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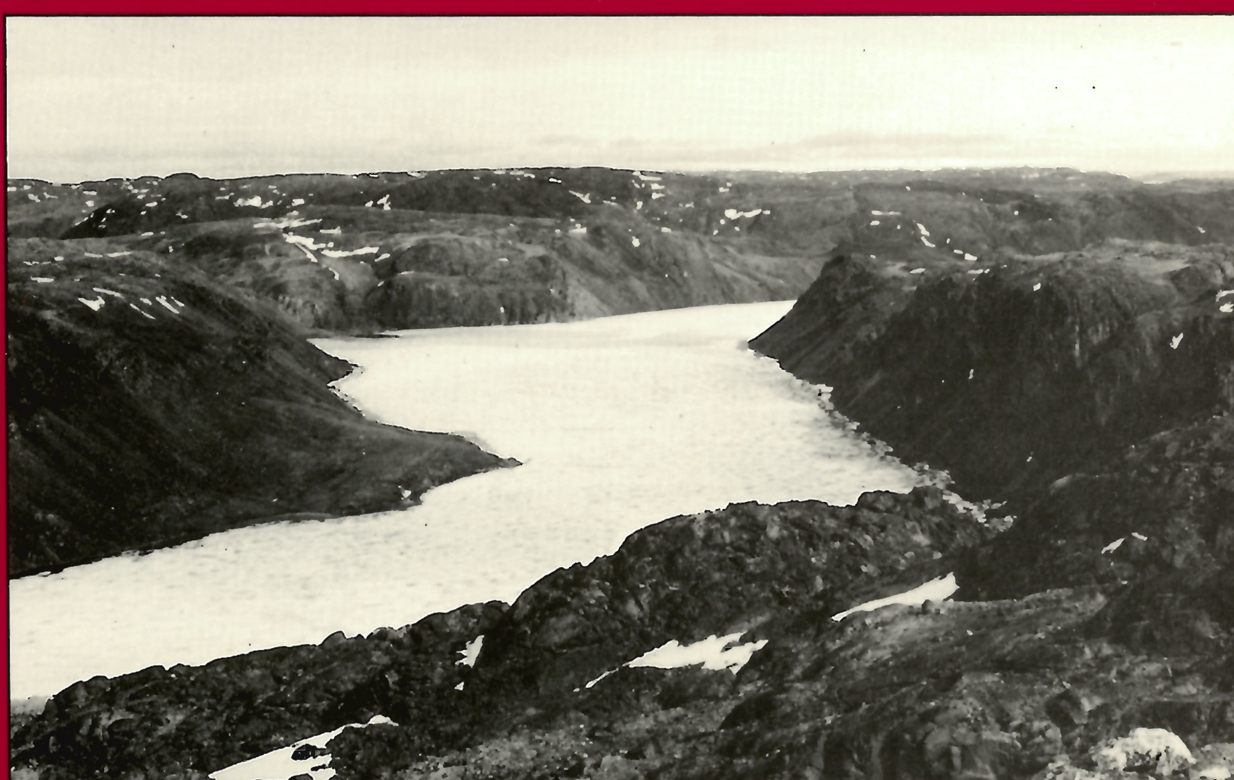
Commission géologique du Canada

MEMOIR 407

QUATERNARY GEOLOGY OF BOOTHIA PENINSULA AND NORTHERN DISTRICT OF KEEWATIN, CENTRAL CANADIAN ARCTIC

ARTHUR S. DYKE



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Cover

View east along Amituryouak Lake, a fiordic lake on northern Boothia Peninsula. Ice flowed eastward along this trough during the last glaciation. (GSC 203359-V)

Critical reader

D.A. Hodgson

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Preface

The search for hydrocarbons in the Arctic Island which began in the late 1950's following the initial geological reconnaissance made by the Geological Survey in 1955 resulted in the identification of resources of natural gas. As a result studies were undertaken of terrain conditions in areas which would be crossed by any transportation corridors needed to exploit these resources. The study of the Boothia Peninsula part of a proposed corridor was initiated in 1974 and preliminary maps of the terrain were released to the public soon after compilation of field work. More detailed work was carried out in subsequent years to elucidate the broad, fundamental aspects of the recent geological history of critical areas.

This report describes the surficial geology of 64 000 km^2 of Canada's most northern mainland and complements a similar report on Somerset Island which lies immediately north of Boothia Peninsula being separated from it by the very narrow waters of Bellot Strait. The report discusses the potential use of geological information to future land use projects and to mineral exploration and presents new interpretations regarding the structure and dynamics of the northern part of the last ice sheet to cover Canada.

OTTAWA, June 1982

R.A. Price
Director General

Préface

La recherche d'hydrocarbures dans l'archipel Arctique qui a été mise en branle à la fin des années 1950 à la suite des premiers levés géologiques effectués dans la région, en 1955, par la Commission géologique du Canada a mené à la découverte de gisements de gaz naturel. À la suite de quoi, on a entrepris de déterminer les conditions du terrain dans les régions qui seraient appelées à être traversées par les couloirs de transport nécessaires à l'exploitation de ces ressources. Suite à une étude faite en 1974 de la presqu'île de Boothia où l'on envisage de faire passer un couloir, des cartes préliminaires du terrain ont été publiées peu après la compilation des données. Au cours des années subséquentes, des travaux plus poussés en vue d'élucider les aspects généraux et fondamentaux de l'évolution géologique récente de régions critiques ont été réalisés.

Ce rapport décrit la géologie des formations superficielles de 64 000 km^2 dans la partie la plus septentrionale de la terre ferme du Canada et vient compléter un rapport semblable sur l'île Somerset qui est située juste au nord de la presqu'île de Boothia, de l'autre côté du très étroit détroit de Bellot. On analyse également dans ce rapport la possibilité de se servir de l'information géologique pour des projets d'utilisation des terres et la recherche de minéraux. Enfin, on présente de nouvelles interprétations au regard de la structure et de la dynamique de la partie septentrionale de la dernière nappe glaciaire du Canada.

OTTAWA, juin 1982

R.A. Price
Directeur généraux

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QUATERNARY GEOLOGY OF BOOTHIA PENINSULA AND NORTHERN DISTRICT OF KEEWATIN, CENTRAL CANADIAN ARCTIC

Abstract

A surface of erosional planation, of regional extent in the central Canadian Arctic, was fragmented during the Eurekian Rifting Episode (Miocene-Pliocene) to produce the plateaus, plains, and lowlands of the study area and the rift valleys occupied by the large marine channels.

Seven genetic groups of materials – rock, till, and glaciofluvial, glaciolacustrine, glaciomarine, marine, and fluvial sediments – constitute the surface of the map area, but rock and till are predominant. The till sheet exhibits major lateral facies changes which define two large carbonate dispersal trains. Thick glaciomarine silt and clay extend in a wide belt across the southern part of the map area.

Aspects of possible future engineering concern include: scarcity of sand and gravel in much of the area, natural instability of most unconsolidated material on slopes, high ground-ice content of thick till and fine grained glaciomarine sediment, annual thermal contraction and cracking of soil, frost heaving of soil and bedrock, ice push at sea and lake shores, and ice scouring of river beds and banks.

Analyses of about 750 till samples provide the initial geochemical data base for this region. Variations in background levels of base metals reflect variations in carbonate content.

All surface sediments shown on Map 1570A were deposited during the late Wisconsin glaciation. During the late Wisconsin maximum a major dome of the Laurentide Ice Sheet had a north-south oriented ice divide located over M'Clintock Channel. M'Clintock Dome was contiguous with the main body of the ice sheet to the south and coalesced with the Foxe Dome at the base of the Gulf of Boothia. M'Clintock Dome generated an eastward to east-northeastward flow over Boothia Peninsula and northern District of Keewatin. Important features of this flow regime were two large ice streams at the base of the ice sheet.

Deglaciation began more than 9250 years ago and initial retreat was westward, towards the centre of the M'Clintock Dome. Retreat swung gradually to the south as a calving bay along the western side of Somerset Island and Boothia Peninsula penetrated the central parts of the dome. Boothia Peninsula was largely ice free by 8800 years B.P. and the marine-based part of the M'Clintock Dome had completely disappeared by 8700 years ago. A moraine system of regional extent in Arctic Canada, locally named the Chantrey Moraine System, was deposited shortly after 8700 years B.P. and could reflect increased accumulation on the Laurentide Ice Sheet brought on by creation of a new moisture source when the sea invaded the region previously occupied by the northern part of the M'Clintock Dome.

Emergence during deglaciation exceeded 30 m per century, at least in places, and averaged more than 53 cm per century during the last 4500 years.

Résumé

Au centre de l'Arctique canadien, une surface d'aplanissement par érosion, d'étendue régionale a été fragmentée au cours de la phase tectonique eurékienne (Miocène-Pliocène) en plateaux, plaines et plaines littorales qui font actuellement l'objet d'études, et en fossés d'effondrement qui sont aujourd'hui de larges bras de mer.

Sept groupes génétiques de matériaux – roche, till et sédiments fluvioglaciaires, glaciolacustres, glaciomarins, marins et fluviaux – constituent les formations superficielles de la surface cartographiée, mais la roche et le till prédominent. L'épandage de till compte d'importantes variations latérales de faciès qui délimitent deux grosses trainées de dispersion de roches carbonatées. Un épais dépôt d'argile et de silt glaciomarins forme une large zone en travers de la partie méridionale de la région cartographiée.

Voici les aspects techniques dont il faudra peut-être tenir compte dans l'avenir: région en grande partie dépourvue de sable et de gravier; instabilité naturelle des versants qui pour la plupart sont recouverts de matériaux non consolidés; épais dépôts de till et sédiments glaciomarins à grains fins à forte teneur en glace de fond; contraction et fendillement du sol en raison des variations annuelles de température; soulèvement du sol et des couches rocheuses par le gel; poussée ascendante des glaces sur les rives marines et lacustres; et affouillement des lits et des berges de rivières par les glaces.

Les résultats d'analyses d'environ 750 échantillons de till constituent la base initiale de données géochimiques pour cette région. Les variations de la proportion courante de métaux communs traduisent les variations de la teneur en carbonate.

Tous les sédiments superficiels montrés sur la carte 1570A se sont déposés à la fin de la glaciation du Wisconsin. Au cours du maximum glaciaire du Winconsin supérieur, la nappe glaciaire Laurentide s'est divisée selon l'axe nord-sud au-dessus du chenal M'Clintock à partir d'une accumulation en forme du dôme. Le dôme M'Clintock était contigu au sud à la masse principale de la nappe glaciaire et s'est fusionné au dôme Foxe au fond du golfe de Boothia. Le dôme M'Clintock a provoqué une coulée de glace allant vers l'est et l'est-nord-est sur la presqu'île de Boothia et le nord du district de Keewatin. Deux larges langues glaciaires à la base de la nappe caractérisent entre autres cet écoulement glaciaire.

La déglaciation a commencé il y a plus du 9 250 ans et le retrait s'est d'abord effectué vers l'ouest, vers le centre du dôme M'Clintock. Ce retrait s'est graduellement orienté vers le sud à mesure qu'une baie de vêlage se formait le long du côté ouest de l'île Somerset et de la presqu'île de Boothia et pénétrait au centre du dôme. Vers 8 800 ans BP, la presqu'île de Boothia était en grande partie libre de glace et la partie submergée du dôme M'Clintock avait entièrement disparue vers 8 700. Dans l'Arctique canadien, un système morainique régional, localement connu sous le nom de système morainique Chantrey, s'est formé peu de temps après 8 700 ans BP et pourrait représenter une accumulation accrue à la surface de la nappe glaciaire Laurentide résultant de l'apparition d'une nouvelle source d'humidité lorsque la mer a envahit la région couverte auparavant par la partie septentrionale du dôme M'Clintock.

Pendant la déglaciation, l'émersion a été supérieure à 30 m par siècle en certains endroits du moins, avec une moyenne de plus de 53 cm par siècle, au cours des 4 500 dernières années.

INTRODUCTION

General

During the mid and late 1970s several teams from the Geological Survey of Canada investigated terrain conditions along a proposed pipeline route from the natural gas fields of the northwestern Arctic Islands to Ontario. One team in 1974 investigated Boothia Peninsula and northern District of Keewatin, an area of about 64 000 km^2 (Fig. 1). Their field work consisted mostly of checking and correcting airphoto interpretation by means of helicopter surveys. Final airphoto interpretations were released on Open File (Boydell et al., 1975a) and general attributes of the terrain were described (Boydell et al., 1975b).

The reconnaissance survey of 1974 was followed by field work and airphoto interpretation by the author in 1978 and 1979. The later work concentrated on reconstructing the sequence of late Quaternary glacial and sea level events and on analyzing the geology of the late Wisconsin till sheet – the most extensive Quaternary deposit in the area. Some results for the northern half of Boothia Peninsula are given in Dyke (1979a; 1980).

This final report, which is the companion of a report on the Quaternary geology of Somerset Island (Dyke, in press), describes the surface materials of the study area with the aid of Map 1570A, discusses the attributes of these materials that could be of possible future economic concern, and summarizes the record of glacial and sea level fluctuations. In the discussion of till, data from those parts of Somerset Island that were overridden by late Wisconsin Laurentide ice are incorporated in this report. In the discussion of the pattern and chronology of deglaciation (Quaternary History), data from Somerset Island and from north-central District of Keewatin are incorporated. The most important conclusion regarding the Quaternary history of the region is that, during the late Wisconsin maximum, a major ice dome (M'Clintock Dome) was centred over M'Clintock Channel, which is situated to the west of the map area and to the east of Victoria Island. In developing this conclusion, aspects of the Quaternary geology of Prince of Wales Island, King William Island, Victoria Island, and Melville Island are considered. Aspects of the Quaternary geology of the Foxe Basin region and of central District of Keewatin are discussed in an examination of the relationship of the M'Clintock Dome and its ice divide to the Foxe Dome and to the Keewatin Ice Divide. Hence, although this report is concerned principally with the map area, it should be of interest to students of the Quaternary of Arctic Canada in general and to those dealing with reconstructing the dynamics of the Laurentide Ice Sheet.

Acknowledgments

The 1974 field party consisted of A.N. Boydell (party chief), K.A. Drabinsky, and J.A. Netterville. C. Tarnocai, Soil Survey of Canada, Department of Agriculture, studied plants and soils. During the 1978 field season I was assisted by R.G. Hélie and S.J. Black and during the 1979 season by R.G. Hélie and A. Taal. Polar Continental Shelf Project, Energy, Mines and Resources Canada, provided logistical support during 1978 and 1979.

Climate

Weather records have been kept at two Inuit communities, Spence Bay and Pelly Bay, and at a former Defense Early Warning Line site, Shepherd Bay, all in the southern half of the map area. The three stations have closely similar weather patterns with long, cold dry winters and brief, cool, damp summers. The climate at Pelly Bay seems slightly more marine than that at Spence Bay or Shepherd Bay in that it receives appreciably more precipitation in both winter and summer and has slightly cooler and slightly shorter summers (Table 1). Pelly Bay has only 52 frost-free days on average each year compared to 73 days at Spence Bay (Atmospheric Enviroment Service, 1975).

Vegetation and Soils

Tarnocai and Boydell (1975) have divided the map area into two ecological regions. Simpson Lowland and the entire area lying north of a line connecting Inglis Bay and the mouth of Pelly Bay (Fig. 2) fall within the mid-Arctic region and the remainder of the area falls within the low Arctic region. The low Arctic region has continuous vegetation cover and the northern limit of ericaceous shrubs and tussock-forming **Eriophorum** is the limit of the region. The mid-Arctic region has sparser vegetation cover with little or no plant cover on highly calcareous materials. Low Arctic soils exhibit much more B-horizon development and less frost churning than do mid-Arctic soils. Low Arctic soils on fine grained marine sediments are termed brunisolic turbic cryosols and gleysolic turbic cryosols; brunisolic turbic cryosols dominate tills. Most soils in the mid-Arctic, because of lack of B-horizon development, are classed as regosolic turbic cryosols.

Topography

Boothia Peninsula and northern District of Keewatin consist of extensive plateaus, plains, and lowlands (Fig. 2). Boothia and Wager plateaus (Bostock, 1969) form the axis of the landmass. Boothia Plateau is the higher of the two and in its central part consists of low rolling bedrock hills whose

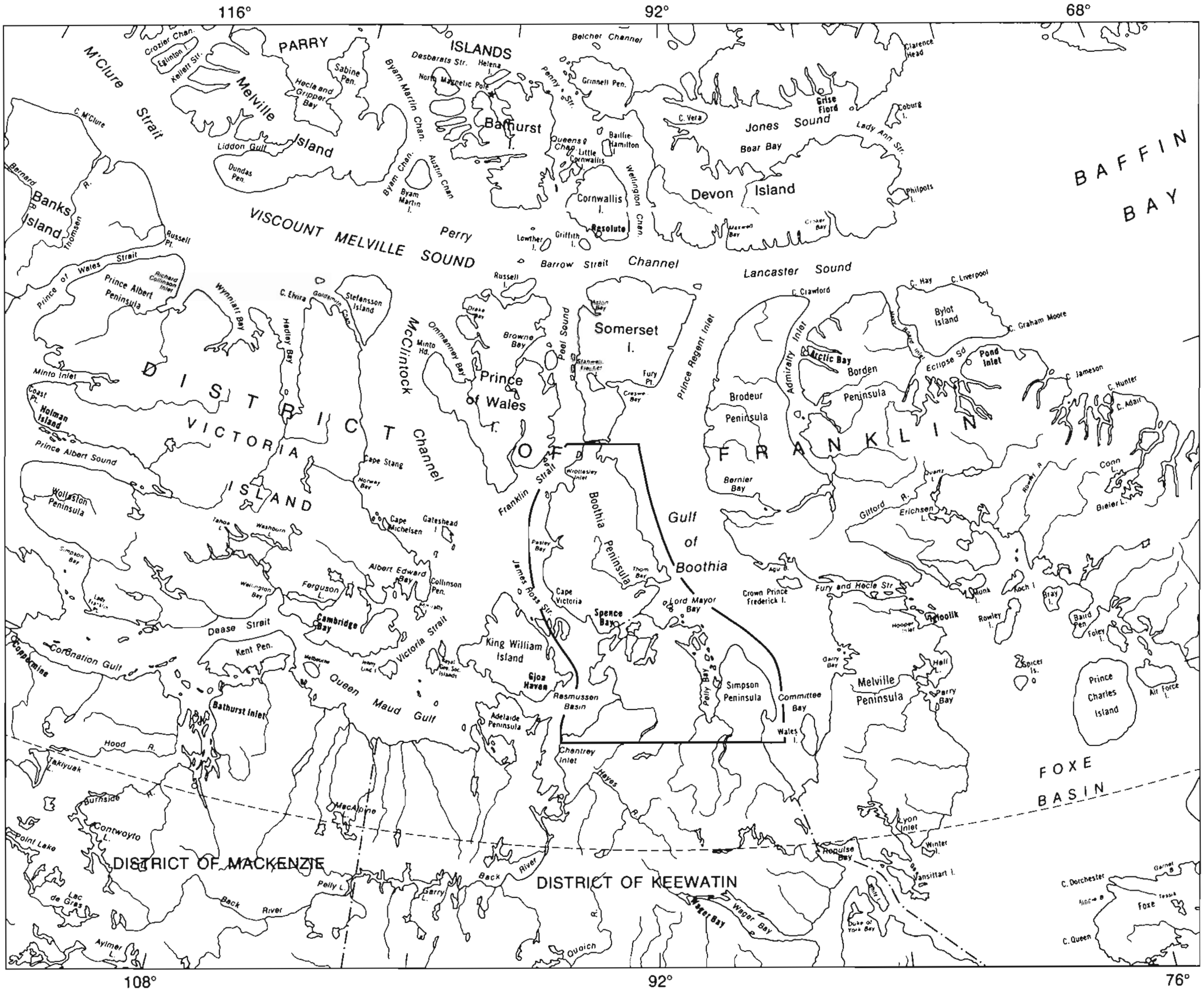

Figure 1. *Part of Arctic Canada showing location of the map area (outlined) and place names.*

