R. Baragar.

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DEPARTMENT OF MINES
AND TECHNICAL SURVEYS

ECONOMIC GEOLOGY
REPORT No. 22

GEOLOGY OF IRON DEPOSITS IN CANADA

Volume I

General Geology and Evaluation of Iron Deposits

G. A. Gross

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1965

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General Geology and Evaluation of Iron Deposits

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PLATE I. RUTH LAKE MINE IN THE KNOB LAKE IRON RANGES OF QUEBEC AND LABRADOR, NEWFOUNDLAND. (JULY 1957)

G.A.G. 3-1-57

PREFACE

In the decade ending 1962 Canada became one of the world's leading producers of iron ore. During that time production increased from 5 million tons a year to 25 million tons and iron ore became the fourth largest mineral commodity—nearly 10 per cent of the mineral wealth of the country.

A special study of the geology of iron deposits in Canada was begun by the author in 1957, the results of which will appear in several volumes of this series. Volume I is a general discussion of iron deposits and serves as an introduction to the subsequent volumes in which iron deposits in specific geological provinces will be described. Because iron-formations or bedded deposits provide most of the iron ore in Canada, their characteristics and origin are considered in detail. A revised classification is proposed.

The section giving statistics on iron-ore production was prepared by the Mineral Resources Division of this Department.

J. M. HARRISON,
Director, Geological Survey of Canada

OTTAWA, February 25, 1964

WIRTSCHAFTSGEOLOGISCHER BERICHT Nr. 22 —DIE GEOLOGIE DER KANADISCHEN EISENERZLAGER

Band 1: Allgemeine Geologie und Bewertung der Eisenerzlager.

Von Gordon A. Gross

Dieser Bericht enthält Klassifikationen von Eisenformationen.

ЭКОНОМИЧЕСКАЯ ГЕОЛОГИЯ ОТЧЁТ № 22 ГЕОЛОГИЯ ЗАЛЕЖЕЙ ЖЕЛЕЗНЫХ РУД КАНАДЫ.

Том I: Общая геология и оценка залежей железных руд.

Гордон А. Гросс.

В отчёте дается классификация железных орудных формаций.

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GENERAL GEOLOGY AND EVALUATION OF IRON DEPOSITS

Abstract

In 1938, after a 15-year idle period, iron ore production was resumed in Canada; a 25-million-ton output in 1962 reflected a five-fold increase within 10 years. Reference data and an introduction to geology of iron ore deposits are provided in this first volume of *Iron Deposits in Canada.*

The ubiquitous distribution of iron in rocks and the conditions affecting its solution, migration, and precipitation are emphasized in discussion of its geochemistry. The common minerals found in iron ore are described, and a review of the genesis of these minerals indicates more widely diversified conditions for the formation of iron oxides than for the formation of carbonates, silicates, and sulphides.

Twenty different types of iron ore deposits are described in an empirical classification, but the most extensive concentrations of iron occur in four kinds of iron-formation designated as Algoma, Superior, Clinton, and Minette types. The coarse-grained metamorphosed Superior type (metataconite) and the finer grained cherty iron-formations (taconite) are predominant in Canada. They provide almost unlimited ore reserves and are the basis for the greater part of the iron ore industry. Residual concentrations of iron formed by weathering of iron-formations are also a major source of ore in Canada and in many other parts of the world. Iron-formations are found mainly in eastern Canada and replacement magnetite and other deposits directly associated with igneous rocks are prominent in the west coast region and immediately north of the St. Lawrence River.

Algoma-type iron-formations are intimately associated with volcanic rocks in eugeosynclinal belts whereas the Superior type formed with continental shelf sediments and volcanic rocks in broad shallow coastal basins. These cherty formations are common in Precambrian rock successions but are not restricted to that geological era. The Clinton and Minette types were also formed near the shores of coastal basins and have prominent oölitic textures, complex mineralogy, and a higher alumina and phosphorus content than the Algoma and Superior types, but in contrast they generally lack chert. In the Algoma- and Superior-type formations evidence is stronger for a volcanic source for the iron and silica than for derivation of these elements by weathering of a landmass, which source is usually postulated for iron in the Clinton- and Minette-type beds.

Résumé

En 1938, après une période creuse de 15 années, on a recommencé à produire du minerai de fer au Canada. La production de 1962 (25 millions de tonnes) indique que depuis 10 ans la production a quintuplé. Le premier volume de «Iron Deposits in Canada» comporte des chiffres et une introduction à la géologie des gisements de minerai de fer.

L'auteur fait une étude poussée de la géochimie du fer, sa répartition généreuse dans les roches et les conditions qui régissent sa solution, sa migration et sa précipitation. Il décrit les minéraux communs que l'on trouve dans le minerai de fer, et une étude de leur genèse permet d'établir que les conditions de formation des oxydes de fer sont beaucoup plus diversifiées que celles qui régissent la formation des carbonates, des silicates et des sulfures.

Une classification empirique donne vingt genres différents de gisements de minerai de fer, mais les plus grandes concentrations se trouvent dans quatre genres de formations connues sous les noms d'Algoma, Superior, Clinton et Minette. La formation Superior (métataconite) à gros grains, métamorphisée et les formations cherteuses à grains plus fins (taconite) prédominent au Canada. Elles renferment des réserves presque illimitées de minerai et alimentent la majeure partie de l'industrie du minerai de fer. Les concentrations résiduelles de fer provenant de l'altération des formations ferrifères par les intempérismes sont aussi une source importante de minerai au Canada et dans plusieurs autres parties du monde. Les formations ferrifères se trouvent surtout dans l'Est du Canada, tandis que la magnétite de substitution et les autres gisements directement associés aux roches ignées sont plus répandus dans la région de la côte Ouest de même qu'immédiatement au nord du fleuve Saint-Laurent.

Les formations ferrifères du genre Algoma sont intimement associées à des roches volcaniques dans des bandes eugéosynclinales tandis que les gisements du genre Superior ont été formés avec des sédiments du plateau continental et des roches volcaniques mis en place dans de larges dépressions peu profondes des bassins côtiers. Ces formations cherteuses sont nombreuses dans les successions de roches précambriennes, mais ne sont pas confinées à cette ère géologique. Les formations des types Clinton et Minette se sont aussi formées près des limites des bassins côtiers et présentent des textures fortement oolitiques, une minéralogie complexe et une plus forte teneur en alumine et en phosphore que les formations des types Algoma et Superior, mais, de façon générale, on n'y trouve point de chert. Dans les formations des types Algoma et Superior, tout porte à croire que le fer et la silice proviennent de source volcanique plutôt que de l'altération de masses de terre par intempérisme, phénomène qui serait habituellement à l'origine des concentrations de fer dans les gisements de types Clinton et Minette.

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Chapter I

INTRODUCTION

From the time of man's first use of iron, at least as early as 3500 B.C., his cultural development and his knowledge of the production and use of this grey metal have been inseparable. The strength and influence of nations in world affairs have been measured to a great extent by the size and quality of their iron resources and the ingenuity of their people in utilizing such resources.

Iron ranks fourth in abundance among the elements and occurs in many different forms. Natural occurrences with physical characteristics and compositions suitable for ore are abundant throughout the world, and their efficient development and use have challenged man since antiquity.

In the varied geological settings of the North American Continent the iron deposits are the possible source of many different kinds of iron ore. In Canada these deposits have been developed on a large scale within the past 15 years in order to meet an expanding export market and, to a lesser extent, to supply domestic needs. Ore is shipped to established steel industries in the United States, Europe, and Japan where domestic reserves have been depleted and where higher grade natural ore and concentrate are desired.

In Canada, many different kinds of natural ores are available that can be produced for direct shipment to steel plants or processed to satisfy rigid grade and structural specifications. Records show that more than 1,200 iron deposits or iron ranges occur; in 1962 four provinces produced ore (six had produced previously), and in all other provinces and territories deposits have been explored and developed to some degree. In some remote regions many unexplored deposits have been noted.

This introductory volume on iron in Canada deals with the geochemistry, mineralogy, origin, and distribution of the main types of iron occurrences, as well as with the geological features of particular importance to exploration, and to efficient evaluation, development, and exploitation of the deposits. Various types of deposits are described, and their general distribution in relation to the regional geology and tectonic framework is shown on the accompanying map. It is also the purpose of this volume to give an introduction to the metallogeny of iron deposits and to provide basic reference data for use in the investigation and evaluation of iron occurrences.

Acknowledgments

The interest shown by colleagues of the Geological Survey and by industry in this study has been a constant source of encouragement, and their comments and criticisms regarding the organization and presentation of many parts of this introductory work are gratefully acknowledged. The cooperation received from property owners and their representatives in permitting examination of iron deposits is also greatly appreciated.

History of Iron and Steel Production

Iron ornaments, jewellery or primitive tools are found with the relics of early cultures in nearly all parts of the world and were used at least as early as 3500 B.C. These were made of meteoric iron and there is no evidence to indicate that the ancient peoples knew how to make iron metal from terrestrial material. It is easily understood why metallic iron was valued above the more common and more easily worked gold and copper. Occurrences of native iron of terrestrial origin are rare and native iron was probably in use a long time before the reduction of iron from natural ores was accomplished.

