 194
Wilrich Member ... 459
Wilson Island Group ... 73, 94
Windermere System ... 371, 374, 380
Windfall (31) gas field, Alta. ... 533
Windigo belt ... 157, 159
Lakes ... 731
Windsor Group ... 286–291, 295, 298, 329, 333, 340, 342, 344, 626, 663, 772
(7) mine, N.S. ... 344
(118) mine, Ont. ... 342
Wingfield Formation ... 269, 605
Winisk River ... 732
Winnifred Member ... 436
Winnipeg Formation ... 391
River district, Man. ... 217–220
Winnipegosis Formation ... 407, 519, 614
Winter Harbour moraine ... 742
Winterburn Group ... 408, 409
Wisconsin ... 243
Glaciation ... 677, 705–756
Wishart Formation ... 102
Wokkpash Formation ... 399
Wolf Island Sediments ... 692
Lake ... 424
pluton ... 127
Wolfville, N.S. ... 249, 268
Formation ... 298
Wollaston Lake ... 746
Wolverine Complex ... 372, 375, 415, 416, 424, 425
Wonah quartzite ... 391
Wood Mountain, Sask. ... 522, 690
gravels ... 481

PAGE

Woodbend Group ... 408–410, 528, 533
Woodbridge, Ont. ... 685
Wopmay Belt ... 45, 77
River ... 80
Workman (52b) oil field, Sask. ... 534
Worsley (4) gas field, Alta. ... 533
Worth Point ... 700
Wrangell–Revillagigedo belt ... 427
Wright-Hargreaves (60d) mine, Ont. ... 200, 204
Wuchusk Lake Formation ... 126
Wunnummin Lake ... 732
Wyebridge phase lake ... 728
Wymark Member ... 409
Wynniatt Formation ... 84, 589

Yakoun Formation ... 442, 454
Yalakom fault ... 432
River ... 518
Yankee Belle Formation ... 381
Yarmouth, N.S. ... 249, 268, 710
Interglacial Interval ... 677, 681
Yedhe Mountain ... 380
Yellow Aster Complex ... 400
Yellowknife, N.W.T. ... 200, 201, 212, 218
Bay ... 207
(8, 9) Camp, N.W.T. ... 207–208
Group ... 72–74, 76, 77, 100
Yellowknife–Beaulieu River district, N.W.T. ... 217, 220
Yeoman Member ... 392
Ymir Group ... 434, 442
Yohin Formation ... 418
York River Formation ... 274, 275
Till ... 684
Youngs Cove Formation ... 240, 267
Yukon Coastal Plain ... 18, 21, 25, 695
Crystalline Platform ... 367, 427, 438, 440, 475
Group ... 375, 416, 424, 440
Plateau ... 20, 22, 25, 453, 479, 481
River ... 22, 25, 378, 388, 400, 418, 424, 434, 703, 705, 755
Territory ... 701

Zama Lake (1) oil field, Alta. ... 527
North (1) oil field, Alta. ... 527
Zeballos Iron (42) mine, B.C. ... 515
Zeemel fault ... 167
Zenmac (Zenith) (33) mine, Ont. ... 186, 195
Zig-Zag Lake ... 218
Zinc deposits ... 185, 199, 495, 504, 589
Zymoetz River ... 517







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