Table 1. Climatic data for Spence Bay, Shepherd Bay, and Pelly Bay

	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year	Range
Average monthly and annual daily mean temperatures (°C)														
Spence Bay	-34.4	-34.6	-31.5	-21.9	-10.0	0.8	7.1	6.5	-0.9	-11.9	-23.7	-30.2	-15.4	41.7
Shepherd Bay	-35.3	-36.6	-31.5	-22.7	-10.0	0.4	8.1	6.6	-0.9	-11.7	-24.1	-30.9	-15.7	44.7
Pelly Bay	-33.1	-33.2	-28.9	-20.0	-9.7	-0.6	6.9	5.2	-2.9	-12.2	-22.6	-28.1	-14.9	40.1
Mean monthly and annual precipitation														
Spence Bay														
Rainfall (mm)	0	0	0	T	T	6.9	24.9	29.0	11.9	0.3	0	0	73.0	
Snowfall (cm)	4.6	5.6	6.1	5.6	8.1	8.4	0.3	T	10.2	19.3	7.6	5.1	80.9	
Total (mm)	4.6	5.6	6.1	5.6	8.1	15.2	25.1	29.0	21.8	19.6	7.6	5.1	153.4	
Shepherd Bay														
Rainfall (mm)	0	0	0	0	T	7.4	22.4	39.4	12.7	1.0	0	0	82.9	
Snowfall (cm)	3.3	5.3	2.0	6.1	4.8	1.8	0.3	1.3	7.1	13.0	6.4	5.6	57.0	
Total (mm)	3.3	5.3	2.0	6.1	4.8	9.1	22.4	40.4	19.8	14.0	6.4	5.6	139.2	
Pelly Bay														
Rainfall (mm)	0	0	0	0	0	5.1	33.3	49.5	15.5	1.3	0	0	104.7	
Snowfall (cm)	2.5	1.8	2.8	11.2	12.7	6.1	0.5	3.3	15.7	18.5	8.9	3.3	87.3	
Total (mm)	2.5	1.8	2.8	11.2	12.7	10.9	34.0	52.8	31.2	19.8	8.9	3.3	191.9	

Source: Atmospheric Environment Service, 1975.

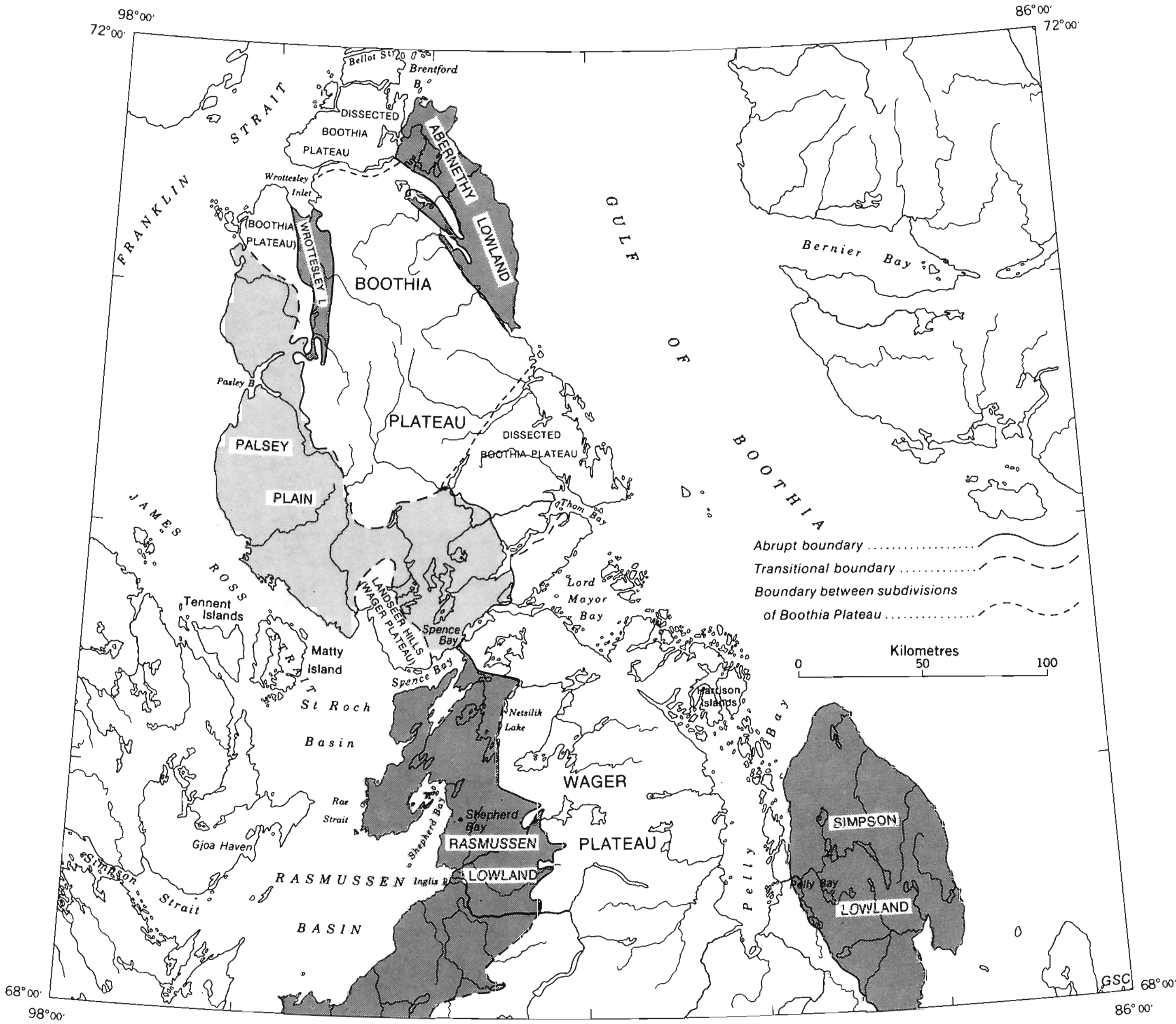

Figure 2. *General topographic regions of the map area.*

summits are 400 to 500 m a.s.l. The northern and southeastern ends of Boothia Plateau are dissected by deep, narrow intersecting valleys excavated along lines of structural weakness, which imparts a blocky appearance to the landscape. Summits in the dissected plateau lie generally between 250 and 400 m a.s.l. Wager Plateau forms the highest ground in northern District of Keewatin but is generally lower than Boothia Plateau (in the District of Franklin). Wager Plateau is more hilly in the southern part of the study area than in its northernmost extension, and summit elevations decline gradually northwards. Many summits in the south are 270 to 330 m a.s.l. whereas most summits in the north are 150 to 200 m a.s.l.

The plateaus are flanked to the east by lowlands. Abernethy Lowland occupies northeastern Boothia Peninsula and contacts Boothia Plateau along a pronounced escarpment. The lowland is submerged off the east-central part of the map area but reappears as Simpson Lowland, which contacts Wager Plateau along a major escarpment in the southeast. Abernethy Lowland has little relief; strike ridges and cuestas a few metres high and raised beach ridges are the most conspicuous landforms. Simpson Lowland is more hilly with several long cuestas and drift ridges which are 10 to 40 m high.

The plateaus are flanked to the west by Rasmussen Lowland and Pasley Plain. Rasmussen Lowland contacts Wager Plateau along a distinct escarpment, except in the southernmost part of the map area where the boundary between lowland and plateau is transitional. The lowland is mantled almost everywhere by thick glacial and marine deposits that exhibit little relief and that extend less than 60 m a.s.l. in most places (Map 1570A). Pasley Plain is an extensively drift-mantled surface of low relief whose elevation is between those of lowlands and plateaus, mostly between 100 and 180 m a.s.l. The plain abuts the plateaus at distinct escarpments east of Pasley Bay and northwest of Lord Mayor Bay but elsewhere the plateaus rise gently from it.

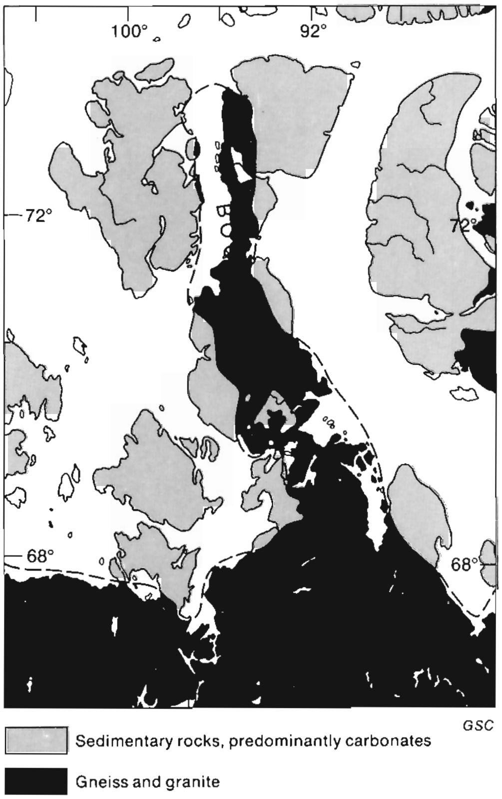

Figure 3. *Bedrock geology of the map area.*

Wrottesley Lowland is a broad scarp-bounded, drift-covered valley set into Boothia Plateau.

Bedrock Geology

The bedrock of Boothia Peninsula and northern District of Keewatin ranges in age from Precambrian to Tertiary (Fig. 3) and the distribution of rock types and their tectonic history has determined the evolution of the major topographic elements (cf. Fig. 2, 3). The oldest rocks, gneiss and granite, complexly folded and faulted, form the Boothia Horst, a structural high that underlies Boothia and Wager plateaus. The horst is flanked on both sides by lower Paleozoic sedimentary rocks, predominantly carbonates, that dip away from it. These constitute the southern part of the Cornwallis Fold Belt, the whole of which, along with Boothia Horst, comprises the Boothia Uplift (Kerr, 1977). The Boothia Uplift was active between late Precambrian and Late Devonian time.

The youngest rocks in the area are small outliers of unlithified quartz sands in Wrottesley valley and on the northeast flank of the Boothia Horst. These are the southernmost exposures of the Eureka Sound Formation in Arctic Canada (Kerr and deVries, 1977), and their preservation was ensured by downfaulting during the Miocene to Pliocene phase of the Eurekian Rifting Episode. During this time the marine channels on either side of Somerset Island and northern Boothia Peninsula as well as smaller structures such as Wrottesley valley and Bellot Strait were formed as grabens (Kerr and deVries, 1977; Kerr, 1980). Although the Tertiary tectonic history of Boothia Peninsula and northern Keewatin is less known than that of the Arctic Islands, it is reasonable to speculate that Abernethy, Simpson, and Rasmussen lowlands and Pasley Plain were dropped down from the general level of adjacent Boothia and Wager plateaus at this time as well. If so, these various discordant levels do not represent multiple stepped erosion surfaces as suggested by Bird (1967), but a single faulted erosion surface.

SURFACE MATERIALS AND LANDFORMS

The surface materials of the study area have been divided into seven genetic groups. These materials and associated glacial landforms are discussed below in order of decreasing age. Some periglacial landforms and processes, however, are common on more than one age and type of material, so these are discussed first.

The entire area lies well within the zone of continuous permafrost and the active layer in most materials is less than 1 m deep. The active layer in bedrock is somewhat deeper than in surficial deposits. At Rankin Inlet, 500 km south of the study area, the active layer in gneiss is 3 m thick (Brown, 1978). On Boothia Peninsula it is probably about 2 m thick.

Most of the surface of the study area has been modified by periglacial processes during postglacial time. Crystalline bedrock has been the least modified, but even that has been somewhat disrupted by frost heaving (Fig. 4). Carbonate bedrock has been more severely frost shattered – where not protected by a cover of surficial sediment, the upper 50 cm or so of rock has been reduced to felsenmeer. On Pasley Plain near Lord Lindsay River, where flat-lying carbonate bedrock is thinly mantled by till, vertical joints in the rock have been dilated by upturning of adjacent strata to produce large furrows similar to ice wedge troughs (Fig. 5).

The surficial deposits exhibit a variety of patterned ground forms depending on their texture and slope angle which probably control drainage and ice content. The principal patterned ground forms on till are mudboils and solifluction stripes, lobes, sheets, and steps. Till that is more than about 2 m thick has, in addition, high centre ice wedge polygons and scattered retrogressive thaw flowslides. Fine grained marine sediments (units 4b and 5b) have earth and mud hummocks and low or high centre ice wedge polygons over most of their surfaces and have thermokarst ponds and beaded streams on exceptionally flat areas. Sand and gravel deposits (units 2a, 2b, 4a, 5a, 5c, 6a, and 6b) are patterned almost everywhere by ice wedge polygons, most of which have high centres, and by turf-banded lobes on slopes, particularly on moist slopes.

Little information is available on the rates and mechanisms of periglacial processes in the study area. L. Dyke (1979, 1981) summarized observations on frost heaving of bedrock, Dyke and Zoltai (1980) measured long-term average rates of movement of mudboils, and A.S. Dyke (1981) measured long-term average rates of movement of four turf-banked lobes in various materials.

*Rock (R)**

Distribution. Bedrock forms the surface material over about 40% of the map area though most outcrop is along the eastern half of Boothia and Wager plateaus, which are composed of Precambrian gneiss and granite. There the rock is obscured only by small patches of till, mostly less than 1 m thick and largely composed of locally derived boulders. In the southernmost part of the map area, where Wager Plateau is most hilly, outcrop is discontinuous. Paleozoic sedimentary rock, which underlies extensive lowlands and plains, outcrops over only small areas. The largest exposures are on Abernethy Lowland in the northeast. Elsewhere outcrops are restricted largely to deep river cuts.

* Unit designator on Map 1570A

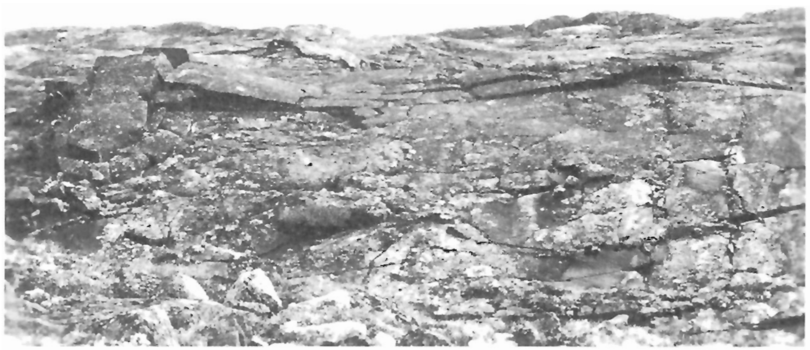

Figure 4

Typical extent of modification of granite gneiss by frost heaving of joint blocks during postglacial time. GSC-203509-L

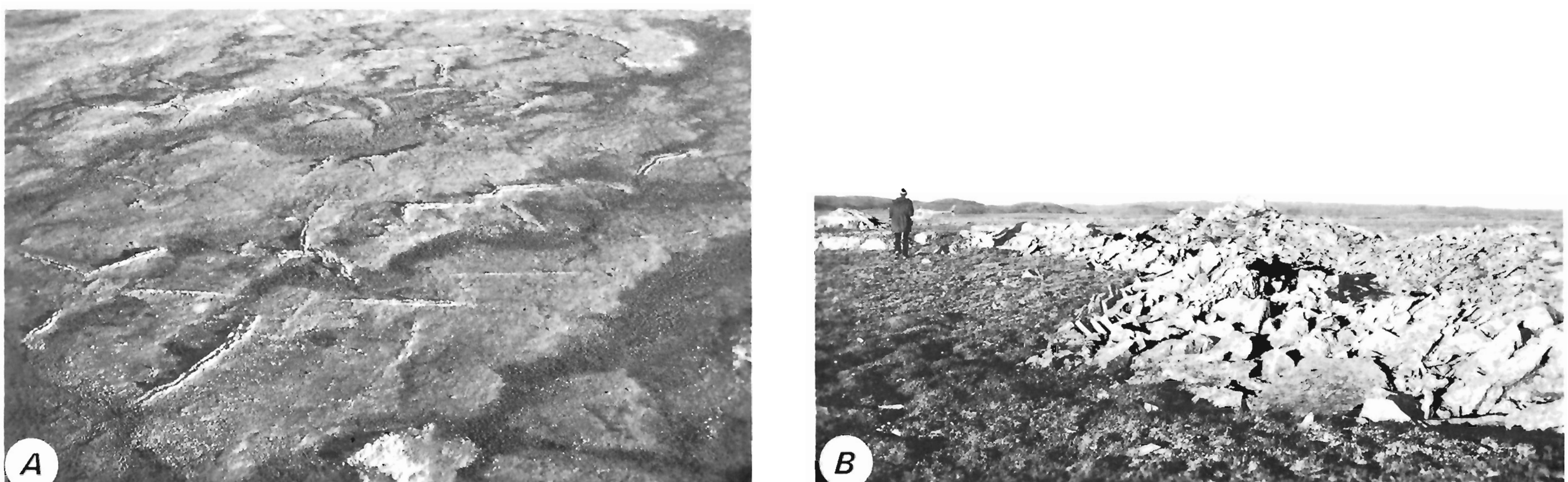

Figure 5

Vertical joints dilated by frost heave in horizontally bedded limestone near Lord Lindsay River. A – oblique aerial view (GSC-203674-G); B – ground view (GSC-203674-F)

Landforms and Ice Flow. The bedrock everywhere shows sign of glacial scour, although as Sugden (1978) noted the central part of Boothia Peninsula is less heavily scoured than other parts of the area. This is reflected most obviously in the lower density of lake basins in that area, but also by the smaller size of ice-moulded rock forms there.

In that zone of weak scour, glacial bedforms consist mainly of roches moutonnées only a few metres long. In contrast, northernmost Boothia Peninsula and southernmost Somerset Island (Dyke, 1979a, in press; Fig. 6A) are crossed by a belt of large streamlined glacial erosional bedforms in bedrock, and a similar set of landforms extends from southeastern Boothia Peninsula to northern District of Keewatin. The direction of ice flow remains nearly parallel throughout the region; flow was eastward in the north and more northeastward in the south.