How man first discovered a method to reduce iron minerals to metal may never be known. Perhaps hot charcoal produced in a campfire or forest fire may have accidentally effected the reduction of iron-rich stones to porous sponge metal and led to conscious attempts to repeat the process. Experiments to produce metallic iron in this manner have, however, failed and it is more probable that iron may have been produced accidentally while smelting copper or gold ores, a process that was carried out long before the working of natural iron was known. Also iron may have been produced by accident in pottery furnaces where sufficient heat and a satisfactory reducing environment with charcoal were attainable.

The making of hard and useful iron as early as 1500 B.C. is attributed to the Hittites and tribes living south of the Black Sea in Asia Minor, of whom Tubal-cain is a legendary craftsman-hero of the metal workers. Iron was not generally used by the Hebrew tribes until after the time of the Exodus — a period from 1400 to 1200 B.C. — and a general increase in its use about 1200 B.C. is looked on as the beginning of the true iron industry. The use and working of iron appear to have spread through the ancient world from Asia Minor, and iron working is mentioned in Greece and the Aegean around 900 B.C., in Tuscany about 800 B.C., in the Alpine regions in the middle of the seventh century B.C., and was later distributed in Central and Western Europe under the Hallstatt influence. There is evidence to indicate that iron was introduced in northern India as early as 1000 B.C. but iron tools did not become common in southern India until after 500 B.C. In China, iron was worked as early as 500 B.C. but the transition from the bronze to the iron age does not seem to have been completed until the Han dynasty of 206 B.C. to

A.D. 220. It is of interest that cast iron may have been made in China almost as early as wrought iron.

Except for meteoric iron, nearly all the primitive iron was wrought iron which is malleable and can be worked and shaped by hammering. The first use of the more brittle cast iron poured from molten metal is not definitely established, but a ring-shaped casting from Moravia is dated back to 600 B.C. and a cauldron from China to around 512 B.C. Cast iron may have been made sporadically in Europe in Roman times but there is no substantial evidence for a significant production of molten metal and castings until the development of a blast furnace in the early part of the fourteenth century. The iron furnaces of medieval times were open hearths built of rough masonry and resemble the Catalan forges first used in northern Spain. The first true blast furnace, built in the district of Liége in Belgium before 1400, was apparently evolved by the Germans by increasing the height of their rudimentary stückoffen furnaces and increasing the air blast by bellows operated by water-wheels. The production of molten iron suitable for casting was achieved by the increased contact of the iron with carbon in such a furnace, and the higher temperatures obtained with the increased air blast causing greater absorption of carbon in the metal. In Burwash cemetery in Sussex, England, a grave slab of cast iron of the fourteenth century suggests that this was the area where early production of cast iron took place in that country, and a breech loading mortar with a cast iron interior of the fifteenth century was found at Battle Abbey. Cannons of cast iron were made in Sussex in 1543. Iron shot was cast in England as early as 1497 and the making of cannon balls was an incentive to the iron-casting industry. The Europeans were preceded by the Chinese in the use of cast iron by many hundred years, indeed cast iron was in general use in China by A.D. 500 and heavy castings were made by A.D. 1000.

The metal tools from ancient cultures show how the skills in using iron metal developed and suggest how the making of steel evolved. The first implements that may be regarded as the forerunners of modern steel were no doubt the inevitable result of carburization of reheated iron and the consequent transition from wrought iron to steel. Egyptian knives of the period from 1200 B.C. were carburized with some control. The Wootz steel from India with its damascene patterns is considered to be the first real steel and was exported from Mysore, India, before the beginning of the Christian era. The special laminated patterns of damascene steel of oriental origin are found in many swords and spears from as early as 600 B.C.; this steel was known and prized by the Romans. Complex forging techniques were known from about the third century A.D. to the Viking period and complex steel of this period has been recovered from Schleswig, Denmark, and from near Cambridge, England. Steel was produced in China as early as the second century B.C. and swords made by the forging of hard and soft wrought steel were used there in the third century A.D. The art of forging steel has been carried out in Europe from Roman time on, and Sheffield, England, was known for its manufacture of steel knives from metal imported from Sweden in Chaucer's day (1340-1400). The modern *Age of Steel* is usually dated from 1856 when Bessemer publicized his process for making steel in a crucible converter where air is blown in to oxidize the carbon, and the iron-manganese compound, spiegeleisen, is added to give the steel certain qualities. Siemens' open-hearth process was developed shortly after this and in 1878 Thomas and Gilchrist discovered that dolomitic limestone linings in converters were effective in extracting phosphorus and deleterious ingredients from the iron melt.

Most of the early iron and steel was smelted and treated with wood charcoal as a fuel and reducing agent. During the sixteenth century extensive cutting of forests in England to supply fuel to the iron industry caused considerable concern and was looked upon in some parts as a public nuisance. In some areas laws were passed prohibiting or controlling wood cutting for this purpose and acts of parliament prohibited the building of furnaces. England's oak was premium wood for shipbuilding and the increase in this industry led to some of the measures to control iron smelting. Various experiments with the use of pit coal as a source of fuel and carbon were carried out in Germany and England from about 1610 and the use of coal and coke to replace wood charcoal was adopted in England about 1710 with extensive use of coke by about 1745. The use of coal in iron production did not originate in Europe but was used on a small scale by the early Romans and by the Chinese. The proximity of coal and iron deposits has been a controlling factor in the economics of the iron and steel industry since about 1710 when coal and coke became essential to large-scale iron production.

Early Sources

Because the method by which the first iron metal made by man is not known, the ore that was its source is likewise unknown. Perhaps magnetite mixed with placer gold formed metallic iron as a slag in the melting and refining of the gold. Iron-rich sand deposits may have been an early source of iron in Mesopotamia, Asia Minor, and Egypt. Roasting and washing alluvial ores to concentrate the iron and remove the sulphur was carried out in the Caucasus Mountain area as early as 1000-1100 B.C. Widely distributed bog iron deposits in which the material is easily reduced by primitive methods are believed to have been the ore from which much of the early iron was procured. With the development of smelting techniques the carbonate ores would be suitable source material and ancient pits in iron carbonate rocks are found in many parts of the world. Specular hematite was used at Ur of the Chaldees, and in early Britain iron-rich clay bands and hematite were used in Roman times. The type of ore used in ancient times varied from place to place depending on what was available, and much ore was apparently collected from surface outcrops where little mining was required. Numerous pits on ore deposits indicate that open-cast mining was common and probably very little ore was mined from underground workings.

The first North American recorded discovery was by Raleigh, who reported iron deposits in Virginia in 1585. Several tons of ore from these deposits was smelted in England in 1608. The first smelting of iron ore in America began in Massachusetts in 1664, in Virginia in 1716, and in Pennsylvania in 1730. The discovery of iron deposits near Marquette, Michigan, in 1844 and in Minnesota in 1865, with production in 1854 and 1884 respectively, marked the beginning of the industrial period in the United States.

Development of Canadian Deposits

Champlain reported iron in Acadia in 1604 and other deposits were known as early as 1667. Iron was made at St. Maurice Forges, near Trois-Rivières, Quebec in 1733, and a small blast furnace was operated there from 1737 until 1883 when the local supply of bog iron ore was exhausted. An iron furnace for processing local ore was erected in Ontario at Furnace Falls, Leeds county as early as 1800. No further significant iron mining was done in Canada until the Blairton mine in Belmont township near Peterborough, Ontario, was brought into production in 1820. This mine, which was operated intermittently until 1875, provided more than 250,000 tons of ore.

In Nova Scotia, iron deposits in the Nictaux-Torbrook area of Annapolis county were discovered between 1820 and 1825, and were worked between 1825 and 1913. Furnaces operated at Clementsport and at Nictaux in the early part of that period. Limonitic ore discovered in the East River region, Pictou county, Nova Scotia, in 1828 led eventually to the production of at least 135,000 tons of ore from seven mines and later to the establishment of the Dominion Steel and Coal Corporation plant at Sydney. In the Londonderry region in Colchester county, the first mining was done in 1849 where over two million tons of ore was produced and where a charcoal blast furnace was operated from 1852 to 1875 and a steel plant from 1870 to 1877. It was at Acadia Mines near Londonderry that Siemens, mentioned previously, made his first commercial experiment in the direct conversion of iron into steel.

The Wabana ore on Bell Island, Newfoundland, was referred to as red sandstone by J. B. Jukes in a report on work done in 1842 and 1843, and references to red rock on the island go back to 1819. Mines were opened on the Wabana deposits in 1895 to supply much of the ore used at the plants at Sydney, Nova Scotia, and have been operated almost continuously since, the ore being shipped mainly to Nova Scotia, Britain, and Germany.

On the other side of the country, in British Columbia, iron deposits were discovered before 1873 and mines on Texada Island were worked between 1883 and 1907, the ore being shipped to Irondale, Washington, U.S.A.