Another common set of landforms in the bedrock terrain is meltwater channels. The distribution of these features is closely related to the zones of scouring referred to above and to the postglacial marine limit. The zones of intensive scouring occur mostly below marine limit and have few if any channels. The scarcity of subglacial channels in such areas could indicate that ice flow was too fast to permit the maintenance of tunnels at the base of the ice sheet. The distribution of eskers, which are largely restricted to areas that lack large ice flow features, supports this suggestion. Ice marginal channels are, of course, restricted to areas above marine limit. These are much smaller features (1-5 m deep, 100-300 m long) than the subglacial channels, the largest of which are 60 m or more deep and 15 km long. Many subglacial channels served also as proglacial meltwater routes during deglaciation.

Till (1)

Distribution and Thickness. On Map 1570A, the regional till sheet is divided into two units: blanket and veneer. Till blankets are sufficiently thick to completely obscure the structure and form (relief) of the underlying bedrock, whereas the surface of till veneers mimics the underlying rock surface. Till is both widespread and in many places thick over Pasley Plain and Rasmussen Lowland and over the adjacent western flanks of Boothia and Wager plateaus. East of the plateaus, till is widespread on Abernethy and Simpson lowlands. In most places on Abernethy Lowland, the till is thin but on Simpson Lowland it attains thickness of about 20 m.

Landforms and Ice Flow. The most common and most significant landforms of the till sheet are end moraines, ice marginal meltwater channels, and flutings.

The largest field of streamlined till forms (flutings) extends from southwestern Boothia Peninsula to northern District of Keewatin (Fig. 6A) and thus occurs generally up glacier of or among the large streamlined bedrock forms described above. Another much smaller field of streamlined till forms occurs at the base of Simpson Peninsula, and isolated features or groups of features occur elsewhere.

Of the two types of ice marginal landforms, the meltwater channels are restricted to Boothia Peninsula, because most of northern District of Keewatin lies below marine limit. End moraines composed of till[1] are of two main types. Low, short, and closely spaced ridges, which vary in texture from lines of boulders to low ridges of essentially unmodified basal till, are interpreted as De Geer moraines. They occur only on ground that was deeply submerged by the postglacial sea, and their volumes and spacing indicate that they were formed during brief, perhaps annual, stands of the ice front. Other end moraines composed of till occur both above and below marine limit but they are much more bulky features and are more laterally continuous. The largest such moraines occupy central Simpson Lowland where they form a set of north-south oriented, curving to slightly sinuous, and in places crosscutting, ridges. Some sit atop limestone cuestas indicating that these features served to stabilize the ice front for some time during deglaciation. Other large end moraines composed of till occur to the north, east, and particularly south of Pasley Bay and along the escarpment that divides Rasmussen Lowland from Wager Plateau just south of Netsilik Lake.

Composition. The till ranges from extremely calcareous to noncalcareous. The proportion of till that was derived from sedimentary (predominantly carbonate) rocks was measured in two ways: (1) the "carbonate" content of the matrix was determined by attacking subsamples of the less than 2 mm size fraction with 10% HCl and measuring weight loss; (2) the proportion of sedimentary clasts of fine gravel size (2 to 5.6 mm) was determined by visual identification and weighing. Two problems reduce the precision of the HCl method: (1) the acid apparently dissolves some noncalcareous material (e.g. iron oxides) to give some (<10%) weight loss on purely shield-derived tills and (2) a considerable proportion of the sedimentary rock is insoluble, so that even till derived entirely from limestone and dolostone could have a "carbonate" content of only 70%. The visual identification method using the fine gravel fraction eliminates these problems but is extremely tedious and time consuming.

Despite these problems, both the "carbonate" content of the matrix and the sedimentary rock content of the fine gravel fraction (Fig. 6B, 6C) reveal the same major facies changes in the till sheet. The features of most interest are two large trains of moderately to extremely calcareous till that occur in the northern and southern parts of the study area[2]. The northern train is 150 km wide and the southern one, 220 km wide; the two are separated by an area of relatively weak dispersal that is 110 km wide. In the area of weak dispersal, which corresponds to the area of less strongly scoured bedrock mentioned above, the sedimentary rock component of the till declines to zero 10 to 50 km down ice from the western edge of the crystalline bedrock. It increases again to more than 90% about 10 to 15 km beyond the eastern edge of the shield. In the areas of strong dispersal, which correspond to the areas where large bedforms in both bedrock and till occur, highly calcareous till has been spread across the entire width of the crystalline rocks and the sedimentary component remains as high as 70% more than 100 km from the source area. Both trains, but especially the southern one, have sharply defined edges and both have midstream plumes marked by belts of farthest transport. The midstream plume of the southern train is particularly spectacular. The plumes correspond to areas of slight convergence apparent in the pattern of ice flow features (Fig. 6A). The edges of the dispersal trains do not coincide with any topographic features that could have served to confine an ice stream.

Grain Size. A till sheet with such extreme lateral variations in chemical and petrological composition would be expected to portray correlative textural variations. The texture of the till is portrayed by contour maps of the sand and silt content of the matrix (Fig. 6D, 6E). The sand content ranges from less than 20% to more than 90% and there is an obvious negative correlation between sand content and carbonate content (Fig. 6D, 6B). The silt content ranges from less than 10% to more than 50% and has an obvious positive correlation with carbonate content (Fig. 6E, 6B). The clay content has a somewhat narrower range and shows no obvious consistent

[1] Other end moraines composed of stratified drift are described below.

[2] The parts of Somerset Island that were inundated by Laurentide ice during the last advance are included in this discussion of tills. For further discussion of the Quaternary geology of Somerset Island, see Dyke (in press).

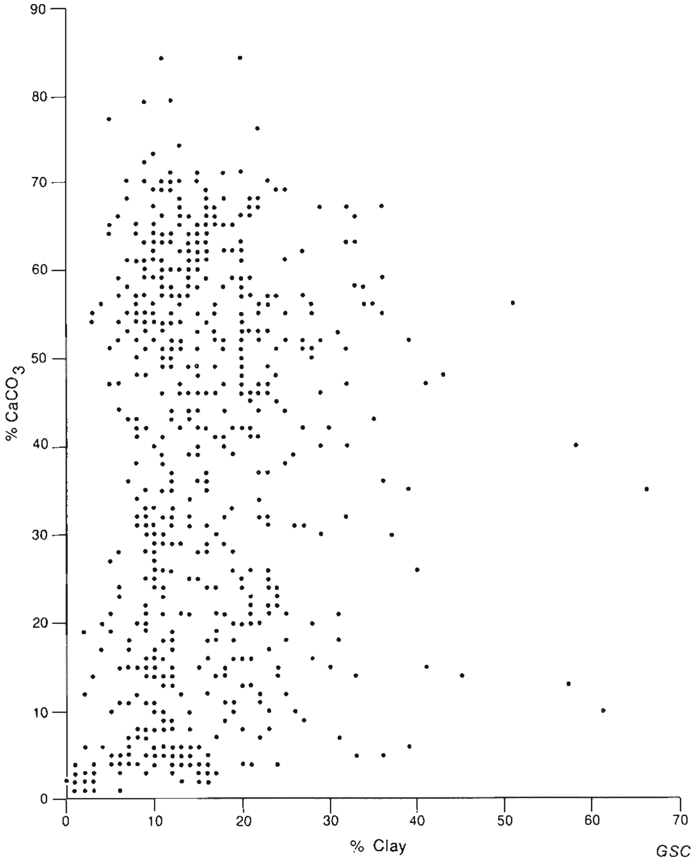

Figure 7. *Plot showing the lack of correlation between clay content and $CaCO_3$ content of the till matrix.*

correlation with carbonate content (Fig. 7). The strong positive correlation between carbonate content and silt content and the lack of correlation between carbonate content and clay content suggest that most limestone and dolostone erratics are crushed and ground to silt size and that little material is comminuted farther to clay size. This agrees with the terminal modes for calcite and dolomite suggested by Dreimanis and Vagners (1972) for tills from southern Canada. The negative correlation between sand content and carbonate content can be restated as a positive correlation of sand content and crystalline rock content. Hence, crystalline rocks are crushed to sand size with less than 30% of material being comminuted to silt size. Clay is equally difficult to produce from either carbonate or crystalline rock.

Given the correlation between grain size distribution and composition, it is not surprising that the large dispersal trains are recognizable in the texture maps (Fig. 6D, 6E). Where sample density is sufficient, smaller features such as the midstream plumes are also recognizable. These are displayed better on the map of sand content than on the map of silt content, indicating that the coarser fraction better reflects the lateral facies changes. On neither map is the southern edge of the southern dispersal train recognizable; this indicates that a greater sample density would be necessary in order to recognize these trains on a purely textural basis and that texture of the till is controlled by factors other than just bedrock type.

Glaciofluvial Sediments (2) and Landforms

Glaciofluvial sediments have been divided into two groups: those deposited in contact with the ice sheet and those deposited in front of the ice sheet by proglacial meltwater.

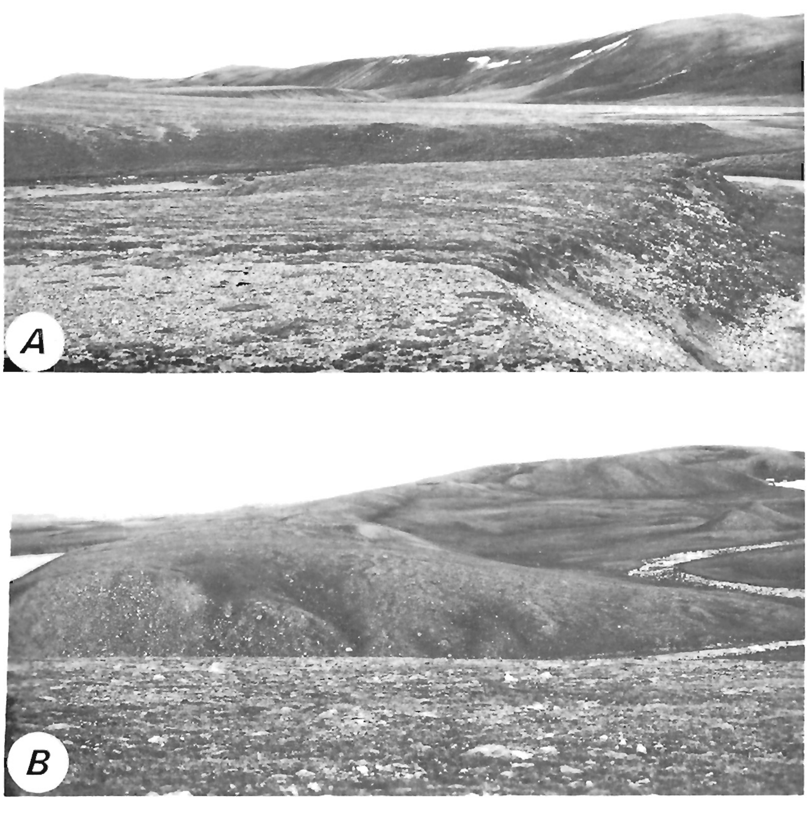

Figure 8

Large end moraine near Lord Lindsay River composed of stratified drift. A – 60 m-high distal face of moraine forming skyline (GSC-203508-W); B – 5 m-high proximal face of moraine with moulin kames behind it (GSC-203508-V)

By far the greater bulk of sediment is of ice contact origin. The largest concentration of ice contact deposits is the kames, kame morraines, and crevasse fillings that occur on Pasley Plain and on the adjacent flank of Boothia Plateau where the deposits abut or overlie thick, highly calcareous till. Northwest of Josephene Bay, several east-west to northeast-southwest oriented gravel ridges have distinct ice contact faces on their southerly sides and are, therefore, interpreted as end moraines composed of stratified drift. The most spectacular belt of ice contact stratified drift extends southwestward from Sanagak Lake. The southeast margin of the belt consists of an end moraine with a 60 m-high distal face and a much lower, scalloped, proximal face (Fig. 8). Beyond the moraine, to the southeast, is an extensive till plain; behind it, to the northwest, is a 5 km-wide belt of kames and crevasse fillings with intermediate kettles. Another end moraine of stratified drift worthy of note occurs at the eastern end of Amituryouak Lake where it formed at the snout of a large valley glacier during deglaciation. Eskers constitute the final category of ice contact sediments. Only a few short, low eskers occur on Boothia Peninsula, notably on Abernethy Lowland and on southwestern Pasley Plain, but they are more common and more continuous in northern District of Keewatin where two of them are more than 40 km long.

Proglacial outwash is limited to areas above the postglacial marine limit, and hence mostly to Boothia Peninsula. Many small deposits, commonly bouldery, are associated with individual meltwater channels and are not shown on Map 1570A. The largest deposits, mostly composed of gravel, developed at the mouths of numerous lateral meltwater channels that formed as the ice margin retreated downslope on the east side of upper Pasley River valley.

Glaciolacustrine Sediments (3) and Landforms

During deglaciation, as the ice front receded downslope on the western side of the height of land on Boothia Peninsula, some 20 or so small ice dammed lakes were formed. The former extent of these lakes is marked best by weakly developed strandline features. Most of the lakes must have been short-lived because very little sediment accumulated in them. Below the strandlines that mark their limits, however, the surfaces that were flooded (mostly till) support a more luxurious vegetation. This probably indicates that a small amount of lacustrine silt and clay has been admixed with the till in the active layer and has increased its moisture retaining capacity. The areas are mapped as till rather than as glaciolacustrine sediments, but the strandline features are shown. The only areas mapped as glaciolaustrine sediments are 11 small basins on Pasley Plain, where the sediments form small flats and are probably no more than 5 m thick.

Glaciomarine Sediments (4) and Landforms

Two types of marine sediments were deposited as the ice margin receded across the area. Deltaic sediments, mostly sand and gravel, prograded as much as 10 km seaward of the marine limit, are the most unequivocal indicators of the position of marine limit along the east coast of Boothia Peninsula and throughout much of northern District of Keewatin. They occur throughout the map area along the limit of submergence and are transitional between the proglacial outwash (unit 2b) described above and the glaciomarine silts (unit 4b) described below. In two places in the southernmost part of the field area, one east and one west of Murchison Lake, these deltas have distinct ice contact escarpments at their proximal ends. In most other places they occur at the downstream ends of meltwater channels, such as those just west of Abernethy River, or at the downstream ends of eskers, such as those west of Murchison Lake.

Unlike the glaciomarine deltas, the glaciomarine silts are mostly restricted to a broad belt along the southern 20% of the map area and extend south of the map area another 30 km or so farther into Keewatin (Thomas and Dyke, 1979a, b). Judging by the erosional relief that has developed on these deposits since emergence, they probably reach thicknesses of 100 m. In places, and in particular on Rasmussen Lowland, the glaciomarine silts overlie thick till and adjacent to the larger stream courses they are overlain by veneers of deltaic or beach sand that are too small to show on Map 1570A.

The silts south and west of Pelly Bay have been severely gullied since emergence, resulting in extensive and spectacular badlands. The lack of a vegetation cover in the badlands indicates that this is an ongoing process. On Rasmussen Lowland, perhaps because the regional slope is less, the silts have been much less dissected. There the main postemergence modification has been the development of ice-wedge polygons, thermokarst lakes, and beaded drainage.

Marine Sediments (5) and Landforms

Three types of marine sediment were deposited after the ice front had passed through the area and the plume of fine glaciomarine sediment had settled to the bottom. These were deposited in deltaic, nearshore, and beach environments.

Deltaic deposits that are extensive enough to show on Map 1570A are not common, which attests to the small amount of sediment carried seaward during postglacial time. The largest deltaic accumulations occur along Kellet River, which has cut through glaciomarine silts. Smaller deltas occur mainly in the vicinity of Pasley Bay where streams are eroding thick deposits of fine grained till. The postglacial deltas contain much less coarse grained material than do the glaciomarine deltas. Boulder gravel topsets, which are typical of the latter, are replaced by sand and fine gravel topsets of the former.

Thin patches of nearshore silts occur in association with raised beaches throughout the map area. They probably originated by suspension of fine sediment in the littoral and immediate prolittoral zone due to wave agitation of till followed by resettling a short distance offshore. Possibly though, some of these thin silts are actually glaciomarine sediments. Glaciomarine silts could be distinguished from postglacial (nonglacial) marine silts by radiocarbon dating of enclosed faunas, but such an extensive effort is not warranted by the minor importance of the problem.

Most large, continuous flights of raised beaches are restricted to the lower third of the area of postglacial submergence, where regression of the sea occurred more slowly than at higher elevations[1], but in places beaches extend to the marine limit. The beaches are underlain almost everywhere by highly calcareous till and, hence, are composed mostly of subangular carbonate rock fragments forming gravel about 1 m thick. The thickness of the beach gravel is limited because, as the gravel developed, it armoured the underlying till, thus protecting it from further wave action. Longshore transport of gravel, which would have allowed accumulation of thick deposits of beach material, was limited as indicated by a lack of raised bars and spits. Elsewhere beaches were formed by wave reworking of eskers, particularly in the southwestern part of the map area. These beaches, compared to those that overlie till, are more sandy and clasts are better rounded.

[1] See discussion of postglacial emergence at the end of the Quaternary History section.

Fluvial Sediments (6) and Landforms

Fluvial sediments are divided into two types: those that lie within the present flood zone and those that are perched above it. Both types commonly occur together though neither is common. As with postglacial deltaic sediments, the most extensive fluvial sediments occur along Kellett River, where their source is eroded glaciomarine silt and fine sand. Other deposits, large enough to show on Map 1570A, occur along Arrowsmith, Becher, Murchison, Inglis, Lord Lindsay, Pasley, and Wrottesley rivers. The small amount of fluvial sediment results from the shortness of the stream flow season, low water levels during much of the season, permafrost beneath the river beds and banks, and the many sediment traps that exist along the poorly graded stream courses.

ECONOMIC GEOLOGY

This chapter is concerned (1) with those aspects of the surficial geology of the map area that might influence or pose problems for future land use activities and (2) with the use of till – the most widespread Quaternary sediment – as a mineral prospecting medium. For some land use concerns or planning, the surficial geology maps at 1:125 000 scale (Boydell et al., 1975a) may be more useful than the 1:500 000 scale map accompanying this report because the larger scale map identifies smaller units and contains information on relief, slope, and periglacial modification. Attributes of the terrain that are deemed to be of most importance to land use planning are the availability and quality of granular resources, Atterberg limits of sediment, ground ice type and distribution, slope stability, and trafficability. A great deal more data are available on some of these topics than on others, and certain baseline information on permafrost thickness and temperature profiles are non-existent.

Land Use Concerns

Granular Resources

The abundance and composition of sand and gravel vary widely throughout the study area. Large volumes of both materials are stored in the ice contact stratified drift deposits of eastern and central Pasley Plain. These materials are highly calcareous. On Boothia and Wager plateaus, granular materials are scarce; the most important deposits there are glaciomarine deltas, eskers, and kames, all of which are small and widely scattered. Most clasts in these deposits are hard crystalline rocks. On Rasmussen Lowland the best sources of granular material are the two large eskers near its southern end; these features are composed predominantly of shield-derived material.

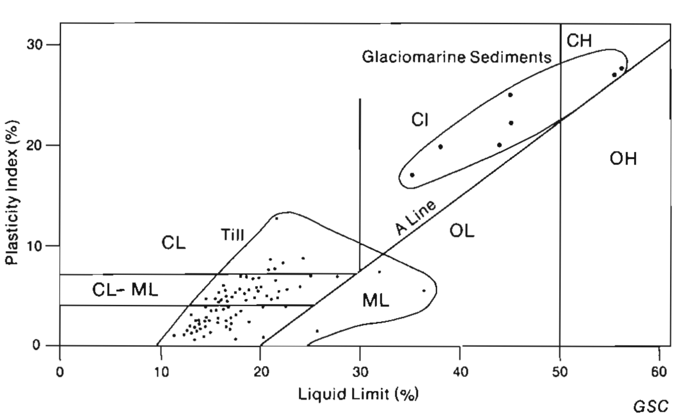

Figure 9. *Classification of till matrix and glaciomarine sediments according to the unified Soil Classification System.*

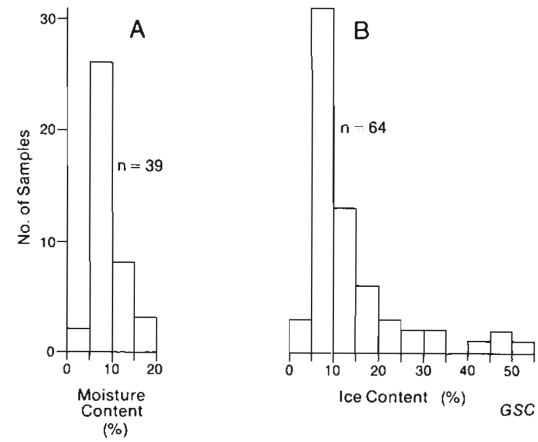

Figure 10. *Moisture content of the unfrozen active layer (A) and of the upper metre of permafrost (B) in till, both expressed as per cent weight of dry sediment.*

The most widespread gravels are those of the raised beaches, but because the deposits are so thin, broad areas would have to be scraped if large supplies were needed. These gravels are composed mostly of subangular carbonate fragments with little or no matrix.

Projects in this region that will require large supplies of granular material will have to resort either to long distance haulage or to extensive rock crushing. A possible undesirable effect of such a project might be depletion of granular resources in parts of the area.

Atterberg Limits

Atterberg limit tests were run on representative samples of till matrix (Fig. 9). Despite the wide variation in texture of till (Fig. 6D, 6E), all samples plot as ML, CL-ML, or CL soils – silty sands, silts, and clay silts – of slight to low plasticity. The narrow range of plasticity and liquid limits is probably a function of the narrow range over which the clay content of the regional till sheet varies.

Seven samples collected from boreholes in glaciomarine sediments in the southern part of the map area tested as silty clay of medium to high plasticity (CI and CH soils). These sediments probably are more variable, both vertically and laterally, than the few samples might indicate.

Ground Ice and Soil Moisture Content

The moisture content of the active layer and ground ice content of the uppermost metre of permafrost were measured on samples collected from 15 pits dug in till. The active layer samples were collected during a late, wet spring (1978), but moisture contents still did not exceed 20% by weight even on imperfectly drained sites (Fig. 10). Most samples had 5 to 10% moisture by weight. The uppermost metre of permafrost shows more variability in moisture content than does the unfrozen active layer (Fig. 10) but most samples cluster in the 5 to 10% range. Most pits showed little or no difference between the moisture content of the thawed and frozen layers but a few showed increasing ice content with depth. Most visible ice exposed in the pits occurred as tiny lenses and as coatings around stones. Extensive ice-wedge polygons and scattered retrogressive thaw flowslides indicate that massive ice is also present just below the permafrost table in till.

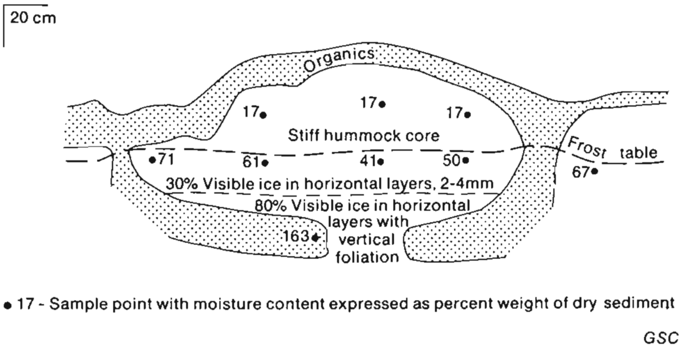

Figure 11. *Profile of earth hummock and underlying sediment showing moisture contents and notes on ground ice.*

Moisture and ground ice contents of marine silt and clay were investigated in one pit and two boreholes. The pit wall (Fig. 11) bisected an earth hummock and revealed a stiff hummock core, having 17% moisture content, over a slightly bowl-shaped frost table. The uppermost frozen layer contained about 30% visible ice by volume distributed in numerous 2 to 4 mm-thick horizontal layers. Samples of this layer yielded moisture contents by weight of 41 to 71% with the drier material occurring under the hummock crest. Visible ice content increased with depth to about 80% by volume. Earth hummocks are exceedingly common on marine silts in the map area, as they are on soils of similar texture in the Western Arctic (Mackay and MacKay, 1976; Tarnocai and Zoltai, 1978), and the single example presented here is probably not atypical. The two boreholes[1] in marine silt and clay extended to 5.5 and 8 m and showed ground ice contents ranging from 25 to 115% by weight (but mostly in the 25 to 50% range) and from 10 to 60% visible ice by volume. Most visible ice was in the form of randomly oriented ice lenses.

Slope Stability and Natural Hazards

Unconsolidated material on slopes in the Arctic are naturally unstable in some measure as indicated by the widespread occurrence of solifluction and frost creep features. The long-term rate of advance of these features seems to be very slow, a fraction of a centimetre per year (A.S. Dyke, 1981), and movement is confined to the active layer. Artificial structures that are perched above the surface would be immune to this movement as long as they were anchored in permafrost. Structures that obstruct a slope movement, however, should be built to withstand constantly decreasing support on the downslope side. If moisture supply to a slope is increased, or if soil pore water pressure is increased, the solifluction rate will increase perhaps by an order of magnitude or more. Solifluction rates under natural conditions provide only minimum measures of slope instability likely to be encountered in many land use situations.

Another important indicator of natural slope instability is the retrogressive thaw flowslide. Several inactive features of this type, which indicate rapid failure of soil underlain by massive ice, were noticed on the thick tills, but they do not seem to be as common here as they are on Somerset Island to the north (Dyke, in press).

Natural slope instability is widespread on marine silts, as indicated by the extensive badlands that have developed on these sediments. These areas should be avoided or their natural instability should be taken into consideration in planning land use activities. Large areas of marine silt that form flat to gently sloping terrain have not been eroded to badlands. However, the high ground ice content of these materials, as indicated by thermokarst ponds, beaded streams, and extensive ice-wedge polygons, would likely cause instability in the vicinity of excavations. In addition, lateral migration of thermokarst ponds and streams, which occurs naturally as ice in the uppermost permafrost is melted by the water in the ponds and streams, could weaken or remove the support for artificial structures. Armouring of the eroding banks with gravel might prevent further migration.

Heaving of bedrock blocks by frost heave in the map area and other parts of Arctic Canada has been dealt with by L. Dyke (1979, 1981). He found that blocks at the surface can move several centimetres in a single year and that displacement is largest on the flanks of outcrops that are surrounded by saturated till. Naturally occurring frost heave in soil in the map area has not been monitored. It is undoubtedly occurring, however, wherever segregated ice is actively forming such as on the floors of recently drained thermokast ponds, in fields of actively widening ice wedges, and in actively aggrading colluvium on seepage slopes. Frost heaving in soil would likely be triggered by any artificial processes or structures that impeded slope drainage or that significantly lowered ground temperatures and thus caused migration of unfrozen soil moisture* towards the colder ground.

Rafting of sea and lake ice is active during much or all of the summer. This process, which involves rafting of ice against the shore by wind, results in considerable pressure upon natural and artificial features at the shoreline, in the shallow foreshore, and in gently sloping backshore areas. Rafting annually ploughs up much sediment into ramparts, which in many places consist of boulders. These are constantly moved around as rafting continues. The effects of ice rafting are particularly evident around shores of large lakes, such as Netsilik Lake, that retain much of their ice covers throughout the summer. The effects of ice rafting in coastal areas of the central Arctic are even more severe and have been described by Taylor (1978).

Ice scouring of river beds and banks occurs throughout the Arctic and its intensity is proportional to river size, velocity, and rapidity of breakup. No observations were made of the process or effects of scouring by river ice in the map area, but along Hayes River, just south of the map area, long grooves in the coarse gravel bed, exposed at low water stage, indicate that this is an important process of erosion and transport.

Trafficability

The initial stages of economic development in the Arctic will likely involve considerable offroad vehicle movement. Hence, aspects of the terrain that influence its trafficability need to be considered; the most important aspects are relief, surface roughness problems, and traction problems.

The general accessibility of a terrain to offroad vehicles is governed by relief. In this regard, the northern and southeastern parts of Boothia Plateau, which are deeply dissected by fiord-like valleys with large lakes or arms of the sea, are the least accessible areas. On a larger scale, the badlands south of Pelly Bay present very steep gradients.

The most widespread surface roughness features that could pose problems to some types of offroad vehicles are boulders, ice-wedge troughs, earth hummocks, and sharply domed mudboils. Concentrations of boulders make passage across small areas of ice contact and proglacial deposits (units 2a, 2b) difficult, but they pose a much more serious problem in areas of extremely bouldery till. Till with boulder cover of 50% or more is widespread on central Boothia

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[1] Boreholes by F.M. Nixon, 1974.

* Considerable proportions of soil moisture, even in permafrost, can be unfrozen in fine grained soils.

Plateau; its distribution coincides with areas where the till matrix is more than 60% sand (Fig. 6D). Ice-wedge polygons pattern most sand and gravel deposits; the troughs are generally less than 50 cm deep and less than 100 cm wide. Steeply domed earth hummocks are widespread and closely spaced on silt and clay deposits. Most mudboils have little relief, but fields of steeply domed features, comparable to earth hummocks, occur on fine textured till.

Traction problems likely will be encountered on most till surfaces, particularly on lower slopes and in depressions, during spring frost table lowering and snowmelt at which time the till is saturated and highly prone to liquefaction. In mid and late summer the active layer in till is generally drier and better drained and hence has greatly increased bearing capacity. Flat-lying areas of marine silt, such as cover much of Rasmussen Lowland, are poorly drained and covered largely by thin peat and standing water throughout the summer when bearing capacity will remain low. Areas that pose traction problems are probably those that will experience most damage from offroad vehicle movement.

Base Metal Content of Till

All till samples collected from the study area were analyzed for concentrations of several base metals and uranium in the less than 2 μm size fraction. Shilts and Klassen (1976) reported on the uranium content of samples collected in 1974 and showed a broad area of relatively high uranium concentration – 5 to 12 ppm – in the south-central part of the map area. These concentrations could indicate elevated background levels in that area; background levels elsewhere in the map area are less than 2 ppm. Dyke (1980) presented contour maps of concentrations of copper, lead, zinc, cobalt, nickel, chromium, uranium, molybdenum, manganese, and iron in till on the northern two-thirds of Boothia Peninsula and discussed areas where concentrations of several metals considerably exceeded background concentrations. Dyke (in press) discussed the concentrations of the same suite of metals in till and weathered bedrock on Somerset Island, established background levels, and discussed all samples with concentrations in excess of background levels.

All samples showing concentrations in excess of background levels have been discussed in the references cited above. What remains is to examine broad regional variations in background and to explain these variations. To this end the concentrations of copper, zinc, nickel, and cobalt are contoured in Figure 12.

The cause of the variations in background levels in these metals in the regional till sheet is obvious upon comparison of Figure 12 and Figures 6B and C. Areas with low background levels are those with highly calcareous tills (regardless of underlying bedrock composition), and areas with higher background levels are those with tills composed almost entirely or entirely of crushed granite and gneiss. The correspondence is so close that the two major carbonate dispersal trains, shown in Figures 6B and 6C, can be picked out quite accurately in the base metal maps. The map of zinc concentration reflects the carbonate content closest of all – the edges of the large carbonate dispersal trains correspond to the 150 ppm contours and the central plume of the southern dispersal train is picked out by an area with less than 50 ppm zinc.

These data constitute the initial geochemical data base for this large region. Obviously, its proper interpretation requires an understanding of the late Quaternary history of the region and in particular recognition of the facies changes in the late Wisconsin till sheet. Future prospecting or geochemical studies, especially those utilizing soils, lakes, and streams as sampling media, should benefit from this finding.

QUATERNARY HISTORY

All the sediments and features shown on Map 1570A were deposited and formed during the last episode of glaciation to affect the area and during postglacial time. Sediments deposited during earlier glacial and nonglacial episodes are exposed in river bluffs in Pasley River valley on western Boothia Peninsula and in a small part of Abernethy Lowland where it forms a re-entrant in Boothia Plateau. The Quaternary history discussed here deals only with late Wisconsin and postglacial events. The abundance of sediments and clarity of features dating from those times, as well as considerable geochronologic control, allow detailed reconstructions to be made.

Data gathered during six years of mapping on Boothia Peninsula and northern District of Keewatin (Map 1570A), in north-central Keewatin (Thomas and Dyke, 1979a, b), and on Somerset Island to the north (Dyke, in press) allow, for the first time, a factually based reconstruction of the configuration and dynamics of the northern part of the Laurentide Ice Sheet. This reconstruction differs radically from earlier speculative models (e.g., Hughes et al., 1977) but is in accordance with the concept of a multi-domed Laurentide Ice Sheet that has arisen from recent work in the southern Arctic (Shilts et al., 1979; Shilts, 1980) and in the eastern Arctic (Andrews and Miller, 1979). Before discussing reconstruction from the study areas, then, it might be useful if the evolution of ideas on the structure of the Laurentide Ice Sheet was summarized briefly.

Models of the Laurentide Ice Sheet

The popular concept of the configuration of the Laurentide Ice Sheet has alternated between (1) an equilibrium ice sheet with a single, central dome over the Hudson Bay area, which generated radial flow to its margin, and (2) a non-equilibrium ice sheet with several domes, each generating its own flow pattern. Dawson (1886, 1890) and Chamberlin (1894) realized that most glacial deposits and landforms east of the Cordillera had been produced by a single ice sheet centred on the Laurentian Shield (now called Canadian Shield) and named the ice sheet after the shield. Later, Tyrrell (1898a,b, 1913) suggested that the ice sheet had had three distinct areas, or centres, of dispersal; these became known as the Keewatin, Labradorean, and Patrician ice sheets. Flint (1943) argued against this concept and for a single dome over Hudson Bay. Flint's concept has remained popular since then and has been repeated by several people (Andrews, 1973; Ives et al., 1975; Hughes et al., 1977), even though Flint himself (1971) expressed some doubt. Following Mackay (1965), Flint suggested that flow patterns shown on the Glacial Map of Canada (Prest et al., 1968), which obviously relate to dispersal areas over Keewatin and Labrador, might indicate the form of the ice sheet at its maximum. Others, however, contended that these patterns were "clearly" the product of dispersal centres that became established during deglaciation (Ives et al., 1975).

Most recently, systematic analysis of the late Wisconsin till sheet in central and southern District of Keewatin (Shilts et al., 1979), the dispersal of distinctive erratics in areas peripheral to southern Hudson Bay (Shilts, 1980), and the distribution of limestone erratics along eastern Baffin Island (Andrews and Miller, 1979) have favoured the multi-domed ice sheet model with major dispersal areas in Keewatin, Labrador-Ungava, and Baffin Island. Further support for this model can be found in the delevelling of raised marine shorelines and proglacial lake shorelines (Andrews and Barnett, 1972; Dyke, 1974; Vincent and Hardy, 1979; Gray et al., 1980).

With respect to the Boothia Peninsula region, there would be little difference between the effects of a dispersal centre over Keewatin or over Hudson Bay; either would have generated a northward flow of ice with vigorous streams flowing down the major marine channels as envisaged by Hughes et al. (1977). This is the popular concept of the last glaciation of this region, and it is against that that the data set forth below should be evaluated.

Last Glacial Advance and Maximum

Ice Flow Pattern

Ice flow features were mentioned above in the description of surface materials and landforms and are summarized in Figure 6A. Obviously, no evidence of northward flow of ice over the study area, as postulated by several of the papers dealing with the character of the Laurentide Ice Sheet (eg. Hughes et al., 1977), has been found. Instead, the numerous ice flow features show a remarkably uniform eastward flow across the area extending from northwestern Somerset Island to northern Boothia Peninsula. This swings to an equally uniform east-northeastward to northeastward flow over the rest of Boothia Peninsula and northern District of Keewatin. The flow pattern is at right angles to the axes of the marine channels and of the landmass; hence, it defies the gross topographic grain of the region.

Over large areas flow lines are mostly parallel, but slight convergences and divergences of flow can be recognized (Fig. 6A). A zone of convergence over northern Boothia Peninsula and southern Somerset Island coincides with the narrowest part of the landmass and with an area where the plateau surface is slightly lower (by 50 to 100 m) than the surface to both the north and south. A zone of slight convergence north of Spence Bay and of divergence near Shepherd Bay had a common origin: flow across the lowland south of Spence Bay was deflected to a more northerly course by the escarpment that marks the boundary between Rasmussen Lowland and Wager Plateau and between crystalline shield and sedimentary rocks. Once the ice overtopped the escarpment, flow in adjacent areas forced it back on its original course.

Bedforms

Apart from striae, the bedforms shown in Figure 6A are features large enough to recognize on 1:60 000 scale airphotos; most range in length from 500 m to 2 km. Striae occur on bedrock throughout the region, but the larger bedforms are concentrated in two main zones. A zone of strongly ice-moulded bedrock dominates the terrain of northern Boothia Peninsula and southern Somerset Island, one of the two zones of ice flow convergence described above. The second major concentration of large bedforms extends from southernmost Boothia Peninsula to northern District of Keewatin and includes the second zone of convergence described above. Streamlined till forms dominate the western and central part of this field and strongly ice-moulded bedrock forms dominate elsewhere. The largest of the minor concentrations of bedforms, that inland from Abernethy Bay, occurs on an area of relatively steeper regional slope on the lee side of Boothia Peninsula.

Dispersal Trains

As discussed above in the description of facies changes in the regional till sheet and as shown in Figures 6B and 6C, two large trains of far-travelled calcareous till occur in the two zones of strongly developed bedforms. Only a minimum estimate of the distance of transport of material in these streams can be given because they continue eastward beneath the Gulf of Boothia. However, as much as 70% of the till on southeastern Boothia Peninsula (in the southern dispersal train) came from a distance of 100 km or more. The actual distance of transport of material in the trains could be several hundred kilometres.