A number of small iron mines in eastern Ontario and southern Quebec, some of which were worked before Confederation of Canada in 1867, operated round the turn of the century and much of the ore was exported to the United States. Following is a list of the larger mines:

Mine	Location	Operated	
		From	To
Chaffey	South Crosby tp. Leeds co.	1858	—
Matthews	North Crosby tp. Leeds co.	1860	1871 ?
Orton	Tudor tp. Hastings co.	1912	1913
Bessemer	Mayo tp. Hastings co.	1902	1913
Blairton	Belmont tp. Peterborough co.	1820	1875
Bluff Point	Bagot tp. Renfrew co.	1881	1901
Bristol	Bristol tp. Pontiac co., Que.	1872 reopened	1894 1958
Calabogie	Bagot tp. Renfrew co.	1883	1901
Coehill	Wollaston tp. Hastings co.	1884	1887
Forsyth and Baldwin	Hull tp. Ottawa co., Que.	1848	1858
Glendower	Bedford tp. Frontenac co.	1873	1888
Marmora	Marmora tp. Hastings co.	1955	present
Radnor	Grattan tp. Renfrew co.	1901	1907
Seymor	Madoc tp. Hastings co.	1837	1845
Victoria	Snowdon tp. Haliburton co.	1883	—
Williams	Bagot tp. Renfrew co.	1910	—
Belmont	Belmont tp. Peterborough co.	—	1914
Wilbur	Lavant tp. Lanark co.	—	1909

Although the deposits of this area are all small they were mined intermittently for about a century, but with development of the vast deposits of the Lake Superior region their operation became unprofitable. Revival of activity and interest in the area in the last decade has seen the reopening in 1958 of the Bristol mine, now called the Hilton mine, and the discovery in 1950 of the Marmora mine and its opening in 1955.

In the Michipicoten area, Algoma district, Ontario, the Helen mine was worked from 1900 to 1918 by the Algoma Steel Corporation during which time nearly three million tons of hematite ore was produced. The New Helen mine, opened in 1939, has been a producer of siderite ore.

Other iron mines operated in Canada at the beginning of the century include the Atikokan in Rainy River district, Ontario, operated from 1907 to 1912; the

Moose Mountain in Sudbury district, Ontario, operated from 1908 to 1915 and reopened as the Lowphos mine in 1959; the Bathurst in Gloucester county, New Brunswick, operated from 1910 to 1915; and the Magpie in Algoma district, Ontario, operated from 1913 to 1921. It was at Moose Mountain mine that iron concentrates were first briquetted, anticipating the present methods of treating low grade iron deposits by about 40 years.

With the development of Canadian sources of ore, Canadian processing plants were constructed. The first forging plant of the group of industries that eventually amalgamated as the Dominion Steel and Coal Corporation Limited with its steel plants at Sydney, Nova Scotia, was in New Glasgow, Pictou county, Nova Scotia, and was owned by the Nova Scotia Forge Company. This company has been the principal user of ore from Bell Island, Newfoundland, and of Nova Scotia coal for coke. The Algoma Steel Company was organized in 1901 and operations were started at a plant in Sault Ste. Marie, Ontario, in 1902, ore coming from the Helen mines in the Michipicoten area of Ontario and mines in the United States. The Steel Company of Canada Limited was formed in 1910 by the amalgamation of a number of smaller steel and iron processing plants, and has its blast furnaces and steel making units at Hamilton, Ontario. This company obtained the bulk of its ore from mines in the United States that it either owned or partly owned and by purchase of ore. More recently it has become involved in development of large iron deposits in Labrador, Newfoundland. Another large producer of steel in Canada is Dominion Foundries and Steel, Limited at Hamilton, founded in 1912. This company purchases ore from sources in Canada and the United States as required and has recently participated in the development of deposits in Canada. A number of smaller steel companies, situated mainly in the central part of the country, produce considerable amounts of diversified types of steel.

The development of major iron ore areas within the last 20 years has gained Canada an important position in the world iron ore trade. The rapid development of deposits in Canada has been particularly impressive when it is realized that no iron ore was produced in this country between 1924 and 1938. The beginning of the significant period of production in Canada was the opening of the sinter plant at Wawa, Ontario, to produce iron-oxide sinter from the siderite ore of the New Helen mine. The entry of Newfoundland (with Labrador) into Canada in 1949 brought with it important new iron ore production and reserves.

The discovery and development of large iron ore reserves at Steep Rock Lake, Ontario, was one of the vital steps in Canada's rise to importance in the international ore trade. As early as 1897, W. H. C. Smith and Wm. McInnes, geologists with the Geological Survey of Canada, reported hematite boulders in glacial deposits south of Steep Rock Lake and deduced that "an iron-bearing horizon with hematite of good quality, appears to be generally covered by the waters of the Lake" (McInnes, 1897)[1]. The presence of large orebodies under the lake was finally confirmed by diamond drilling in 1938. The ingenuity displayed in the engineering and geological

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[1] Names and/or dates in parentheses are listed in Selected Bibliography.

work that guided the development of this iron range is outstanding in Canadian mining history. Before the first ore was shipped in 1944 the Seine River had to be diverted, the lake dewatered, and up to 100 feet of silt and clay stripped from the bedrock by mechanical and hydraulic means.

The development of the iron ranges in Labrador and New Quebec is the most significant step in the establishment of the iron mining industry in Canada, and they will no doubt be one of the world's most important sources of iron for many years. The location and regional distribution of the iron-formations in the Labrador geosyncline was first indicated by A. P. Low of the Geological Survey of Canada in his reconnaissance of northern Quebec and Labrador. Systematic exploration of the iron ranges in the Knob Lake-Schefferville area on the height of land between Quebec and Labrador was started by private prospecting companies about 1936. By 1950 sufficient hematite-goethite ore of direct-shipping quality had been proven to warrant the development of the area, construction of 360 miles of railway north from the seaport of Sept-Iles, the establishment of docking facilities there, and the establishment of hydro power facilities to serve the industry. This development work was started in the autumn of 1950 and the first ore was shipped over the new railway in 1954. Facilities have been established to mine up to 13 million tons of hematite-goethite ore a year.

The southwestern part of the Labrador-Quebec belt is of much greater long range importance to the iron industry in Canada than the ore deposits near Schefferville, Quebec. Very large deposits of metamorphosed iron-formation suitable for beneficiation and production of high grade iron-ore concentrate are widely distributed throughout the southern part of the Labrador geosyncline. Iron ranges were first examined in the Wabush Lake area in 1929 but it was not until the last decade that the area from Wabush Lake to Matonipi Lake was explored in detail. The potential iron ore distributed in this 150-mile-long belt is estimated to be several billion tons of high grade concentrate. Very large mines and plants are established at Lac Jeannine and at Wabush Lake and the total annual production of ore concentrate planned for this area by 1964 is between 15 and 20 million tons. Ore production commenced in 1961.

Besides the development of these large areas a number of smaller iron mines have been active in Canada during the last ten years. Iron mining was resumed on the west coast with production from the Argonaut mine near Campbell River, British Columbia in 1951, from the mines on Texada Island in 1952, from Empire mine west of Port Hardy on Vancouver Island in 1957, and from the Nimpkish mine and a deposit on Head Bay on Vancouver Island in 1959. Small mines operated in eastern Canada include the Marmora mine which started production in 1955, the Hilton mine which was reopened and started production in 1958, the Canadian Charleson mine which started to recover hematite concentrate from glacial gravels in 1958, and Moose Mountain mine which was reopened for production of iron ore concentrate in 1959. Iron ore was produced as a by-product from sulphide ores at Port Robinson and then at Cutler, Ontario from 1954 to 1959, at

Copper Cliff, Ontario starting in 1956, and pig iron has been produced from titanium ores at Sorel, Quebec since 1950. A plant has been constructed at Trail, British Columbia for the production of iron and steel products by utilizing the pyrrhotite-rich tailings from base metal ores. This material will be sintered and processed in electric furnaces.

The domestic steel industry in Canada consumed about 6.5 million tons of iron ore in 1959 but during the same period nearly 22 million tons was produced and marketed from Canadian mines. It is expected that a very large percentage of iron ore produced in Canada will continue to be exported to meet increasing demands in the United States, Europe, and Japan. The production of iron ore in Canada has risen from less than 3.5 million tons to nearly 22 million tons within 10 years, and Canada has become the fourth largest producer in the world, being surpassed in 1959 by Russia, the United States, and France, and surpassing China, Sweden, West Germany, Venezuela, and the United Kingdom by 1.5 to about 7 million tons.

Chapter II

THE GEOCHEMISTRY OF IRON

The distribution of elements in the earth's crust and the kinds of minerals in which they occur are determined by the physical-chemical environment. Chemical systems in nature are usually much more complex than those simulated in laboratory experiments, and for this reason the geochemistry of mineral occurrences still depends to a great extent on empirical methods of investigation. It is often necessary to make deductions from very limited data, and principles of chemistry demonstrated by experiment may be inadequate to account for the results under conditions encountered in nature. Despite these limitations, a great deal is known about the chemistry of the earth and its role in geological processes.

Chemistry of the Ferride Elements

Chemical data pertinent to an understanding of the origin, distribution, and character of natural concentrations of iron are presented below as a necessary background for the discussion of various types of deposits. Because of similarities in chemical properties between iron and other elements in the ferride group, the chemistry of the entire group must be considered. Most ferride elements, if present in more than minimum amounts, are undesirable constituents in iron ore. Therefore, if an iron deposit is to be of ore quality the natural processes that led to its formation must have resulted in a high degree of separation of iron from these elements that have so close a chemical affinity.

Iron Isotopes and Ions

Pure metallic iron is silver-white, very ductile, and strongly magnetic; melts at 1,528°C, boils at 3,000°C, and has a specific gravity of 7.85 to 7.88 (20°C); its atomic number is 26 and atomic weight 55.85. It belongs to Group VIII of the periodic chart of the elements and is the chief element of a series with atomic numbers from 22 to 28, titanium, vanadium, chromium, manganese, iron, cobalt, and nickel, which are known as the iron-family or the ferrides. There are four stable isotopes of iron, Fe^{54}, Fe^{56}, Fe^{57}, Fe^{58}, and four other isotopes including Fe^{52}, Fe^{53}, Fe^{55}, and Fe^{59} with half lives of 8.2 hours, 8.9 minutes, 260 years, and 45.1 days, respectively. Isotope Fe^{56} is the most abundant, constituting 91.68 per cent of natural iron, the remainder being Fe^{54}, Fe^{57} and Fe^{58} making up 5.84, 2.17, and 0.31 per cent in that order. Two other isotopes, Fe^{60} and Fe^{61}, are known that have half lives of 3×10^5 years and 6.0 minutes respectively. Due to the electronic configuration of the atoms in the ferride series, electrons are added to an inner atomic shell as different atoms of the series are built up and the number of electrons in the outer shells does not change, except to provide valence changes. This property accounts for the similarity in the chemistry of the elements in the series. Native iron is rare; iron in most minerals is present as ferrous iron (Fe^{2+}) which is bivalent iron with an ionic radius of 0.74 Å, or as ferric iron (Fe^{3+}), trivalent iron with an ionic radius of 0.64 Å.

In geochemical and mineralogical studies, elements are classified as lithophile, chalcophile, siderophile, and atmophile, depending on the types of compounds they form; these in turn are controlled by the electronic configuration of the atoms. Berry and Mason (1959, p. 214) define lithophile elements as "those which ionize readily or form stable oxyanions (such as CO_3^{-2}, SO_4^{-2}, PO_4^{-3}), and in their compounds bonding is largely ionic in character; chalcophile elements ionize less readily, and thus tend to form covalent compounds with sulphur (and selenium and tellurium, when present); the siderophile elements are those for which metallic bonding is the normal condition, and which do not readily form compounds with oxygen or sulphur". Atmophile elements include the inert gases, nitrogen, and oxygen when distributed in the atmosphere.

Of the ferride group, all the elements except nickel and cobalt are strongly lithophile and form compounds with oxygen or oxyanions. Nickel and cobalt are strongly chalcophile, combining readily with sulphur, but also form lithophile compounds. Iron and manganese are chalcophile as well as lithophile and commonly occur as sulphides. Iron also has siderophile tendencies in nature and is occasionally found as native iron.

Iron in the Earth's Crust

Mineral deposits are abnormal concentrations of elements in the earth's crust but most geological processes lead to some concentration of certain elements. The average composition of the various types of rocks suggests the extent to which elements are concentrated by certain of these processes.

Ninety-nine per cent of the earth's crust consists of eight elements present in amounts greater than 1 per cent, and only four other elements present in amounts greater than 0.1 per cent. Iron ranks fourth in abundance; its amount and volume relative to the other main constituents are shown in Table I (taken from Mason, 1958, p. 46).

Table I

Major Chemical Elements in the Earth's Crust

	Weight %	Volume %
O	46.60	93.77
Si	27.72	0.86
Al	8.13	0.47
Fe	5.00	0.43
Mg	2.09	0.29
Ca	3.63	1.03
Na	2.83	1.32
K	2.59	1.83

Table II

Average Chemical Composition of the Lithosphere and Selected Rock Groups

Chemical Constituents	Lithosphere	Oceanic Regions	Suboceanic Regions	Young Folded Belt	Continental Shield	Igneous Rocks	Silicic Igneous Rocks	Intermediate Igneous Rocks	Subsilicic Igneous Rocks	Ultramafic Rocks	Av. all Sediments	Deep Ocean Sediments	Continental Shield Sediments	Young Folded Belt Sediments	Suboceanic Sediments
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
SiO_2	55.2	56.6	49.5	58.4	59.8	60.18	68.9	54.5	48.2	43.8	44.5	28.5	60.6	51.9	43.2
TiO_2	1.6	2.9	1.9	1.1	1.2	1.06	0.5	1.4	1.9	1.7	0.6	0.4	0.4	0.5	0.7
Al_2O_3	15.3	15.0	15.1	15.6	15.5	15.61	14.5	17.2	15.6	6.1	10.9	8.1	8.9	11.4	11.6
Fe_2O_3	2.8	3.8	3.4	2.8	2.1	3.14	1.7	3.2	3.0	4.5	4.0	5.0	2.4	2.6	4.6
FeO	5.8	8.0	6.4	4.8	5.1	3.88	2.2	4.6	7.8	8.7	0.9	—	1.2	2.0	0.6
MnO	0.2	0.2	0.2	0.2	0.1	—	0.07	0.16	0.17	0.18	0.3	0.6	—	—	0.3
MgO	5.2	7.8	6.2	4.3	4.1	3.56	1.1	3.2	8.2	22.5	2.6	1.8	2.9	3.8	2.4
CaO	8.8	11.9	13.2	7.2	6.4	5.17	2.6	5.8	10.5	10.1	19.7	30.5	10.6	12.6	21.1
Na_2O	2.9	2.5	2.5	3.1	3.1	3.91	3.9	5.1	2.6	0.8	1.1	0.8	0.8	1.3	1.1
K_2O	1.9	1.0	1.3	2.2	2.4	3.19	3.8	3.6	0.9	0.7	1.9	1.2	2.1	2.4	1.8
P_2O_5	0.3	0.3	0.3	0.3	0.2	0.30	0.16	0.39	0.30	0.30	0.1	0.2	0.1	0.1	0.1
BaO	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
CO_2	—	—	—	—	—	—	—	—	—	—	13.4	22.9	10.0	11.4	12.5
H_2O+	—	—	—	—	—	—	0.6	0.8	0.8	0.6	—	—	—	—	—
H_2O+ (105°C + CO_2)	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
S	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
SO_3	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
C	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Total Fe	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—

1 to 5. Determined by A. Poldervaart in "Chemistry of the Earth's Crust, from Crust of the Earth"; *Geol. Soc. Amer.* Sp. Paper 62, 1955, p. 133.

6. Based on 5,159 analyses as calculated by Clarke and Washington and listed by Brian Mason in Principles of Geochemistry, 2nd ed., John Wiley and Sons, Inc., New York, 1958, p. 42.

7. Determined by S. R. Nockolds based on 794 analyses and includes granites, adamellites, granodiorites, tonalites, and their volcanic equivalents; from Nockolds, S. R., Chemical Compositions of Igneous Rocks, *Bull. Geol. Soc. Amer.*, vol. 65, No. 10, p. 1032 (1954).

8. Determined by S. R. Nockolds based on 810 analyses and includes syenite, monzonites, mangerites, diorites, nepheline syenite, and monzonite and their volcanic equivalents; from Nockolds, op. cit.

9. Determined by S. R. Nockolds based on 721 analyses and includes gabbros, norites, anorthosites, troctolites, basalts, essexite, glenmuirite, theralite, teschenite and their volcanic equivalents; from Nockolds, op. cit.

10. Determined by Nockolds based on 182 analyses and includes perknites, dunites, and peridotites; from Nockolds, op. cit.

11 to 15. Determined by A. Poldervaart, op. cit., 1955, p. 132.

Chemical Constituents	All Sediments Except Deep Sea Group (Clarke)	Av. Shale	Av. Sandstone	Av. Limestone	Cleveland Ironstone	Wabana Iron-Formation	Biwabik Iron-Formation	Mesabi Taconite	Gunflint Iron-Formation	Cuyuna Iron-Formation	Gogebic Iron-Formation	Marquette Iron-Formation	Sokoman Iron-Formation	Sokoman Iron-Formation
	16	17	18	19	20	21	22	23	24	25	26	27	28	29
SiO_2	57.95	58.10	78.33	5.19	8.51	12.59	47.34	51.12	58.23	41.10	39.52	50.02	49.97	55.03
TiO_2	0.57	0.65	0.25	0.06	0.36	0.27	n.d.	small	tr.	0.04	0.15	n.d.	n.d.	n.d.
Al_2O_3	13.39	15.40	4.77	0.81	6.12	5.71	1.11	1.31	0.06	1.88	0.80	.97	0.92	0.17
Fe_2O_3	3.47	4.02	1.07	0.54	1.77	75.12	19.77	10.81	5.01	12.90	1.05	10.05	10.45	39.09
FeO	2.08	2.45	0.30	—	36.91	—	23.26	23.86	18.41	21.34	31.67	28.29	28.57	1.04
MnO	—	—	—	—	0.42	0.06	n.d.	0.46	0.25	2.42	.99	0.74	0.64	0.27
MgO	2.65	2.44	1.16	7.89	3.75	0.42	3.12	3.26	9.29	3.98	2.88	4.13	3.12	0.36
CaO	5.89	3.11	5.50	42.57	5.54	1.49	2.53	0.75	0.38	2.05	0.59	2.63	0.18	0.30
Na_2O	1.13	1.30	0.45	0.05	0.05	—	n.d.	small	n.d.	n.d.	n.d.	0.08	0.01	0.01
K_2O	2.86	3.24	1.31	0.33	0.03	—	n.d.	small	n.d.	n.d.	n.d.	n.d.	0.11	0.01
P_2O_5	0.13	0.17	0.08	0.04	1.30	1.63	n.d.	0.04	0.03	1.28	0.02	0.09	0.02	0.06
BaO	—	0.05	0.05	—	—	—	—	—	—	—	—	—	—	—
CO_2	5.38	2.63	5.03	41.54	20.70	—	.43	4.77	5.22	11.60	21.15	1.55	—	—
H_2O+	3.23	5.00	1.63	0.77	(4.05) (10.00)	—	small	3.43	2.01	1.53	.92	0.42	—	—
H_2O+ (105°C + CO_2)	—	—	—	—	—	—	—	—	—	—	—	—	2.62	0.66
S	—	—	—	—	0.05	—	—	—	0.07	—	—	—	0.001	0.015
SO_3	0.54	0.64	0.07	0.05	—	—	—	—	—	—	—	—	—	—
C	0.66	0.80	—	—	0.27	—	0.34	—	—	.17	n.d.	—	—	—
Total Fe	—	—	—	—	—	—	30.95	26.11	17.81	25.63	25.34	29.20	20.50	28.20