Interpretation - M'Clintock Dome

The consistent eastward to northeastward ice flow direction indicated by the large bedforms and by the dispersal trains requires that a major area of ice dispersal existed to the west of the study area during the height of the last glacial episode. Because ice flow direction was nearly parallel throughout the region, because ice flowed across the region everywhere for about the same length of time, and because calcareous material was equally available to the ice sheet in all places west of Boothia Horst, the two large carbonate dispersal trains must have been formed by ice streams at the base of the ice sheet that flowed much faster than the ice over the areas of relatively weak dispersal. So, in summary, a large glaciodynamic feature located west of the study area generated not only an eastward to northeastward flow pattern, which defied the gross topographic grain of the region, but also two large ice streams (Fig. 13).

The Glacial Map of Canada (Prest et al., 1968) summarizes what is known of ice flow patterns on islands west of the study area based on mapping by Fyles (1963) and Craig (1964). Prince of Wales Island and King William Island both lie up ice from the study area. On both islands the dominant flow patterns are northward to northwestward, but small fields of eastward to northeastward oriented features also occur there; I suggest that these latter features represent the flow directions during the height of the last glacial episode and that the northward to northwestward oriented forms were produced later during phases of deglaciation.

A complex mosaic of ice flow patterns occurs on Victoria Island (Fyles, 1963), but on the eastern and southern parts of the island the oldest flow features, judging by crosscutting relationships, are oriented westward. Possibly these features represent the flow direction during the last glacial maximum and the crosscutting patterns were produced during various phases of deglaciation. The oldest flow features require a major area of ice dispersal to the east of Victoria Island.

If the westward flow on Victoria Island occurred during the last glacial maximum, then it was coeval with the eastward flow on Somerset Island and Boothia Peninsula. Thus, the two flow patterns define an ice divide, more than 700 km long, located over M'Clintock Channel (Fig. 13). It is proposed here that this be named the *M'Clintock Ice Divide* and that the ice mass of which it formed the central part be named the *M'Clintock Dome* of the Laurentide Ice Sheet.

A zone of southwestward oriented flow features in northeastern District of Mackenzie, which has not been explained previously, probably describes flow beneath the southwestern slope of the M'Clintock Dome (Fig. 13).

Margin of M'Clintock Dome

The margin of the northwestern part of the late Wisconsin Laurentide Ice Sheet (M'Clintock Dome) has been mapped where it terminated on land. On its eastern side, the M'Clintock Dome terminated on land only on Somerset Island (Fig. 13). Most of Somerset Island that lay beyond the limit of the M'Clintock Dome was covered by a radially flowing local ice cap, which coalesced with the main ice sheet in places and was separated from it by nunataks in other places

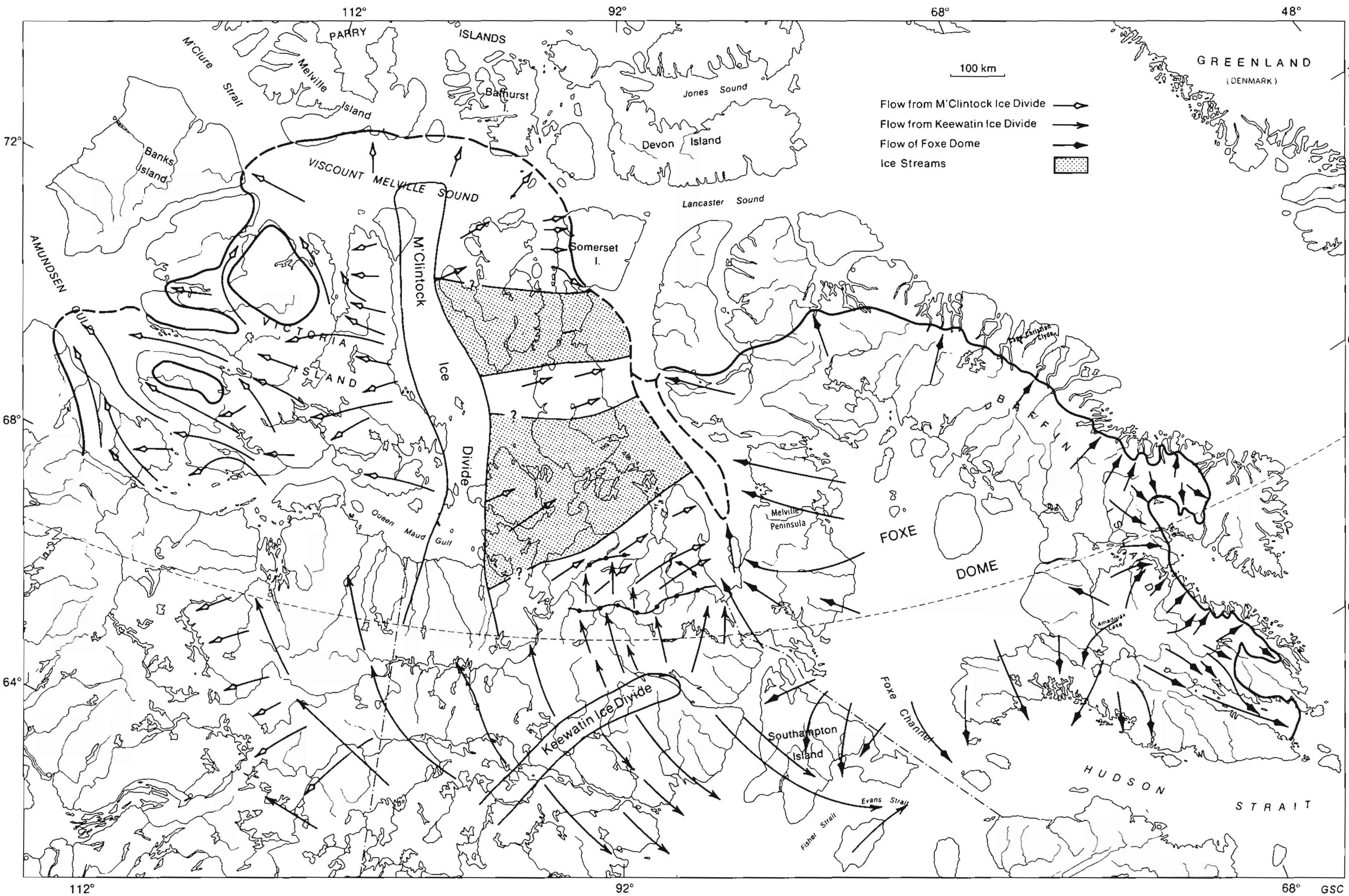

Figure 13. Configuration and dynamics of the northern part of the Laurentide Ice Sheet showing flow patterns and ice streams of the M'Clintock Dome, the Foxe Dome, and the Keewatin ice remnant. The final position of the Keewatin Ice Divide is shown.

(Dyke, in press). Southernmost Somerset Island, Boothia Peninsula, and northern District of Keewatin were completely inundated by M'Clintock ice so a speculative margin is shown offshore in the Gulf of Boothia. Beyond that, an ice shelf could have formed and flowed northward down Prince Regent Inlet although no positive evidence of such an ice shelf has yet been found. North of Somerset Island, the margin is drawn between Lowther and Griffith islands because Barnett et al. (1976) reported signs of "fresh" glaciation on one island but not on the other. On southern Melville Island, the ice sheet formed the Winter Harbour moraine at its maximum stand (Fyles, 1967; Hodgson et al., in press). According to Vincent (1980, 1982), late Wisconsin Laurentide ice barely touched the northeastern tip of Banks Island and much of western Victoria Island lay beyond or above its limit.

In short, the northwestern part of the Laurentide Ice Sheet consisted of an ice mass that had a free margin on three sides – east, north, and west – but that was contiguous with the main part of the ice sheet to the south (Fig. 13). Such a marginal configuration is fully consistent with the structure and dynamics of the M'Clintock Dome as proposed above.

Relation of M'Clintock Ice Divide to Keewatin Ice Divide

The Keewatin Ice Divide is one of the better known dynamic features of the Laurentide Ice Sheet. Lee et al. (1957) and Lee (1959) considered the ice divide to have been a short-lived (about 1000 years) feature that generated flow towards Hudson Bay on one side and away from the Bay on the other side during the last phase of deglaciation of Keewatin, i.e., following removal of the central mass of the Laurentide Ice Sheet upon penetration of the sea into Hudson Bay. Shilts et al. (1979) and Shilts (1980) have shown, however, that the southeastward flow towards Hudson Bay was maintained for a long time, in all likelihood, from the late Wisconsin maximum until deglaciation. They have also shown that towards the close of late glacial time, during deglaciation, the Keewatin Ice Divide migrated southeastward and generated northward to northwestward flow across northern District of Keewatin. That younger flow crosscuts both the northeastward and the southwestward flow from the M'Clintock Ice Divide (Fig. 13). The position of the Keewatin Ice Divide shown in Figure 13 is its last, most southeasterly, position. At the late Wisconsin maximum the M'Clintock Ice Divide probably extended farther southward than shown in Figure 13, and that southward extension was probably the precursor of the late glacial Keewatin Ice Divide; if so, the southeastward flow towards Hudson Bay during the late Wisconsin maximum could represent flow beneath the southern slope of the M'Clintock Dome.

Relation of M'Clintock Dome to Foxe Dome

There are two reasons to believe that a dome, comparable in size to the M'Clintock Dome, covered the Foxe Basin region during the last glacial maximum. First, the large, spatially uniform flow features in that region define a radially flowing ice mass (Fig. 13)[1]; that flow was maintained, at least across Melville Peninsula, for sufficient time to disperse till more than 100 km westward (Sim, 1960). Second, the pattern of postglacial rebound reveals a distinct dome on the uplift surfaces over Foxe Basin (Fig. 3-2 to 3-10 in Dyke, 1974). The margin of the Foxe Dome on Baffin Island (Fig. 13) follows the late Wisconsin Laurentide limit proposed by Miller and Dyke (1974), modified for southernmost Baffin Island according to Miller (1980).

The ice flow features (Fig. 13) indicate that both M'Clintock ice and Foxe ice flowed in an unimpeded manner into the Gulf of Boothia. The only sound evidence of coalescence of the two ice masses is in the form of northward oriented ice flow features on Wales Island at the base of the Gulf of Boothia. For this reason, it is possible that the margins of these ice masses remained separate well into the southern part of the Gulf, as indicated in Figure 13 It is likely, however, that an ice shelf would have formed where ice entered a shallow, elongate inland sea from opposite sides.

Deglacial Events and Features

Detailed reconstruction of the pattern of deglaciation requires careful mapping of ice marginal features. Hence, the following discussion is restricted to the map area and to areas to the north and south where ice marginal features and marine limit were mapped in detail (Thomas and Dyke, 1979a, b; Dyke, in press). Following the discussion of radiocarbon dating control on the timing of deglaciation, some implications of the results for interpreting the late glacial history of Arctic Canada are discussed.

Ice Retreat and Marine Limit

A map showing the pattern of ice marginal recession (Fig. 14) was made by tracing all ice marginal features (moraines, lateral meltwater channels, ice contact escarpments in stratified drift, and proglacial lakes) on 1:250 000 scale topographic bases and correlating the features on the basis of trend and relative topographic position; the map was then reduced to 1:1 000 000 scale.

Discussion of the pattern of deglaciation requires reference also to marine limit (Fig. 15, Map 1570A). Along the east coast of Boothia Peninsula, in Wrottesley valley, and throughout much of northern District of Keewatin the position of marine limit is based primarily on the elevation of perched glaciomarine deltas and to a lesser extent on the highest beaches and patches of marine silt. Along the west coast of Boothia Peninsula south of Wrottesley Inlet, the position of marine limit is clearly marked by distinct washing limits and wave cut notches in till and by the highest beaches, as well as by glaciomarine deltas at a few places.

The oldest ice marginal features, those farthest down ice, are De Geer moraines on Abernethy Lowland of northeastern Boothia Peninsula and end moraine ridges on Simpson Lowland. Both sets of moraines were deposited below the surface of the adjacent postglacial high sea and both define a westward receding, straight to broadly arcuate ice margin. On the higher parts of Boothia Peninsula, which stood above the postglacial sea and which have more relief than lower areas, the configuration of the receding ice margin was naturally more digitate, but recession continued to be generally westward. On the western flank of Boothia Peninsula, ice marginal features, including large end moraines, indicate a southward receding ice margin. This change in the direction of ice retreat is best explained as the result of the progressive southward extension of a marine calving bay along the western side of the Peninsula. Another calving bay penetrated westward across southern Boothia Peninsula and northern District of Keewatin. By the time that bay had reached Rasmussen Basin and joined the sea that had invaded the area west of Boothia Peninsula, deglaciation of the Peninsula was complete and the central part of the M'Clintock Dome had disintegrated.

[1] In fact, a subsidiary dome of considerable size covered southern Baffin Island, where a radial flow pattern centred on Amadjuak Lake (Amadjuak Dome) exists (Blake, 1966; Dyke et al., 1982).

Table 2. Radiocarbon dates pertaining to marine limit on Boothia Peninsula and northern District of Keewatin (cf. Fig. 15)

Date (Years B.P.)	Lab. No.	Sample Elevation (m)	Marine Limit (m)	Location
9430 ± 210	GSC-2093	225-230	240-250	68°58'N, 89°59'W
9230 ± 130	GSC-2720	185	215	71°11'15"N, 93°45'W
9040 ± 100	GSC-2722	123-126	155	69°57'30"N, 95°24'W
8790 ± 80	GSC-2917	130	136	70°25'15"N, 96°01'W
8730 ± 230	GSC-3072	171	184	68°58'45"N, 2°56'30"W
8870 ± 140	GSC-44	155	240-250	68°30'N, 91°05'W
8700 ± 120	GSC-47	171	240-250	68°05'N, 90°09'W

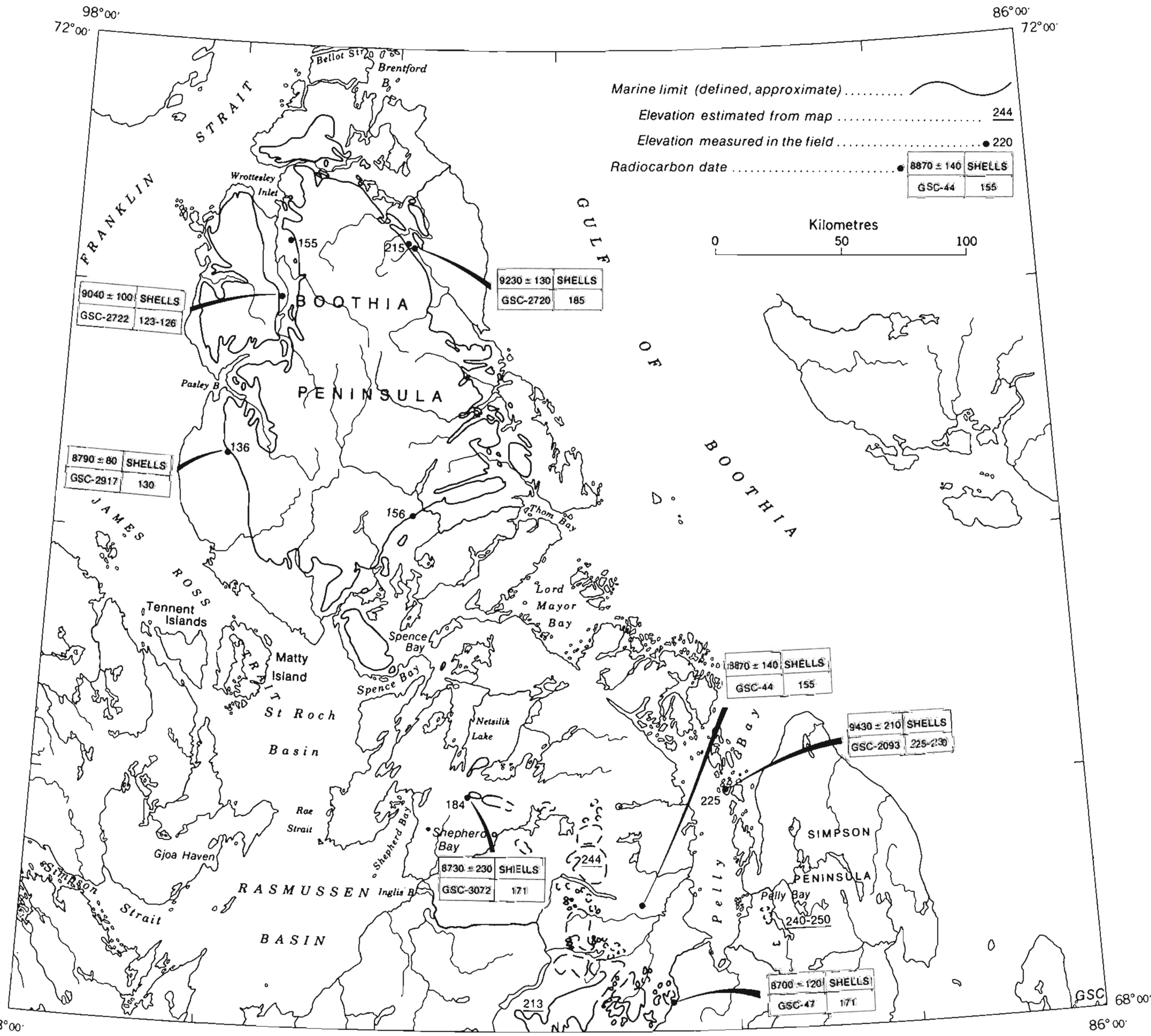

Figure 15. *Area submerged by the postglacial sea on Boothia Peninsula and northern District of Keewatin, showing elevations and ages of marine limit.*

From Shepherd Bay and Pelly Bay south to the general limit of marine submergence and beyond (Fig. 15) the ice margin retreated southward. This pattern is indicated by De Geer moraines in Rasmussen Basin, Pelly Bay, and Chantrey Inlet areas; by larger end moraines between Chantrey Inlet and Committee Bay; and by numerous small ice-dammed lakes that formed in almost every south-facing drainage basin above marine limit.

Elevation of Marine Limit

It is obvious from the preceding discussion of the pattern of ice retreat that the postglacial high sea reached different parts of the study area at different times, but that in general it invaded the east prior to the west and the north prior to the south. This general sequence of events is substantiated by the spatial variations in elevation of the diachronous marine limit (Fig. 15). Marine limit is consistently lower on the west coast than on the east: for example, it is 215 m a.s.l. on northeastern Boothia Peninsula compared to 155 m a.s.l. in the northwest. The same pattern holds at the south end of the study area: marine limit lies 240 to 250 m a.s.l. in the Pelly Bay area but only 184 m a.s.l. in the Shepherd Bay area. Dyke (in press) noted the same relationship on Somerset Island where marine limit declines from 157 m a.s.l. at the head of Creswell Bay on the southeast coast to about 90 m a.s.l. on much of the west coast.

The elevation of marine limit at any given site is a function of two spatially variable factors: (1) the amount of original depression of the crust caused by the ice load and (2) the amount of recovery of that depression (rebound) that occurred beneath the ice cover prior to deglaciation of the site. Because ice was thicker to the west and south than to the east and north of the study area, the original depression of the crust should have been largest on the western and southern side. Because the western and southern parts were deglaciated later than the eastern and northern parts, however, a larger proportion of the recovery occurred beneath the ice cover on the western and southern sides than on the eastern and northern sides. Hence, it is the second factor that accounts for much of the variations in marine limit elevation in the study area.

Age of Marine Limit, Date of Ice Retreat, and Elevation of the 9300 B.P. Shoreline

The sea reached marine limit throughout the study area at the instant of deglaciation. Therefore, a radiocarbon date on a local marine limit feature is also a date on deglaciation. Seven radiocarbon dates on marine shells that relate closely to the marine limit are presently available (Table 2).

The two oldest samples were collected at sites close to marine limit along the east coast (Fig. 15). The oldest date is 9430 ± 210 years B.P. (GSC-2093) on fragments of **Hiatella arctica** from a pocket of glaciomarine silt at 225 to 230 m a.s.l. on the summit of an island in outer Pelly Bay. The marine limit stood above the summit of the island. Both west and east of Pelly Bay, marine limit follows the 800 foot (~244 m) contour and, therefore, is estimated to lie 240 to 250 m a.s.l. Because the shells do not occur in littoral sediment, but in silts deposited in more than a few metres of water, the date is taken as a good measure of the age of marine limit and date of deglaciation. The end moraines on Simpson Peninsula lie down ice from the shell site and, therefore, are somewhat older than the shells. The sample that falls next in age (though the 2σ error terms overlap) dated 9230 ± 130 years B.P. (GSC-2720). It came from distal foreset sand of a glaciomarine delta on northeastern Boothia Peninsula. The sample site is 185 m a.s.l. but the marine limit, marked by the topset gravel of the delta, is 215 m a.s.l. The distal foreset sand and the topset gravel either are of very similar age or are contemporaneous, so the date is a good measure of the age of marine limit. De Geer moraines that occur down ice from the shell site are somewhat older than the shells.

The 2σ error terms of these two dates on marine limit overlap; they also overlap the eleven oldest radiocarbon dates that relate to marine limit and onset of deglaciation on Somerset Island (Dyke, in press). Hence, the oldest ice marginal features and shorelines in the entire area extending from the north coast of Somerset Island to Simpson Peninsula are essentially contemporaneous. Shorelines of about the same age have been dated at several other points in the central and northern Arctic (Table 3), so the present elevation of the 9300 year B.P. shoreline can be sketched (Fig. 16). The isobase pattern shows a rise in elevation of the shoreline towards M'Clintock Channel and the continental mainland from the east, north, and west. The pattern south of Parry Channel is taken to reflect recovery of the crustal depression caused by the M'Clintock Dome.

The oldest of the three samples associated with marine limit along the west coast of the map area is from the northernmost locality (Table 2, Fig. 15). Paired valves of **Mya truncata, Hiatella arctica,** and **Macoma calcarea** were collected from a small deposit of glaciomarine silt and sand that fills a depression in till at 126 m a.s.l. These are not beach sediments, so the shells are thought to have lived in some depth of water. Marine limit in this vicinity is best marked by flat glaciomarine delta tops at 155 m a.s.l. The date of 9040 ± 100 years B.P. (GSC-2722), which was obtained on the **Mya** valves, is considered a reasonable approximation of the date of marine limit and of deglaciation.

The other two samples associated with marine limit along the west coast overlap in age. The most northerly sample consisted mostly of whole valves of **Hiatella arctica** collected from horizontally bedded silty sand 130 m a.s.l. That sand represents distal foreset or upper bottomset beds of a glaciomarine delta. The topset gravel of the delta forms a plain that marks marine limit at 136 m a.s.l. Therefore, the date on the shells, 8790 ± 80 years B.P. (GSC-2917), is a date on marine limit. Inland from Shepherd Bay, marine limit is marked by a distinct washing limit and wave cut notch in till at 184 m a.s.l. Fragments of **Mya truncata** collected from the till surface at 171 m a.s.l. yielded an age of 8730 ± 230 years (GSC-3072)[1]. Because the shells lived in some depth of water, the date is a reasonable approximation of the age of marine limit.

Two dates are available for samples collected by members of the Geological Survey in 1960 from sites west and southwest of Pelly Bay (B.G. Craig in Dyck and Fyles, 1963; and Fig. 15). Both samples came from sediments depicted on Map 1570A as nearshore glaciomarine sediments. Marine limit in the vicinity is estimated at 240 to 250 m a.s.l. The samples came from fine sand and silt at 155 m and 171 m and dated 8870 ± 140 (GSC-44) and 8700 ± 120 (GSC-47) years B.P., respectively. The ages overlap each other and also overlap the two age determinations from the west coast discussed above. Both are reasonable dates on deglaciation.

Emergence Rates During Deglaciation

The difference in elevation of marine limit on the west coast and that of marine limit on the east coast is a minimum measure of the amount of emergence that occurred during the period of deglaciation. The difference is only a minimum measure because the west coast was rising faster, and thus shorelines of any given age dip eastward as does the 9300 year B.P. shoreline (Fig. 16). For this reason, a shoreline on the east coast that is the same age as marine limit on the west coast lies lower than that marine limit.

[1] Collected by J.D. Aitken for B.G. Craig, July 19, 1960.

Table 3. Radiocarbon dates and other data pertaining to the elevation of the 9300 year B.P. shoreline in the central and northern Arctic (cf. Fig. 16)

Superscript on Figure 16	Elevation of 9300 B.P. shoreline(m)	Type and source of data
1	215	Glaciomarine delta with top at 215 m dated 9230 ± 130 years B.P. (GSC-2720); date on shells from foreset sand at 185 m (this report).
2	>225	Shells from silt at 225-230 m dated 9430 ± 210 years B.P. (GSC-2093) (this report).
3	>250	Marine limit in this vicinity at ca. 250 m but dates from about 8700 years B.P.; hence, more than 250 m of emergence in last 9300 years (this report).
4	157	Glaciomarine delta with top at 157 m dated 9240 ± 90 years B.P. (GSC-2561); date on shells from clayey silt at 107 m (Dyke, 1979b).
5	125	Delta (glaciomarine?) with top at 125 m dated 9270 ± 90 years B.P. (GSC-2596); date on shells from upper foreset sand at 120 m (Dyke, 1979b).
6	90	Delta (glaciomarine?) with top at 90 m dated 9310 ± 90 years B.P. (GSC-2272); date on shells in sand 5 m below top (Dyke, 1979b).
7	76	Whale bone from beach at local marine limit dated 9210 ± 120 years B.P. (S-1390) (Dyke, 1979b).
8	102	Delta (glaciomarine?) with top at 102 m dated 8990 ± 210 years B.P. (GSC-2732); date on shells from top of delta (Dyke, 1979b).
9	122	Highest beach marking local marine limit with shells dated 9380 ± 180 years B.P. (GSC-319) (Dyke, 1979b).
10	81	Highest beach, possibly marking local marine limit, with shells dated 9280 ± 150 years B.P. (GSC-241) (Dyck et al., 1965).
11	95	Shells from frost boil in silt slightly above highest distinct beach dated 9260 ± 150 years B.P. (GSC-392) (Dyck et al., 1966).
12	>119	Shells from surface of marine silt at 199 m dated 8830 ± 170 years B.P. (GSC-183) (Dyck et al., 1965).
13	<80	Shells from 75 m in ice contact delta graded to 80 m dated 9500 ± 75 years B.P. (GSC-3318) (Klassen, 1982).
14	>3	Wood in beach gravel at 3 m dated 5990 ± 130 years B.P. (GSC-2064) (Lowdon and Blake, 1979).
15	>50, ≤80	Shells from mud on surface at 50 to 52 m dated 9220 ± 90 years B.P. (GSC-2531) and 9330 ± 110 years B.P. (GSC-3183); highest beach in vicinity at 80 m (Lowdon and Blake, 1981).
16	>>82	Shells from surface at 82 m dated 8090 ± 70 years B.P. (GSC-2692) (Lowdon and Blake, 1979).
17	>107	Shell fragments from surface at 107 m dated 8590 ± 150 years B.P. (GSC-840) (Lowdon et al., 1971).
18	>105	Shell fragments from surface at 105 m dated 8830 ± 130 years B.P. (GSC-881) (Lowdon and Blake, 1975).
19	>120	Shell fragments from surface at 120 m dated 8720 ± 110 years B.P. (GSC-858) (Lowdon and Blake, 1975).
20	118	From emergence curve (Blake, 1975).
21	>114	Shell fragments from beach sand and gravel at 114 m dated 9260 ± 100 years B.P. (GSC-866) (Lowdon and Blake, 1975).
22	60	From emergence curve (Barr, 1971).
23	≥119	Shell fragments from surface at 112 to 119 m dated 9470 ± 150 years B.P. (GSC-322) (Lowdon et al., 1967).
24	>114	Shells on surface at 114 m dated 8550 ± 350 years B.P. (W-1220) (Blake, 1970).
25	>107	Shell fragments from surface at 107 m dated 9160 ± 160 (GSC-722) (Lowdon and Blake, 1968).

Table 3. (cont.)

Superscript on Figure 16	Elevation of 9300 B.P. shoreline(m)	Type and source of data
26	>94	Shells from surface at 94 ± 5 m dated 8430 ± 140 years B.P. (GSC-1128); shells within 30 m of marine limit (Lowdon et al., 1971).
27	≥100	Marine limit here ≤ 100 m and dates from 9000 years B.P. or earlier (Hodgson, 1982).
28	>116	Shell fragments from surface of beach at 116 m dated 8480 ± 140 years B.P. (GSC-244) (Dyck et al., 1965).
29	≥100	Marine limit here ≥100 m and dates from 9000 years B.P. or earlier (Hodgson, 1982).
30	>128	Shell fragments from surface at 128 m dated 9240 ± 160 years B.P. (GSC-182) (Dyck et al., 1965).
31	<137	Basal peat at 137 m dated 9210 ± 170 years B.P. (GSC-180) (Dyck et al., 1965).
32	≤0	Peat from pebbly sand of ice pushed ridge at present sea level (GSC-473, J-S. Vincent, personal communication, 1982).
33	>10	Driftwood log in silt and clay at 10 m dated 7850 ± 140 years B.P. (GSC-1171) (Lowdon et al., 1971).
34	>71 ≤82	Shells on surface at 71.5 m dated 9075 ± 275 years B.P. (I-730); marine limit in vicinity about 82 m; (Trautman, 1964, p. 271).
35	<55	Shells from delta with top at 49 m dated 10 380 ± 160 (GSC-338); higher terrace at 55 m (Hodgson et al., in press).
36	<48	Post Winter Harbour marine limit dating about 9800 years B.P. lies at 48 m (Hodgson et al., in press; Hodgson and Vincent, in press).
37	25	From emergence curve in McLaren and Barnett (1978).
38	<20	Post Winter Harbour marine limit dating about 9800 years B.P. lies at 20 m (Hodgson et al., in press; Hodgson and Vincent, in press).
39	<24	Post Winter Harbour marine limit dating about 9800 years, B.P. lies at 24 m (Hodgson and Vincent, in press).
40	<50	Post Winter Harbour marine limit dating about 9800 years B.P. lies at 50 m (Hodgson and Vincent, in press).
41	>30	Shells in deltaic sand at 30 m dated 9010 ± 120 years B.P. (GSC-3422) (J-S. Vincent, personal communication, 1983).
42	>113, <160	Whale bone at 113 m dated 8640 ± 140 years B.P. (GSC-266) (Dyck et al., 1966); marine limit in vicinity lies at about 160 m; also shells from surface at 153 m dated 9540 ± 150 (GSC-255) (Dyck et al., 1965).
43	>>198	Shells from surface at 198 m dated 8370 ± 100 years B.P. (GSC-115) (Dyck and Fyles, 1964); marine limit in vicinity about 229 m probably dates from about 9200 years B.P. (see location 44).
44	>186	Shells from surface of beaches at 186 m dated 9190 ± 210 years B.P. (Dyck and Fyles, 1964).
45	<220	Shells from silty sand at 200 ± 15 m dated at 9620 ± 130 years B.P. (GSC-3584) (D.A. St-Onge, personal communication, 1983); local marine limit at 220 m.
46	≥150	Shells from surface at 150 m dated at 9100 ± 180 years B.P. (I(GSC)-16) (Craig, 1960).
47	<170	Marine limit here lies at 170 m and dates 9880 ± 90 years B.P. or more (GSC-3327) (St-Onge, in press).
48	≤0	From emergence curve in Vincent (1980).
49	≤0	Late Wisconsin/Holocene marine limit on west coast of Banks Island and along western Arctic mainland coast ≤0 m – that is, region has undergone submergence since last glaciation (Prest et al., 1968; Vincent, 1980).

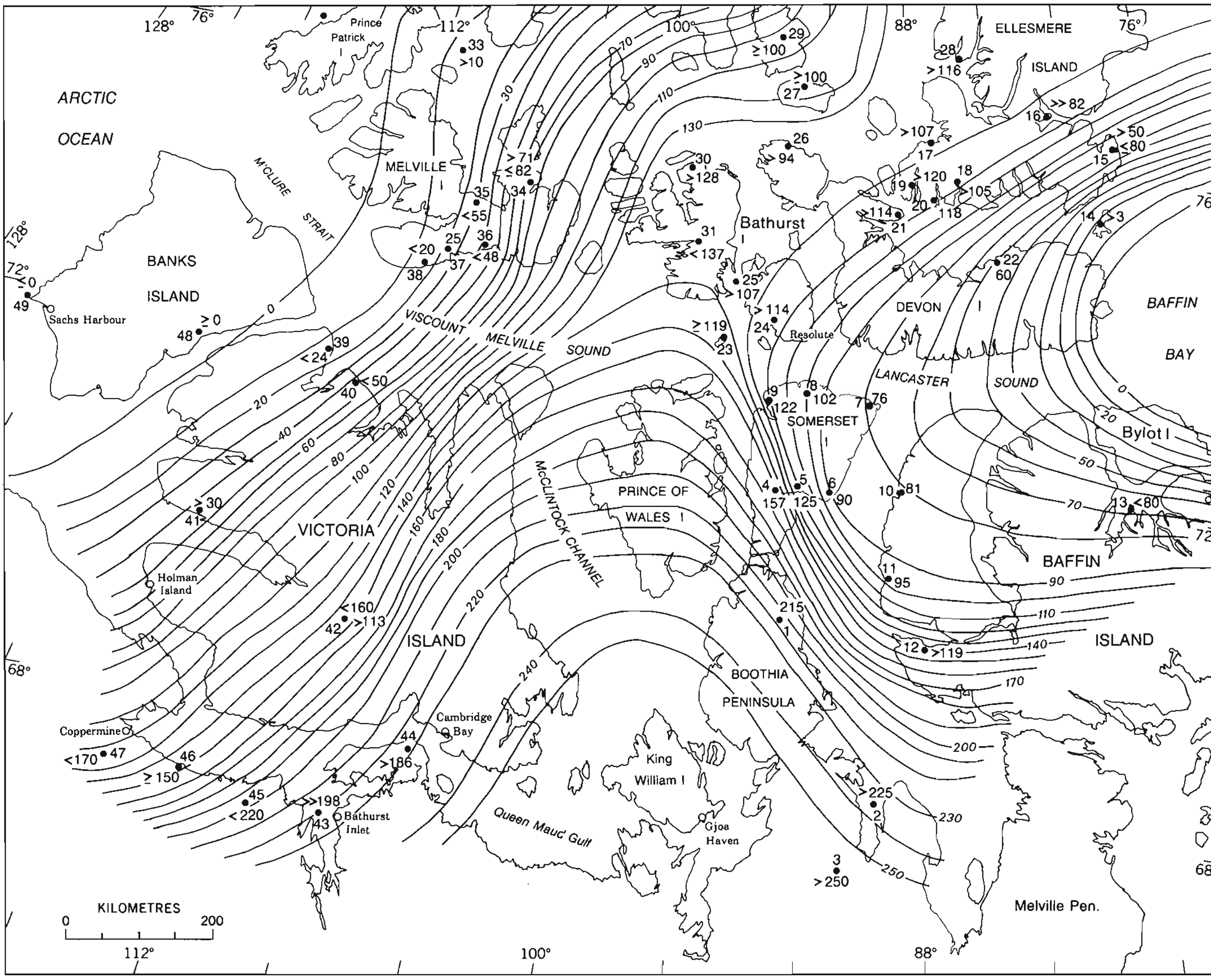

Figure 16. *Isobases on the 9300 year B.P. shoreline in the central Canadian Arctic. Numbers give elevation of the shoreline in metres. Superscripts refer to comments in Table 3.*

Two examples can be given for Boothia Peninsula and northern District of Keewatin. On northwestern Boothia Peninsula, the marine limit at 155 m a.s.l. is 9040 ± 100 years old. It is 60 m lower than the marine limit on northeastern Boothia Peninsula, which lies at 215 m a.s.l. and is 9230 ± 130 years old. Because the shorelines dip eastward, the 9040 ± 100 year old shoreline on the northeast coast must lie more than 60 m below the 215 m marine limit (Fig. 17A). Hence, more than 60 m of emergence occurred in the two centuries or so between formation of the two shorelines. This gives an emergence rate, ignoring the standard errors on the dates, of more than 30 m per century. Taking the 2σ error extremes, the emergence rate could not have been less than 14.3 m per century (>60 m in 420 years).[1]

In the Shepherd Bay area the marine limit at 184 m a.s.l. is 8730 ± 230 years old. It is 56 to 66 m lower than the marine limit in the Pelly Bay area, which lies at 240 to 250 m a.s.l. and is about 9430 ± 210 years old. Because the shorelines dip eastward, the 8730 year-old shoreline must be less than 184 m a.s.l. in the Pelly Bay area and hence more than 56 m below the marine limit in that vicinity (Fig. 17B). Hence more than 56 m of emergence occurred in about 700 years. This gives an emergence rate of more than 8 m per century for northernmost Keewatin.

The rate of emergence during deglaciation of northern Boothia Peninsula calculated above is much higher than rates reported for other areas of Arctic Canada, or to my knowledge, for any other part of the world. Perhaps the explanation is that the very rapid emergence occurred during the process of deglaciation in an area that was rapidly deglaciated (see below). Under such conditions the gravitational attraction of the sea by the ice sheet is rapidly reduced and the elastic portion of crustal rebound is concentrated in a short time interval (Clark, 1976; Farrell and Clark, 1976). Alternatively, unloading may have triggered faulting, but no postglacial faults have been recognized.

Ice Retreat Rates

To summarize the set of radiocarbon dates in terms of rates of ice retreat, ice marginal positions shown in Figure 14 are divided into four age groups. The oldest ice marginal features are more than 9250 years old. During the interval prior to 9250 years B.P. the ice margin in most places retreated in the sea. Because this retreat was probably quite rapid, the oldest ice marginal features in the region are unlikely to be more than a few centuries older than 9250 years.

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1 It is unlikely that both dates are in error by the full extent of the error terms.