16. to 19. Determined by Clarke and listed by Pettijohn, F.J., Sedimentary Rocks, Harper and Brothers, New York (1949).

20. Sideritic-chamosite mudstone; Cleveland ironstone; Great Britain from Pettijohn, op. cit., p. 337.

21. Hematite-chamosite; Wabana, Dominion bed, zone 2, Ordovician, Newfoundland, from Pettijohn.

22. Composition of Biwabik iron-formation about 5 miles east of the Tower-Allen Junction Railway track, based on 22 analyses of drill-core by Mines Experiment Station of University of Minnesota and listed by Gruner, J. W., Mineralogy and Geology of the Mesabi Range; Office of the Commissioner of the Iron Range Resources and Rehabilitation, St. Paul, Minnesota, p. 62 (1946).

23. Composition of unaltered taconite of Main Mesabi Range, based on 14 analyses; from Gruner, op. cit., p. 63.

24. Sample of Gunflint Formation at the outlet of Gunflint Lake from Irving and Van Hise, 1892 and listed by Gruner, op. cit., p. 63.

25. Average of five analyses of ferruginous cherts and slates from Cuyuna Range from Harder and Johnston, 1918, listed by Gruner, op. cit., p. 63.

26. Average of four analyses of ferruginous cherts from Gogebic Range by Irving and Van Hise, 1892, listed by Gruner, op. cit., p. 63.

27. Four analyses of grunerite-magnetite schists from the Marquette Range from Van Hise and Bailey, 1897, listed by Gruner, op. cit., p. 62.

28. Unaltered silicate-carbonate iron-formation, from Ferriman mine area, Schefferville, Quebec, from paper by John Stubbins, Roger A. Blais, and I Stephan Zajac; Origin of the Soft Iron Ores of the Knob Lake Range, *Trans. Can. Inst. Min. and Met.*, vol. LXIV, 1961, pp. 37-52

29. Average analyses of three composite samples of fresh metallic iron-formation from the Knob Lake Range, Quebec, from Stubbins, *et al.*, op. cit.

The average weight per cent of other ferride elements in surficial rocks is titanium, 0.44; manganese, 0.1; chromium, 0.02; vanadium, 0.011; nickel, 0.008; and cobalt, 0.0023. The average weight per cent in crustal rocks of other elements that are undesirable constituents of iron ores is phosphorus, 0.118; sulphur, 0.052; zinc, 0.0065; copper, 0.0045; and arsenic, 0.0002 per cent or 2 parts per million.

The relative amounts of the major elements in different parts of the lithosphere and in different kinds of rock are given in Table II. The amounts reported vary somewhat, depending on the methods used in obtaining weighted averages for parts of the crust. The figures given in Table II are regarded as being of the right order of magnitude and relative abundance. The amount of iron and magnesium in the lithosphere as given by Poldervaart (1955) is higher than that indicated by Mason. This is apparently due to the fact that Poldervaart included more analyses of deeper parts of the earth's crust, which consist of rocks rich in ferromagnesian constituents, whereas the averages listed by Mason are based on rock types found at the earth's surface. The average amount of iron in igneous rocks (Table II) is only slightly higher than that for the whole crust. The average shale has less iron than the average igneous rocks and only slightly less than the average for the whole. A number of analyses are given of iron-formations or highly ferruginous sediments that represent a major source of iron ore or the protore from which it is derived. The iron content of this special group of rocks varies from about 20 to 50 per cent where they are unaffected by secondary enrichment, and is 4 to 10 times greater than the average iron content in crustal rocks.

Much of the iron in igneous rocks is in pyroxene, olivine, amphibole, or biotite where it is chemically combined with silica, alumina, or other constituents and is of little significance as iron ore. Iron present in oxide minerals, such as magnetite, hematite, or limonite, or in carbonate minerals, mainly siderite, is in a form suitable for use as ore. Magnetite and lesser amounts of hematite are disseminated in igneous rocks, and local concentrations may form ore deposits of considerable size and significance. The iron in many basic igneous rocks and anorthosites occurs as magnetite and hematite and is intimately associated with ilmenite and iron-titanium oxide. This type of occurrence serves as a source of iron ore only under special circumstances, as the iron cannot be easily separated from the titanium minerals. Other concentrations of magnetite within, or associated with, intermediate igneous rocks are usually free of titanium and provide some of the best quality iron ore. Much of the iron in sedimentary rocks occurs as magnetite, hematite, or siderite, The bedded ironstones or iron-formations are the richest in iron of all sedimentary rocks and, as they are predominantly chemical sediments, the geochemistry of iron is especially important in understanding their origin and composition.

Iron in Meteorites

The composition of meteorites has been studied very carefully with the belief that they may indicate the composition of other planets, or stars, and of the core of the earth. Meteorites consist of two main types, the iron-rich metallic type and the stony chondrites or achondrites. The stony meteorites are much the more common, possibly in the ratio of 5:1 according to Goldschmidt (1937), and consist principally of pyroxene, olivine, plagioclase feldspar, ferrous sulphide (troilite), and minor free iron or chromite. Iron in the stony meteorites is mainly in the ferrous state combined in silicate structures, and the proportion of ferric iron is much less in meteorites than it is in terrestrial rocks (Rankama, Kalervo, and Sahama, 1955).

The average composition of the iron meteorites is Fe, 90.5 per cent; Ni, 8.49 per cent; and Co, 0.57 per cent (Goldschmidt, 1954). The meteorites are made up of solid solutions of these elements or exsolutions of various alloys.

Iron in Sea Water

Iron is not one of the major dissolved constituents of sea water but is present in amounts ranging from 0.002 to 0.02 parts per million (grams per ton). The constituents that form 99.99 per cent of dissolved substances of sea water are listed below with average abundance in parts per million (*after* Mason, 1958):

Cl	—	18.980	K	—	0.380
Na	—	10.556	HCO_3	—	0.140
SO_4	—	2.649	Br	—	0.065
Mg	—	1.272	H_3BO_3	—	0.026
Ca	—	0.400	Sr	—	0.008

The amount of iron is also exceeded by N, Rb, Li, P, Ba, I, and As, in that order of abundance. Other elements of the ferride group are present in the following average amounts in parts per million: Mn, 0.001 to 0.01; Ni, 0.0001 to 0.0005; V, 0.0003; Co, 0.0001. Iron in true solution is present in less than 10^{-9} ppm and Ti and Cr in barely detectible amounts.

The soluble iron content of inshore waters along the State of Washington coast varies from 0.005 to 0.032 ppm with an average of 0.0178 ppm, according to Laevestu and Thompson (1958), and the content of particulate iron varies from 0.051 to 0.163 ppm with a mean of 0.114 ppm. The amount of particulate and soluble iron fluctuates widely, depending on seasonal variations and regional and depth distribution. As considerably more iron is carried by streams than is found in sea water, it is evident that practically all the iron delivered to the ocean must be precipitated in the sediments (*see* Rankama, Kalervo, and Sahama, 1955, p. 295).

Iron in Stream Water

The amount of iron carried by streams varies considerably depending on conditions such as the pH and Eh of the water, the amount of organic material present, the type of terrain being eroded, the climate, the amount of rainfall, the amount and kind of vegetation, and the population density of the river basin being drained. Because of these variables and the difficulty in obtaining representative samples of river waters from different depths and at regular intervals throughout the year, the amount of iron being transported by streams is imperfectly known.

Analyses of stream and lake water in the Upper St. Lawrence River-Central Great Lakes drainage basin in Canada show that the amount of iron dissolved in these waters is generally below 0.05 ppm and that the total iron is usually less than 1 ppm. Analyses from other major drainage basins in Canada indicate that the range in iron content in other streams is about the same. The amount of iron carried in colloidal suspension or in organic compounds is not precisely known because of sampling and analytical difficulties in dealing with these small quantities. Moore and Maynard (1929) report analyses from streams and lakes in the Canadian Shield showing the Fe_2O_3 content to range from 1.3 to 14.4 and a few as high as 61 ppm, but many analyses now available show much less. The average iron content in streams of North America as estimated by Clarke (1924) is less than 1 ppm and for the world less than 2 ppm, although some rivers in tropical areas, such as the Amazon, carry from 2 to 7 ppm.

Generally the amount of iron carried increases as the amount of organic constituents increases, and swamp waters are higher in iron than the streams that drain them. Further information on chemical factors affecting the solution and transport of iron are given in another section of this chapter.

It is concluded that iron is present in true solution in lake and stream waters in exceedingly small amounts and that relatively little information is available on the content of colloidal iron, organic-iron compounds, or suspended iron in streams and lakes.

Chemistry of Iron in Nature

Solubility

The chemical behaviour of ions in aqueous solutions and their solubility are dependent on a number of physical-chemical factors. The manner in which iron ions in particular are held in solution, transported, and precipitated to form part of sedimentary deposit depends on their ionic potential, the hydrogen-ion concentration (pH of the solutions), and the oxidation-reduction potential (Eh of the solution) in which they occur. Water and aqueous solutions are the natural media in which the solution and transport of iron takes place.

The water molecule, H_2O, with two very small hydrogen ions bonded to the larger oxygen ion is dipolar with a positive charge on the side where the hydrogen ions are bonded and negative on the opposite side. The water molecules are attracted to one another by this dipolar property and the properties of water are due to the nature of this ionic structure.

Ionic Potential

The ionic potential of an ion is expressed by a factor that is the ionic charge divided by the radius of the ion. As the ability of ions to attract and hold water molecules depends on their size and the intensity of their charges, ionic potential indicates much about the solubility of ions in solution. Of the two common ions of iron, the ferrous or bivalent ion is larger (0.74 Å) than the ferric trivalent ion (0.64 Å) but the ferric ion has a much higher ionic potential (4.7) than the ferrous ion (2.7). The ionic potential may be looked on as a measure of the intensity of the positive charge on the surface of the ion. This positive charge will repel the protons of the water molecule and where the positive charge is sufficiently great, protons may be detached resulting in the neutralizing of the charge of the central cation and the precipitation of an insoluble cation (Mason, 1958). It can be understood from this mechanism why ferric iron ions are decidedly less soluble than ferrous ions in an aqueous solution. It also explains why some ions of other elements with low ionic potential are more soluble and have highly soluble cations, why other ions of higher ionic potential predominantly form hydrolysates and others form soluble complex anions.

In brief, the solubility of ions is dependent on their ionic potential, which is a controlling factor in causing the separation of various ions in sedimentation processes.

Hydrogen-Ion Concentration

The extent to which a solution is acid or alkaline is defined by its hydrogen-ion concentration. The hydrogen-ion concentration in pure water at 20 degrees is 10^{-7} moles per litre, if greater than this the solution is considered acid and if less, alkaline. The hydrogen-ion concentration (pH) of a solution is shown as a logarithmic function ($pH = -\log H+$ conc.). The pH of pure water is 7, of sea water which is slightly alkaline 8 or greater, of river waters 7 or less, of rain water 6, and of some peat waters 4. Mason pointed out that a saturated solution of CO_2 at its partial pressure in the atmosphere has a pH of 5.2 and a solution of calcite in air saturated water has a pH near 8.

The pH of a solution is one of the most important factors in determining whether ions can be held in solution and what concentrations are in equilibrium. Ferric iron is much less soluble than ferrous iron, most ferric iron being precipitated when the pH rises to 3. Considerable amounts of ferrous iron are soluble in solutions with higher pH values and even in neutral solutions (pH 7). The solubility of both iron ions increases with an increased acidity (a decrease in pH value). For example, the solubility of iron in a solution with a pH of 6 has been shown to be about 10^5 times greater than in one with a pH of 8.5. It is evident from this that much of the iron carried in river waters (pH 7) will be precipitated when it reaches the sea (pH 8), which partly accounts for the fact that river waters bear an average of about 2 ppm iron and sea waters only 0.002 to 0.02 ppm.

Oxidation-Reduction Potential

The stability of an ion in one oxidation state depends on the amount of energy change involved in the addition or removal of electrons from its shell. The quantitative measurement of this energy change is recorded by a factor known as the oxidation-reduction potential (Eh), the standard of reference being the energy change involved in the ionization of a hydrogen gas molecule H_2, to $2H^+ + 2e$, to give hydrogen ions, the oxidation potential for this being regarded as 0.00.

Iron occurs in nature in three different states of oxidation with a valence of 0 in native iron, of 2^+ in the ferrous state, and of 3^+ in the ferric state.

The amount of energy involved in the removal of an electron from an ion for the oxidation of Fe^{2+} to $Fe^3 + e$ is a positive value of 0.77 (E_o in volts) but for the change of $Fe(OH)_2+OH^-$ to $Fe(OH)_3+e$ is a negative value of -0.56. The reduced form in any reaction couple has sufficient energy to reduce the oxidized form of any couple of higher potential and the ion in reduced form of lower potential becomes oxidized or loses an electron in the process. For example, $Mn(OH)_2$ with a negative Eh potential may reduce $2H^+$, and H_2 reduces Fe^{3+}.

Most chemical reactions of geochemical significance involve hydrogen and hydroxyl ions and as the oxidation potential varies with varying concentration of the reacting substance, it is obvious that variations in pH must be considered in conjunction with E_o values. E_o values are strongly affected by changes in pH and the oxidation potential decreases as the hydrogen ion concentration decreases (increasing pH values). As a result, oxidation takes place more readily in alkaline solutions. Other variables having a marked effect on the oxidation potential include temperature, atmospheric pressure with associated differences in the partial pressures of various contained gases, and especially variations in concentrations of different ions. The oxidation potential for ferrous iron to ferric iron in acid solutions is about 0.77 volts, but where the pH increases to the point where ferric hydroxide is precipitated, the oxidation potential drops rapidly to a negative figure and the oxidation of ferrous iron and the precipitation of ferric hydroxide proceeds nearly to completion. Ferrous ions are relatively stable in acid solutions and are only slowly oxidized by air. The deposition of ferrous compounds in nature therefore requires either a very acid environment or one with a low oxidation potential. The oxidation of ferrous to ferric iron is demonstrated very clearly in nature where groundwaters bearing ferrous ions in solution emerge as springs at the surface. Where they meet the oxygen of the atmosphere or stream waters of higher pH the ferrous ions become oxidized, relatively insoluble ferric hydroxide or oxide is precipitated, and rusty accumulations form in the soil or stream bed. Deposits of bog iron form in this way.

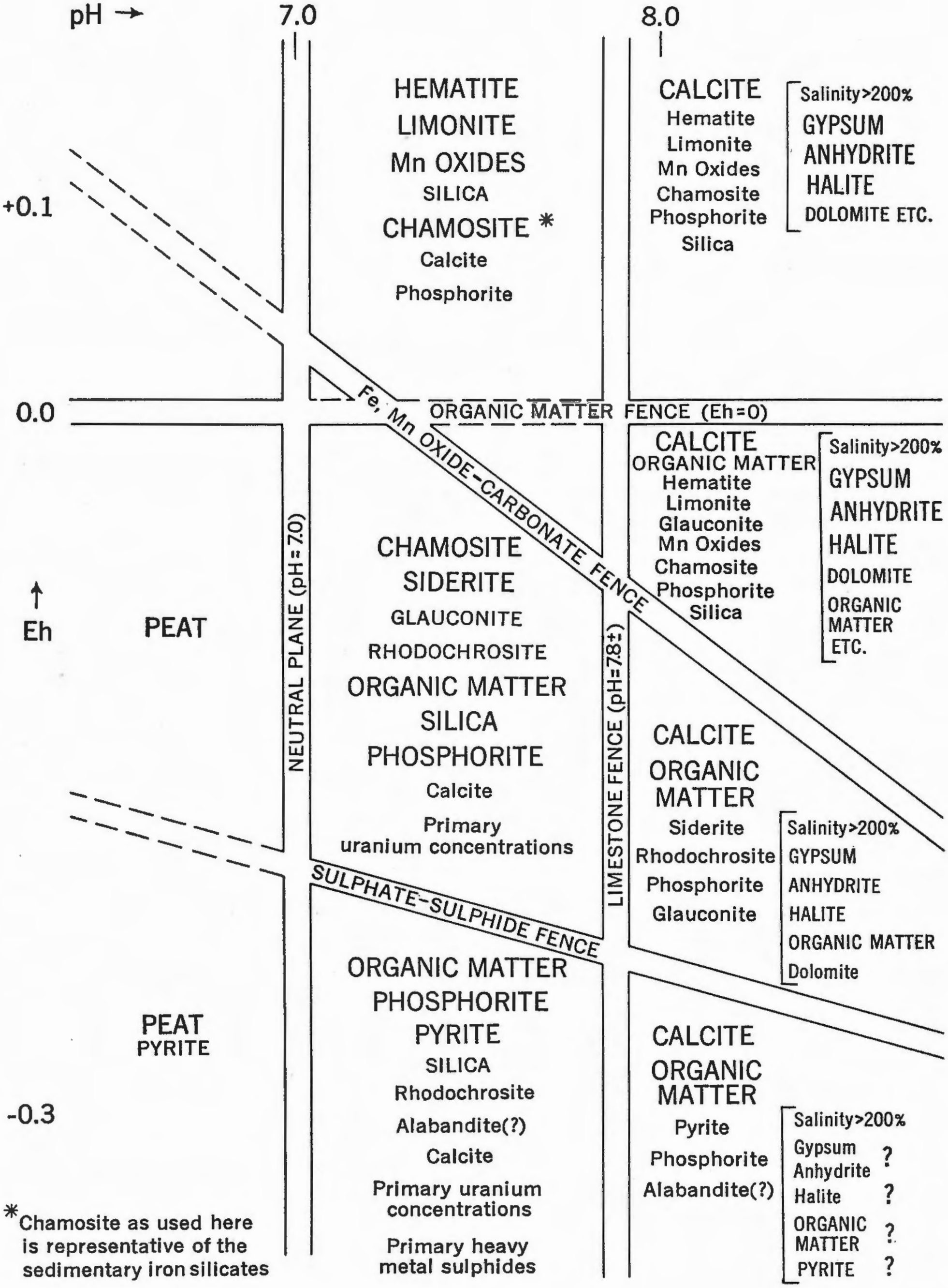
Closely related elements, such as those of the ferride group, become separated in nature because they are affected differently by variations in Eh and pH. Supergene processes involving solution, oxidation, and precipitation of iron and manganese often lead to a nearly complete separation of these elements. Iron and manganese occur in various proportions in many primary minerals in which their ions are interchangeable in the crystal structure. During weathering the iron is converted to relatively insoluble ferric iron by a much lower oxidation potential than that required to change manganese to manganese dioxide. The iron is therefore precipitated as ferric oxide or hydroxide much more readily than the manganese, which remains in solution longer and is precipitated only under stronger oxidizing conditions. In a similar way iron, cobalt, and nickel, found together in primary minerals, become separated because cobalt and nickel are less easily oxidized than iron and remain in solution after mildly oxidizing conditions have caused the precipitation of the iron. The higher oxide of nickel is not found in nature and cobaltic oxide is only formed under very strong oxidizing conditions. The conditions are well illustrated where ultrabasic rocks are weathered to form laterite, there most of the iron remains as a residual deposit and the relative amounts of cobalt and nickel are diminished.