During the interval between 9250 and 8800 years B.P. most of Boothia Peninsula was deglaciated. The ice retreated on land on the central part of the peninsula at an average rate of 200 m per year. In the area west of Boothia Peninsula and Somerset Island the ice margin retreated in the sea as a southward calving bay. During the interval 9250 to 8800 years B.P. the average retreat rate was about 670 m per year. In the Pelly Bay area, the ice margin retreated in the sea as well during this same interval; however, the average retreat rate was only 100 m per year, much slower than the retreat in the sea to the northwest and only half the rate of retreat on land on central Boothia Peninsula.

In the two centuries following 8800 years B.P., and possibly in as little as one century, the ice margin retreated to what, at that time, was the continental mainland coast between Chantrey Inlet and Committee Bay. This was the final event in the disintegration of the central part of the M'Clintock Dome. Almost everywhere the margin retreated in the sea. In the west the retreat rate was about 3000 m per year whereas in the east it was only about 400 m per year. Hence, although the retreat rate in the area south of Pelly Bay increased in this time interval, it remained much less than the rate along the west coast. At the end of this interval, end moraines were built in the vicinity of marine limit east of Chantrey Inlet and west of Committee Bay. The moraines are best interpreted as representing a period of equilibration of the ice sheet profile following oversteepening of the ice front due to rapid calving.

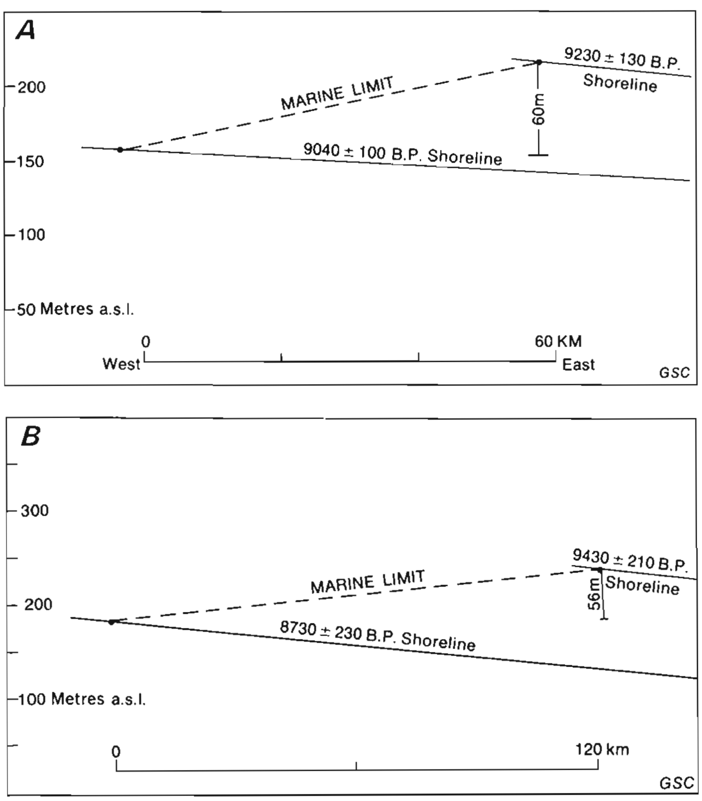


Figure 17. *Approximate profiles of radiocarbon-dated early Holocene shorelines and marine limits on northern Boothia Peninsula (A) and northern District of Keewatin (B); these allow calculation of minimum rates of emergence during the period of deglaciation.*

Last Major Moraine-Building Episode

The youngest ice marginal features shown in Figure 14 are about 8600 years old and younger. Except in the area south of Chantrey Inlet, the ice margin retreated mostly on land. Interruptions of retreat in the early part of this time interval (about 8500 years B.P.?) resulted in deposition of the several end moraines that extend from the head of Chantrey Inlet to the head of Committee Bay (between points A and B on Fig. 14). Unlike the moraines at marine limit (discussed above), these moraines do not simply represent readjustments of the ice sheet profile following rapid mechanical disintegration. Instead, they represent stabilizations or readvances of the ice front that could only have been brought on by climatically induced improvements in the mass balance of the ice sheet. Because of their paleoclimatic significance, it is proposed here that these moraines be named collectively the *Chantrey Moraine System*.

These moraines have been correlated with a large end moraine on Melville Peninsula and with the Cockburn Moraines of Baffin Island (Craig, 1965; Falconer et al., 1965). Moraines of Cockburn age on Baffin Island represent glacier responses to major climatic events just before widespread disintegration of the central area of the Laurentide Ice Sheet. I suggest that the climatic event(s) responsible was increased accumulation on the Foxe Dome brought on by creation of a large new source of moisture when the region formerly occupied by the northern two-thirds of the M'Clintock Dome was invaded by the sea culminating about 8600 years ago. This event would have favoured positive mass balance conditions in the Canadian Arctic even at a time of globally increasing temperatures.

The Cockburn Substage has been formally defined as the interval 8000 to 9000 years B.P. (Andrews and Ives, 1978). Hence, all ice marginal features on southwestern Boothia Peninsula and in northern District of Keewatin are of Cockburn age. Of these features, however, only the Chantrey Moraine System represents significant climatically induced glacier responses. These moraines fall roughly in the middle of Cockburn time (ca. 8500 years ago). On Baffin Island, moraines of Cockburn age that are likely correlatives of the Chantrey Moraine System include the Ranger Moraines, which are about 8600 years old (Dyke, 1979c), and possibly the Frobisher Bay Moraine, which is 8000 to 8500 years old (Blake, 1966; Miller, 1980).

Postglacial Emergence

Emergence during the process of deglaciation, which might be termed "deglacial emergence",[1] was discussed above and shown to be rapid, exceeding 14 m and possibly 30 m per century on northern Boothia Peninsula. The entire study area continued to emerge during postglacial time as it does today.

Curves that describe the progress of emergence through time are truly valid only if they are based on radiocarbon-dated samples collected from a small area (less than 100 km^2 or so). Samples collected from a larger area still can be used to show the general form of emergence, but curves based on them are really composite curves. Eleven radiocarbon-dated samples from northernmost District of Keewatin (Table 4 and Fig. 18), an area of about 17 000 km^2, are used here to construct an emergence envelope.

The older part of the envelope is based on dates that approximate the age of local marine limits and were discussed above. The middle and lower parts of the envelope

[1] This term is preferred over "restrained rebound", proposed by Andrews (1970), because other processes in addition to rebound (e.g. gravitational attraction of water by ice sheet) are involved.

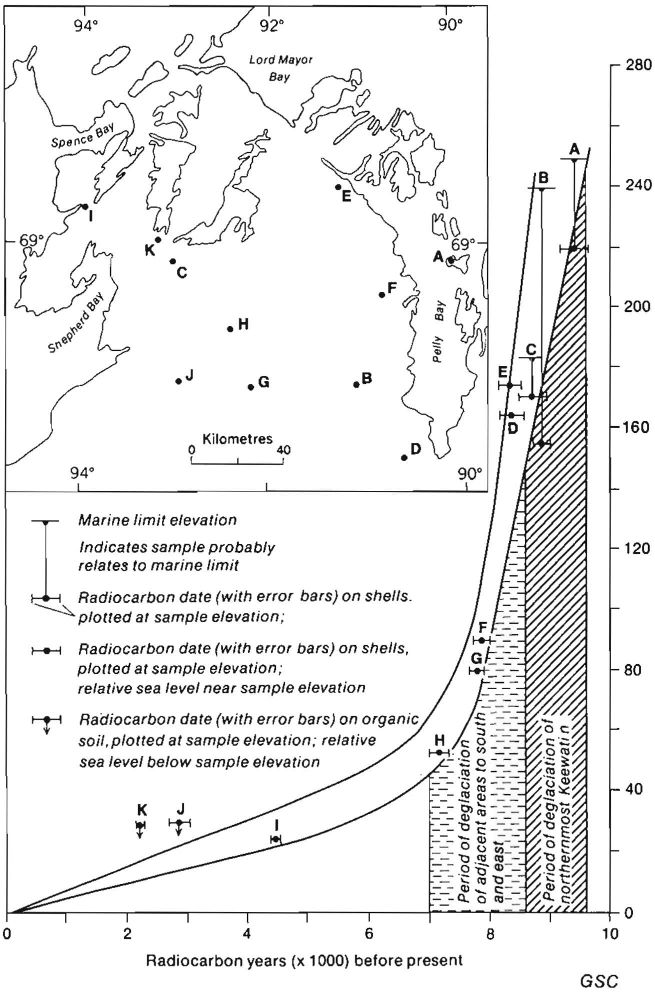


Figure 18. *Regional deglacial and postglacial emergence envelope for northernmost District of Keewatin. Letters on emergence envelope correspond to locations on map.*

are based on dates on shells collected from beach and deltaic sediments and considered by the collectors to have lived at a time when relative sea level stood very near the elevation of the collection sites. In addition to these, two dates on soil organics, which must postdate emergence, constrain the shape of the lowest part of the envelope.

Much of the emergence of northernmost District of Keewatin occurred before 7000 years ago. Some 180 to 200 m out of a total 250 m of emergence, or 70 to 80% of emergence, occurred then at an average rate of about 10 m per century. This corresponds to the period of deglaciation of northenmost District of Keewatin (9600 to 8600 years B.P.) and to the period of deglaciation of areas to the south and east. The major deglacial event to the south occurred about 8000 years ago, when the sea entered Hudson Bay and destroyed the central part of the Laurentide Ice Sheet. The major deglacial event to the east occurred about 7000 years ago, or slightly earlier, when the sea entered Foxe Basin and destroyed the central part of the Foxe Dome (Fig. 13).

Emergence during the last 4500 years occurred at an average rate of at least 53 cm per century, and this is a reasonable estimate of the current rate of emergence. Hence, in parts of Rasmussen Basin, where the regional slope is less than 1:600, the current rate of shoreline regression is more than 3.18 m per year (600 x 5.3 mm per year).

The sharp break in rate of emergence at about 7000 years ago, and its coincidence with the last major unloading event in this region of the Arctic – disintegration of the Foxe Dome – suggests that different processes controlled emergence during the period after 7000 years B.P. than during the preceding period. As mentioned above in the discussion of deglacial emergence rates, the early part of emergence is probably dominated by the elastic portion of rebound and by reduction of gravitational attraction of the sea by the ice sheet. The viscous portion of rebound, caused by flow of the upper mantle, undoubtedly contributed to early emergence as well, but this is difficult to separate from the other processes. Emergence after 7000 years B.P., on the other hand, must be due predominantly, if not solely, to rebound caused by viscous flow of upper mantle material

Table 4. Radiocarbon dates pertaining to deglacial and postglacial emergence of northernmost District of Keewatin

Location on Fig. 18	Date (years B.P.)	Lab. No.	Location	Elevation (m)	Comments	Reference
A	9430 ± 210	GSC-2093	68°58'N 89°59'N	225-230	Shells from glaciomarine silt; sea level at or near marine limit at 240-250 m a.s.l.	This paper
B	8870 ± 140	GSC-44	68°30'N 91°05'W	155	Shells from glaciomarine silt; sea level considerably above site.	Craig in Dyck and Fyles, 1963
C	8730 ± 230*	GSC-3072	68°58'45"N 92°56'30"W	171	Shells from till surface; sea level probably at marine limit at 184 m.	This paper
D	8370 ± 200	I(GSC)-179	68°12'N 90°34'W	165	Shells from nearshore marine silt.	Craig, 1961
E	8360 ± 175	I(GSC)-215	69°15'N 91°16'W	175	Shells from nearshore marine silt.	Craig, 1961
F	7880 ± 150	I(GSC)-213	68°51'30"N 90°48'N	90	Shells from delta foreset beds slightly below contemporaneous sea level.	Craig, 1961
G	7790 ± 100	GSC-46	68°29'N 92°09'W	80	Shells from terrace in horizontally bedded sand.	Craig in Dyck and Fyles, 1963
H	7160 ± 160	I(GSC)-212	68°42'N 92°27'"	53	Shells from delta foreset beds slightly below contemporaneous sea level.	Craig, 1961
I	4460 ± 80	GSC-45	69°09'N 93°59'W	24	Shells from highest beach ridge on crest of a drumlin.	Craig in Dyck and Fyles, 1963
J	2860 ± 90	BGS-246	68°30'N 92°52'W	30	Peat in an earth hummock; sea level lower than site at this time.	Zoltai et al., 1978
K	2210 ± 60	GSC-3017	69°02'15"N 93°12'W	29	Humus under solifluctued till; sea level lower than site at this time.	A.S. Dyke, 1981

*With the exception of this date, all dates are not corrected for isotopic fractionation. This date without correction is 8700 ± 230 B.P.

towards the area beneath the glaciated regions. For this reason, the lower parts of emergence curves are better used as indicators of mantle viscosity than are the upper parts or total curves.

SUMMARY

Pre-Quaternary Evolution

Boothia Peninsula and northern District of Keewatin consist of plateaus flanked by lowlands and plains. The evolution of these major topographic elements was controlled by the distribution of rock types and by structurally controlled tectonic events. The marine channels on either side of Somerset Island, Boothia Peninsula, and northern District of Keewatin, as well as smaller structures such as Wrottesley valley and Bellot Strait, were formed as grabens during the Eurekian Rifting Episode of Miocene to Pliocene times. The lowlands that flank the plateaus are emergent parts of the graben floors. Prior to rifting, these floors probably were accordant with the plateau surface and the two formed part of an extensive surface of erosional planation recognizable throughout much of the central Arctic (Barrow Surface of Bird (1967)).

Surface Materials

Seven genetic groups of surface materials are depicted on Map 1570A. Bedrock forms the surface material over about 40% of the map area. It everywhere shows signs of glacial scour, although the central part of Boothia Peninsula is less heavily scoured than elsewhere. Zones of maximum scouring have large glacial bedforms eroded in rock. Both subglacial and ice marginal meltwater channels are common in bedrock. Till forms the surface material over another 40% of the map area. It is most widespread and thickset over Paleozoic carbonate bedrock and over gneiss situated down-ice from carbonates. End moraines, flutings, and meltwater channels are common landforms of the till sheet. The till sheet ranges from extremely calcareous and fine grained to noncalcareous and coarse grained. The fine grained highly calcareous till defines two large dispersal trains, 150 and 210 km wide. Glaciofluvial sediments, forming kames, eskers, end moraines of stratified drift, and outwash trains are scattered throughout the map area. Glaciolacustrine sediments and landforms define about 20 small ice-dammed lakes on western sloping parts of Boothia Peninsula. Glaciomarine sediments include small ice contact deltas, which record the position of marine limit, and a vast belt of silt and clay, as much as 100 m thick, which extends across the entire southern part of the map area. Marine sediments deposited during postglacial time include small emerged deltas; scattered, thin nearshore silts; and thin but extensive beach gravel. Postglacial fluvial sediments are not extensive; they occur mostly where streams are eroding thick till or glaciomarine sediments.

Economic Geology

Both sand and gravel are plentiful on Pasley Plain and in two large eskers in the southwestern part of the map area. In most of the map area, granular resources are scarce.

Till varies widely in texture and composition but all samples tested classify as ML, CL-ML, or CL soils. Fine glaciomarine sediments tested as CI and CH soils

All unconsolidated materials on slopes are naturally unstable in some measure, and artificial structures should be designed to accommodate that instability. Long-term average rate of movement of soil on slopes is some few millimetres per year under natural conditions, but movement could be greatly increased if moisture supply or pore water pressure was increased. More rapid slope failure occurs naturally in ice-rich sediments, most notably in fine glaciomarine sediment and in thick till. Other natural processes that might prove hazardous to land use activities include thermal contraction and cracking of soil, frost heave of bedrock and soil, ice push at sea and lake shores, and ice scouring of river beds and banks.

Quaternary History

All sediments and features shown on Map 1570A were deposited and formed during the last episode of glaciation (Late Wisconsin) and during postglacial time.

During the height of the last glacial episode, ice flowed eastward across the entire region extending from northern Somerset Island to northern Boothia Peninsula and northeastward to east-northeastward over southern Boothia Peninsula and northern District of Keewatin. The large dispersal trains of fine grained, carbonate-rich till mentioned above were formed by fast flowing streams in the base of the ice sheet.

That flow pattern required a large area of ice dispersal west of the study area. The oldest flow patterns on southern and eastern Victoria Island indicate an area of ice dispersal east of the island. The two flow patterns between them define a north-south oriented ice divide located over M'Clintock Channel. That feature is named the M'Clintock Ice Divide and the ice mass of which it formed the central part is named the M'Clintock Dome of the Laurentide Ice Sheet. The M'Clintock Dome had a free margin on the west, north, and east but was contiguous with the main part of the ice sheet to the south. The southern part of the M'Clintock Dome coalesced with the southwestern part of the Foxe Dome at the southern end of the Gulf of Boothia.

Excellent preservation of numerous ice marginal features and seven radiocarbon dates on glaciomarine deposits associated with marine limit allow detailed reconstruction of the pattern and chronology of ice retreat. The oldest ice marginal features occur along the north and east coasts of Somerset Island, on northeastern Boothia Peninsula, and on Simpson Peninsula, and are more than 9250 years old. Initially retreat was westward but later swung southward as a large calving bay penetrated along the western side of Boothia Peninsula and disintegrated the central mass of the M'Clintock Dome. The northern, marine-based, part of the M'Clintock Dome had completely disappeared by 8700 years B.P. Shortly thereafter, a major moraine system was constructed in northern District of Keewatin by the Keewatin ice remnant, and on Melville Peninsula and Baffin Island by the Foxe Dome, which was still intact. This episode of moraine building is thought to reflect increased accumulation on the Laurentide Ice Sheet brought on by creation of a large new source of moisture when the region formerly occupied by the northern part of the M'Clintock Dome was invaded by the sea.

Emergence during the process of deglaciation exceeded 8 m per century in northern District of Keewatin and exceeded 14 m and perhaps 30 m per century on northern Boothia Peninsula. The entire study area continued to emerge during postglacial time as it does today. In northern Keewatin some 70 to 80% of emergence had been accomplished by 7000 years ago at an average rate of about 10 m per century. Emergence during the last 4500 years has occurred at an average rate of at least 53 cm per century giving a shoreline regression in flat areas of more than 3 m per century.

REFERENCES

Andrews, J.T.
1970: A geomorphological study of postglacial uplift with particular reference to Arctic Canada; Institute of British Geographers, Special Publication No. 2, 156 p.

1973: The Wisconsin ice sheet: dispersal centers, problems of rates of retreat and climatic implications; Arctic and Alpine Research, v. 5, p. 185-200.

Andrews, J.T. and Barnett, D.M.
1972: Analysis of strandline tilt direction in relation to ice centers and post glacial crustal deformation, Laurentide Ice Sheet; Geografiska Annaler, v. 54, series A, p. 1-11.

Andrews, J.T. and Ives, J.D.
1978: "Cockburn" nomenclature and the late Quaternary history of the eastern Canadian Arctic; Arctic and Alpine Research, v. 10, p. 617-633.

Andrews, J.T. and Miller, G.H.
1979: Glacial erosion and ice sheet divides, northeastern Laurentide Ice Sheet, on the basis of the distribution of limestone erratics; Geology, v. 7, p. 592-596.

Atmospheric Environment Service
1975: Canadian Normals – Temperature, Precipitation 1941-1970; Environment Canada, Downsview, v. 1, 198 p., and v. 2, 333 p.

Barnett, D.M., Dredge, L.A., and Edlund, S.A.
1976: Terrain inventory: Bathurst, Cornwallis, and adjacent islands, Northwest Territories; in Report of Activities, Part A, Geological Survey of Canada, Paper 76-1A, p. 201-204.

Barr, W.
1971: Postglacial isostatic movement in northeast Devon Island: a reappraisal; Arctic, v. 24, p. 249-268.

Bird, J.B.
1967: The Physiography of Arctic Canada with Special Reference to the Area South of Parry Channel; The Johns Hopkins Press, Baltimore, 336 p.

Blake, W., Jr.
1966: End moraines and deglaciation chronology in northern Canada, with special reference to southern Baffin Island; Geological Survey of Canada, Paper 66-26, 31 p.

1970: Studies of glacial history in Arctic Canada. I. Pumice, radiocarbon dates, and differential postglacial uplift in the eastern Queen Elizabeth Islands; Canadian Journal of Earth Sciences, v. 7, p. 634-664.

1975: Radiocarbon age determinations and postglacial emergence at Cape Storm, southern Ellesmere Island, Arctic Canada; Geografiska Annaler, v. 57, series A, p. 1-71.

Bostock, H.S.
1969: Physiographic Regions of Canada; Geological Survey of Canada, Map 1254A, 1:5 000 000.

Boydell, A.N., Drabinsky, K.A., and Netterville, J.A.
1975a: Surficial geology and geomorphology maps, Boothia Peninsula and Simpson Peninsula, N.W.T.; Geological Survey of Canada, Open File 285, scale 1:125 000.

Boydell, A.N., Drabinsky, K.A., and Netterville, J.A. (cont.)
1975b: Terrain inventory and land classification, Boothia Peninsula and northern Keewatin; in Report of Activities, Part A, Geological Survey of Canada, Paper 75-1A, p. 393-396.

Brown, R.J.E.
1978: Influence of climate and terrain on ground temperatures in the continuous permafrost zone of northern Manitoba and Keewatin District, Canada; in Proceedings of the 3rd International Conference on Permafrost (Edmonton), p. 15-21.

Chamberlain, T.C.
1894: Glacial phenomena of North America; in The Great Ice Age and its Relation to the Antiquity of Man by J. Geikie; Edward Stanford (London) 3rd Edition, p. 724-775.

Clark, J.A.
1976: Greenland's rapid postglacial emergence: a result of ice-water gravitational attraction; Geology, v. 4, p. 310-312.

Craig, B.G.
1960: Surficial geology of north-central District of Mackenzie, Northwest Territories; Geological Survey of Canada, Paper 60-18, 8 p.

1961: Surficial geology of northeastern District of Keewatin, Northwest Territories; Geological Survey of Canada, Paper 61-5, 6 p.

1964: Surficial geology of Boothia Peninsula and Somerset, King William, and Prince of Wales islands, District of Franklin; Geological Survey of Canada, Paper 63-44, 10 p.

1965: Notes on moraines and radiocarbon dates in northwest Baffin Island, Melville Peninsula, and northeast District of Keewatin; Geological Survey of Canada, Paper 65-20, 7 p.

Dawson, G.M.
1886: Notes to accompany a geological map of the northern portion of the Dominion of Canada east of the Rocky Mountians; Geological Survey of Canada, Annual Report 2, p. 1R-62R.

1890: On the glaciation of the northern part of the Cordillera, with an attempt to correlate the events of the glacial period in the Cordillera and Great Plains; American Geologist, v. 6, p. 153-162.

Dreimanis, A. and Vagners, U.J.
1972: The effect of lithology upon texture of till; in Yatsu, E and Falconer, A. (eds.) Research Methods in Pleistocene Geomorphology, ed. E. Yatsu and A. Falconer; Proceedings, 2nd Guelph Symposium on Geomorphology, Geoabstracts Ltd., Norwick, p. 66-82.

Dyck, W. and Fyles, J.G.
1963: Geological Survey of Canada radiocarbon dates I and II; Geological Survey of Canada, Paper 63-21, 31 p.

1964: Geological Survey of Canada radiocarbon dates III; Geological Survey of Canada, Paper 64-40, 15 p.

Dyck, W., Fyles, J.G., and Blake, W., Jr.
1965: Geological Survey of Canada radiocarbon dates IV; Geological Survey of Canada, Paper 65-4, 23 p.

Dyck, W., Lowdon, J.A., Fyles, J.G., and Blake, W., Jr.
1966: Geological Survey of Canada radiocarbon dates V; Geological Survey of Canada, Paper 66-48, 32 p.

Dyke, A.S.
1974: Deglacial chronology and uplift history: northeastern sector, Laurentide Ice Sheet; Colorado University, Institute of Arctic and Alpine Research, Occasional Paper No. 12, 113 p.

1979a: Glacial geology of northern Boothia Peninsula, District of Franklin; in Current Research, Part B, Geological Survey of Canada, Paper 79-1B, p. 385-394.

1979b: Radiocarbon-dated Holocene emergence of Somerset Island, central Canadian Arctic; in Current Research, Part B, Geological Survey of Canada, Paper 79-1B, p. 307-318.

1979c: Glacial and sea-level history of southwestern Cumberland Peninsula, Baffin Island, Northwest Territories, Canada; Arctic and Alpine Research, v. 11, p. 179-202.

1980: Base metal and uranium concentrations in till, northern Boothia Peninsula, District of Franklin; in Current Research, Part C, Geological Survey of Canada, Paper 80-1C, p. 155-159.

1981: Late Holocene solifluction rates and radiocarbon soil ages, central Canadian Arctic; in Current Research, Part C, Geological Survey of Canada, Paper 81-1C, p. 17-22.

– Quaternary geology of Somerset Island, District of Franklin; Geological Survey of Canada, Memoir 404. (in press)

Dyke, A.S. and Zoltai, S.C.
1980: Radiocarbon dated mudboils, central Canadian Arctic; in Current Research, Part B, Geological Survey of Canada, Paper 80-1B, p. 271-276.

Dyke, A.S., Dredge, L.A., and Vincent, J-S.
1982: Configuration and dynamics of the Laurentide Ice Sheet during the late Wisconsin maximum; Géographie physique et Quaternaire, v. 36, no. 1-2, p. 5-14.

Dyke, L.D.
1979: Bedrock heave in the central Canadian Arctic; in Current Research, Part A, Geological Survey of Canada, Paper 79-1A, p. 241-246.

1981: Bedrock heave in the central Canadian Arctic; in Current Research, Part A, Geological Survey of Canada, Paper 81-1A, p. 157-167.

Falconer, G., Ives, J.D., Loken, O.H., and Andrews, J.T.
1965: Major end moraines in eastern and central Arctic Canada; Geographical Bulletin, v. 7, p. 137-153.

Farrell, W.E. and Clark, J.A.
1976: On postglacial sea level; Geophysical Journal of the Royal Astronomical Society, v. 46, p. 647-667.

Flint, R.F.
1943: Growth of the North American Ice Sheet during the Wisconsin Age; Geological Survey of America, Bulletin, v. 54,p. 325-362.

1971: Glacial and Quaternary Geology; John Wiley & Sons, Inc., New York, 892 p.

Fyles, J.G.
1963: Surficial Geology of Victoria and Stefansson Islands, District of Franklin; Geological Survey of Canada, Bulletin 101, 38 p.

Fyles, J.G. (cont.)
1967: Winter Harbour moraine, Melville Island; in Report of Activities, Geological Survey of Canada, Paper 67-1A, p. 89.

Gray, J., de Boutray, B., Hillaire-Marcel, C., and Lauriol, B.
1980: Postglacial emergence of the west coast of Ungava Bay, Quebec; Arctic and Alpine Research, v. 12, p. 19-30.

Hodgson, D.A.
1982: Surficial materials and geomorphological processes, Western Sverdrup and adjacent islands, District of Franklin; Geological Survey of Canada, Paper 81-9, 37 p.

Hodgson, D.A. and Vincent, J-S.
– A 10,000 yr. B.P. extensive ice shelf over Viscount Melville Sound at the limit of Laurentide ice in Western Arctic archipelago, Canada; Quaternary Research. (in press)

Hodgson, D.A., Vincent, J-S., and Fyles, J.G.
– The Quaternary geology of central Melville Island, Northwest Territories; Geological Survey of Canada, Paper 83-16. (in press)

Hughes, T., Denton, G.H., and Grosswald, M.G.
1977: Was there a late Wurm Arctic ice sheet?; Nature, v. 266, p. 596-602.

Ives, J.D., Andrews, J.T., and Barry, R.G.
1975: Growth and decay of the Laurentide Ice Sheet and comparisons with Fenno-Scandinavia; Naturwissenchaften, v. 62, p. 118-125.

Kerr, J.Wm.
1977: Cornwallis Fold belt and the mechanism of basement uplift; Canadian Journal of Earth Sciences, v. 14, p. 1375-1401.

1980: Structural framework of Lancaster Aulacogen, Arctic Canada; Geological Survey of Canada, Bulletin 319, 24 p.

Kerr, J.Wm. and deVries, C.D.S.
1977: Structural geology of Somerset Island and Boothia Peninsula, District of Franklin; in Report of Activities, Part A, Geological Survey of Canada, Paper 77-1A, p. 107-111.

Klassen, R.A.
1982: Quaternary stratigraphy and glacial history of Bylot Island, N.W.T., Canada; unpublished Ph.D. thesis, University of Illinois, Urbana, 161 p.

Lee, H.A.
1959: Surficial geology of southern District of Keewatin and the Keewatin ice divide, Northwest Territories; Geological Survey of Canada, Bulletin 51, 42 p.

Lee, H.A., Craig, B.G., and Fyles, J.G.
1957: Keewatin Ice Divide (abstract); Geological Society of America Bulletin, v. 68, p. 1760-1761.

Lowdon, J.A. and Blake, W., Jr.
1968: Geological Survey of Canada radiocarbon dates VII; Geological Survey of Canada, Paper 68-2, Part B, p. 207-245.

1975: Geological Survey of Canada radiocarbon dates XV; Geological Survey of Canada, Paper 75-7, 31 p.

1979: Geological Survey of Canada radiocarbon dates XIX; Geological Survey of Canada, Paper 79-7, 57 p.

1981: Geological Survey of Canada radiocarbon dates XXI; Geological Survey of Canada, Paper 81-7, 22 p.

Lowdon, J.A., Fyles, J.G., and Blake, W., Jr.
1967: Geological Survey of Canada radiocarbon dates VI; Geological Survey of Canada, Paper 67-2, Part B, 42 p.

Lowdon, J.A., Robertson, I.M., and Blake, W., Jr.
1971: Geological Survey of Canada radiocarbon dates XI; Geological Survey of Canada, Paper 71-7, p. 255-324.

Mackay, J.R.
1965: Glacier flow and analogue simulation; Geographical Bulletin, v. 7, p. 1-6.

Mackay, J.R. and MacKay, D.K.
1976: Cryostatic pressures in nonsorted circles (mud hummocks), Innuvik, Northwest Territories; Canadian Journal of Earth Sciences, v. 13, p. 889-897.

McLaren, P. and Barnett, D.M.
1978: Holocene emergence of the south and east coasts of Melville Island, Queen Elizabeth Islands, Northwest Territories, Canada; Arctic, v. 31, p. 415-427.

Miller, G.H.
1980: Late Foxe glaciation of southern Baffin Island, Northwest Territories, Canada; Geological Society of America Bulletin, Part 1, v. 91, p. 399-405.

Miller, G.H. and Dyke, A.S.
1974: Proposed extent of late Wisconsin Laurentide ice on Baffin Island; Geology, v. 2, p. 125-130.

Prest, V.K., Grant, D.R., and Rampton, V.N.
1968: Glacial Map of Canada; Geological Survey of Canada, Map 1253A; 1:5 000 000.

St-Onge, D.A.
– Deglaciation pattern, Coppermine River Region, Northwest Territories; Boreas. (in press)

Shilts, W.W.
1980: Flow patterns in the central North American ice sheet; Nature, v. 286, p. 213-218.

Shilts, W.W. and Klassen, R.A.
1976: Drift prospecting in the District of Keewatin – Uranium and base metals; in Report of Activities, Part A, Geological Survey of Canada, Paper 76-1A, p. 255-257.

Shilts, W.W., Cunningham, C.M., and Kaszycki, C.A.
1979: Keewatin Ice Sheet – Re-evalution of the traditional concept of the Laurentide Ice Sheet; Geology, v. 7, p. 537-541.

Sim, V.W.
1960: A preliminary account of late Wisconsin glaciation in Melville Peninsula, Northwest Territories; Canadian Geographer, no. 17, p. 21-34.

Sugden, D.E.
1978: Glacial erosion by the Laurentide Ice Sheet; Journal of Glaciology, v. 20, p. 367-391.

Tarnocai, C. and Boydell, A.N.
1975: Biophysical study of the Boothia Peninsula and Northern Keewatin; in Report of Activities, Part A, Geological Survey of Canada, Paper 75-1A, p. 123-424.

Tarnocai, C. and Zoltai, S.C.
1978: Earth hummocks of the Canadian Arctic and Subarctic; Alpine and Arctic Research, v. 10, p. 581-594.

Taylor, R.B.
1978: The occurrence of grounded ice ridges and shore ice piling along the northern coast of Somerset Island, Northwest Territories; Arctic, v. 31, p. 133-149.

Thomas, R.D. and Dyke, A.S.
1979a: Surficial geology and geomorphology, north-central Keewatin; Geological Survey of Canada, Open File 626, 4 maps, 1:125 000 scale.

1979b: Surficial geology and geomorphology, north-central Keewatin, Northwest Territories (Part II); Geological Survey of Canada, Open File 642, 4 maps, 1:125 000 scale.

Trautman, M.A.
1964: Isotopes, Inc. radiocarbon measurements IV; Radiocarbon, v. 6, p. 269-279.

Tyrrell, J.B.
1898a: Report on the Doobaunt, Kazan, and Ferguson Rivers, and the northwest coast of Hudson Bay; and on two overland routes from Hudson Bay to Lake Winnipeg; Geological Survey of Canada, Annual Report, Series N, v. 9, p. 1-218.

1898b: The glaciation of north central Canada; Journal of Geology, v. 6, p. 147-160.

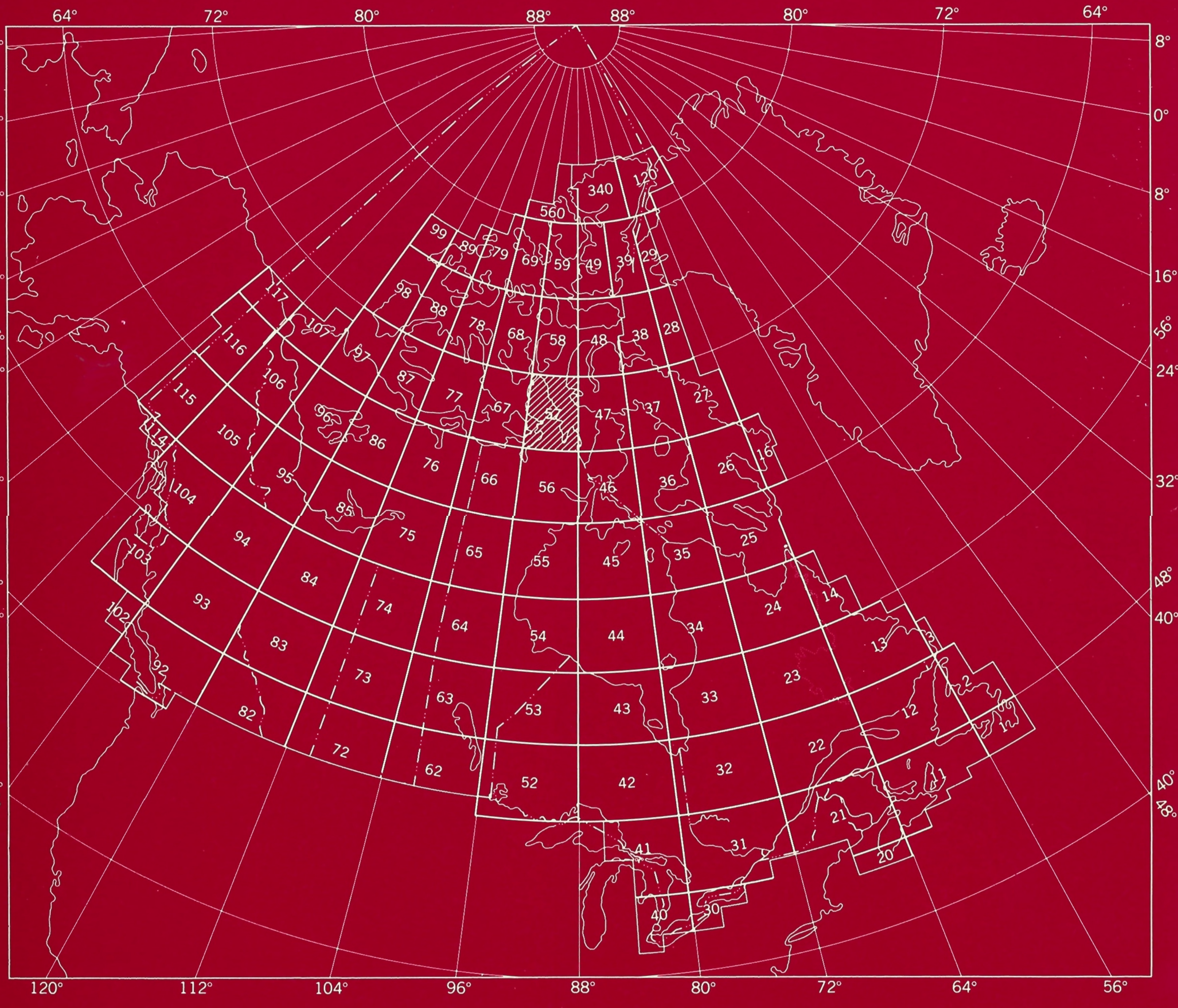
1913: The Patrician glacier south of Hudson Bay; Congrès geologique international, compte rendu de la XII session, Canada, p. 523-534.

Vincent, J-S.
1980: Les glaciations quaternaires de l'isle de Banks, Arctique Canadien; thèse de D.Sc. non publiée, Université de Bruxelles, 248 p.

1982: The Quaternary history of Banks Island; Géographie physique et Quaternaire, v. 36, no. 1-2, p. 209-232.

Vincent, J-S. and Hardy, L.
1979: The evolution of glacial lakes Barlow and Ojibway, Quebec and Ontario; Geological Survey of Canada, Bulletin 316, 18 p.

Zoltai, S.C., Tarnocai, C., and Pettapiece, W.W.
1978: Age of cryoturbated organic materials in earth hummocks from the Canadian Arctic; in Proceedings of the Third International Conference on Permafrost (Edmonton), p. 325-331.



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