The solution and migration of iron are affected greatly by the presence or absence of oxygen and carbon dioxide in solution. Ferrous iron is readily converted to ferric iron in the presence of atmospheric or dissolved oxygen and ferric iron is only carried in solution by waters with a pH less than 3. Where oxygen is absent ferrous iron is dissolved and carried by solutions bearing carbon dioxide. Soluble ferrous bicarbonate, $Fe\,(HCO_3)_2$, is formed where the partial pressure of CO_2 is adequate and will be precipitated as ferrous carbonate when carbon dioxide is removed from solution by diminishing atmospheric pressure, by an increasing temperature, or by its utilization by plants or other processes. Hem (1960) noted that if much bicarbonate is present the amount of iron dissolved may not change with change in Eh. Ferrous iron is also taken into solution when iron sulphides are oxidized and ferrous sulphate and sulphuric acid are formed, but these sulphates are oxidized in the atmosphere where oxygen is available and eventually the iron is oxidized and precipitated as ferric iron.

The solution of iron is also affected by a number of organic substances. Hem (1960) has shown that solutions of tannic acid at concentrations of 5 to 50 ppm reduce dissolved ferric iron to the ferrous state when the pH is less than 4. Solutions in nature are complex and usually contain several metallic ions, and a number of substances contribute to their pH and Eh. To predict the solubility and behaviour of any one substance, the concentrations of all substances must be known and the environment of the system defined. This is usually not possible for natural systems, but the physical-chemical conditions and known limitations for simple systems are of utmost importance in understanding and interpreting the genesis of natural mineral deposits. In brief, the solution and deposition of ionic substances in aqueous solutions is dependent on the ionic potential of the ions, the hydrogen-ion concentration, and the oxidation-reduction potential of the reacting substances.

Effect of Eh and pH on Sedimentation

The effect of the oxidation potential and hydrogen-ion concentrations on sedimentary environment can be appreciated more fully by examination of the 'fence' diagram prepared by Krumbein and Garrels (1952) and shown in Figure 1.



*Chamosite as used here is representative of the sedimentary iron silicates

GSC

FIGURE 1.

Origin and classification of the chemical sediments in terms of pH and oxidation-reduction potentials (*from* Krumbein, W. C., and Garrels, R. M., 1952, *J. of Geology*, vol. 60).

Changes in oxidation potential are shown as the vertical component of the diagram and changes in pH as the horizontal. Boundary lines or fences on the diagram define areas where a given list of minerals form and remain stable under the Eh and pH conditions indicated. From this diagram it is seen that hematite, limonite, manganese oxides, silica, and chamosite, will form when the pH of a system is between 7 and 8 and the oxidation potential is above 0.0, but if the Eh falls to a negative value in this pH range iron and manganese will probably form carbonates or silicates including siderite, rhodochrosite, glauconite, and so forth. If the Eh drops even lower and conditions are as represented below the sulphate-sulphide fence, iron will deposit as a sulphide where sulphur is available. In a similar way the diagram indicates that iron minerals are not likely to form where the pH is less than 7 except where the Eh is very low. This diagram is based on physical-chemical studies of a number of common reactions and the sedimentary environment is defined on the basis of Eh and pH. A study of the mineralogy of a natural deposit will reveal the general conditions in which it formed. James (1954) classified the various kinds of iron-formation in the Iron River District of Michigan as sulphide, carbonate, silicate, and oxide on the basis of the predominant mineral assemblages and the Eh and pH conditions of formation of these sedimentary facies can be inferred from Figure 1.

Transportation of Iron in Colloidal Form

Ferric hydroxide is one of a large number of substances that form colloidal suspensions in nature. Colloidal particles vary in size from 10^{-3} to 10^{-6} mm and elements in colloidal solutions grade from true solutions to colloidal suspensions. The relative amount of iron carried in true solution and colloidal suspension is not accurately known, but available data suggest that the amount of colloidal iron is several times greater than that of dissolved iron in most streams. Colloidal particles are electrically charged with some elements forming positively and others negatively charged particles. Ferric hydroxide particles may be either positively or negatively charged depending on the environment in which they form. Because of opposite charges being present, mixtures of different colloids may cause mutual deposition of oppositely charged particles if there is nothing present to stabilize the suspension. Iron and silica colloids on mixing will coagulate, according to Moore and Maynard (1929), unless they are protected by organic material. They found that 16 ppm of organic matter could hold 36 ppm of ferric oxide in suspension and that a mixture of 9.95 ppm colloidal ferric oxide, 29.85 ppm colloidal silica, and 4.5 ppm organic matter remained stable for two weeks in the presence of considerable quantities of electrolytes. Clarke (1924) showed that some river waters carry 12 ppm of organic matter which no doubt plays an important role in the stabilizing of ferric oxide colloids in stream waters. Moore and Maynard concluded from their experiments that ferric hydroxide is precipitated in a few days when colloidal iron and silica solutions stabilized by organic matter come in contact with electrolytes of the sea. Under similar circumstances silica is not completely coagulated after several months.

Factors in Crystal Chemistry

Except for native iron and sulphide compounds, practically all minerals of the ferride series elements are ionic compounds. The crystal structure of ionic compounds is determined mainly by the size of the ions and their electrical charge as indicated by their valence. The ionic radii of elements in the same group in the periodic table increase as the atomic number increases and in positive ions of the same electronic structure the radius decreases with increasing charge. In elements that exist in several valence states the size of the ion decreases with increases in positive charge. An appreciation of the relationship of ionic charge and ionic size or a number of common elements and their respective ions is shown in Figure 2.

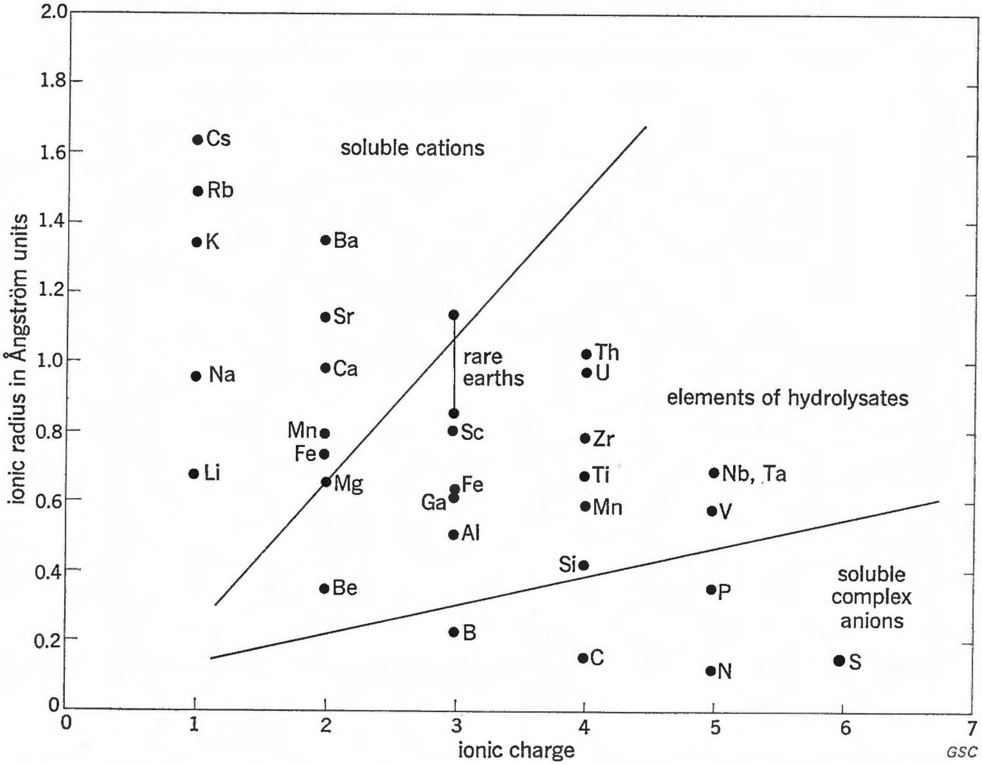


FIGURE 2.

Ionic potential and behaviour of elements in sedimentary processes (*from* Berry, L. G., and Mason, Brian, 1959, *Mineralogy*, W. H. Freeman and Company, San Francisco, California).

In ionic crystal structures it is essential that the overall electrical charge be balanced by an equal number of positive and negative charges on the ions.

The manner in which ions are put together or the packing arrangement of the ionic spheres is dependent mainly on the sizes of the ions incorporated in the crystal structure. Practically all the common minerals of ferride elements, except the sulphides, are oxygen compounds. The oxygen ion O^{-2} has a radius of 1.40 Å and is considerably larger than the ions of common metals. Because of the large size of the oxygen ions they determine the general structural pattern of the mineral. The cations, because of their decidedly smaller size, fit into the interstices of the oxygen ions, and the number and arrangement of the oxygen ions grouped around a cation depend on the size of the cation relative to that of the oxygen anion. The ratio of the size of the cation to that of the anion, known as the radius ratio, is a controlling factor in determining what geometrical arrangement of ions forms stable mineral structures. The coordination number of a cation shows the number of anions that can be arranged around it in a stable fashion. Most of the cations of the ferride series of elements have sixfold coordination with oxygen anions except for Mn^{4+}, V^{5+} and Cr^{6+}, which are smaller and have fourfold coordination with oxygen.

The atomic structures of ionic minerals are built up by the combination of anions and cations in appropriately coordinated arrangements that satisfy the factors in crystal chemistry outlined. In simple oxides the basic units of the crystal are the oxide anions and metallic cations, but silicate structures are more complex. The basic unit in silicates is the silica tetrahedron, consisting of four oxygen anions arranged at the points of a tetrahedron in fourfold coordination around one silica cation. These tetrahedrons are combined in various ways to permit other cations to be arranged in proper coordination in remaining interstices between oxygen ions.

Factors of crystal chemistry are a fundamental guide to understanding what minerals are stable under given circumstances, the extent to which cations are mixed or substituted in a specific mineral (isomorphism, ionic substitution, solid solution), and the manner in which small amounts of substances or trace elements are distributed in mineral species. The order and form in which elements crystallize from a magma and the manner in which they are distributed in stable minerals are governed mainly by laws of crystal chemistry.

Iron in Igneous Rocks

Table II shows that iron, magnesium, and manganese are most abundant in the ultramafic rocks and decrease in abundance as rocks become more silicic. Likewise cobalt, nickel, chromium, vanadium, and titanium are more abundant in basic igneous rocks than in silicic or granitic rocks. Iron, together with lesser amounts of cobalt and nickel, is segregated in sulphide minerals and with chromium and titanium in oxide minerals in the early stages in the differentiation of basic magmas.

The extent to which oxidation has taken place in the magma is significant in determining whether magnetite containing some ferric iron can form and become segregated as layers or disseminations in gabbroic rock. The availability of sulphur and its vapour pressure in relation to temperature, and the state of oxidation of the iron, control the segregation of iron as sulphides in other gabbroic rocks, as outlined by Goldschmidt. In magmas where elements are in a low state of oxidation the iron is present as ferrous iron. It combines with early formed minerals including chromite, spinel minerals, and ilmenite and forms a major constituent in ferrous silicates such as olivine and pyroxene in dunites, pyroxenites, norites, and gabbros. When oxidation proceeds in a magma a limited amount of ferric iron is formed, and magnetite and ilmenite crystallize and increase in abundance as oxidation continues. Ferric iron as hematite will predominate only in rocks crystallized from a magma that was in a very advanced stage of oxidation. The degree to which elements are oxidized in a magma seems to be the controlling factor in determining whether in the early stages of crystallization a large part of the iron is segregated as oxide minerals.

Landergren (1948, p. 47) showed that the degree of oxidation increases as the ratio of quartz to $SiO_2 + CaO + MgO + FeO + MnO$ increases in igneous rocks, indicating that gabbros and basalts generally have a lower degree of oxidation than diorites, and diorites have lower oxidation than granites. This variation in degree of oxidation is not attributed to conditions of differentiation primarily, but to the addition of oxygen to a magma from external sources through assimilation of sedimentary rocks or from surface sources. The degree of oxidation is reflected directly in the ferric to ferrous iron ratio of the rock. The ferric iron content may not actually increase, as Wickman (1943) showed that this does not change greatly during differentiation, but the ferric-ferrous ratio may change because ferrous iron is removed from the magma.

Ferrous and ferric iron ions occur in different ways in igneous rocks due mainly to the differences in size of the ions and valence charges (Fe^{2+} – 0.74 Å, Fe^{3+} – 0.64 Å). As oxidation takes place most of the ferric iron is combined with ferrous iron to form magnetite and there is usually enough ferrous iron present to cause magnetite rather than hematite to crystallize. Some of the ferric iron is taken into the silicate mineral structures where it can substitute to a limited degree for aluminum cations Al^{3+}. The radius of Al^{3+} is 0.51 Å and because of the difference in ionic size the exchange of Fe^{3+} for Al^{3+} is limited and takes place only in the sixfold coordination positions in the lattice. Ferrous ions are closely associated with magnesium and their substitution or exchange for magnesium ions in the silicate structures is continuous and almost unrestricted.

Early differentiates of a magma are very high in magnesium, which is much more abundant than iron, but as differentiation continues the Fe:Mg ratio changes and iron predominates. The Fe:Mg ratio as given by Rankama, Kalervo, and Sahama (1955) for dunite, gabbro, diorite, and granite rocks respectively are 0.2, 1.5, 2.4, and 4.9 illustrating the increase of iron relative to magnesium. As the radius of ferrous iron ions is 0.74 Å and that of the magnesium ions 0.66 Å, the magnesium compounds have a stronger bonding and a higher melting point than iron and are taken up in the early formed minerals, before iron.

Other Ferride Elements in Igneous Rocks

Chromite is associated with iron in igneous rocks and is most abundant in the magnesium-rich silicates, olivine and pyroxenes, that rcystallize in basic and ultrabasic rocks. The ionic radius of Cr^{3+} is close to that of Fe^{3+} and Al^{3+}. Chromium is present in primary minerals such as chromite and other spinels.

Titanium-rich mineral segregations, which occur where oxidation of the magma is advanced to a considerable degree, are found in gabbros and norites and in plagioclase-rich rocks such as anorthosites. Iron and titanium are intimately associated in basic igneous rocks where ilmenite and magnetite form large segregated masses. In the more silicic igneous rocks the iron and titanium occur in separate minerals and titanium is much the less abundant. Iron-titanium minerals are resistant to weathering, but when broken down titanium usually remains as a residual with aluminum and is separated from iron.

The common vanadium ion in magmatic rocks is trivalent and is closer to ferric iron in ionic radius than to trivalent chromium and aluminum. Because of this close similarity in size to ferric iron, vanadium is distributed mainly with ferric iron in magmatic magnetite. In ilmenite-magnetite mixtures vanadium is predominant in magnetite. Some vanadium occurs in chromite and in other chromium minerals but is absent or in trace amounts in aluminous silicates. The amount of vanadium decreases in the silicic igneous rocks. In weathering, vanadium becomes oxidized and in this form is rather soluble. It becomes separated from iron and is precipitated again as the oxide under strong reducing conditions.

Manganese and iron are closely associated in igneous rocks but manganese increases in amount relative to iron in the more silicic rocks and in pegmatites. Manganese is most abundant in the ultrabasic rocks but does not vary greatly in amount in the other igneous rocks except in the highly silicic rocks. Although manganese may form separate mineral species in some igneous rocks, most of it is combined in other minerals and is closely associated with ferrous iron. Minerals low in iron are also usually low in manganese. The geochemistry of manganese is very similar to that of iron except that it is not differentiated to the same extent in igneous rocks and is more soluble than iron in most weathering processes.

Cobalt and nickel have a high affinity for sulphur and have ionic radii close to that of ferrous iron and magnesium thus they are most abundant in the sulphide masses associated with peridotite and norite rocks. Nickel is mixed with ferrous iron in the nickel-iron sulphide pentlandite, and some is present with iron in pyrrhotite. Cobalt is present in excess of nickel in pyrite regardless of its temperature of formation. The Co:Ni ratio in pyrrhotites varies from 0.02 to 0.07 but in late magmatic sulphides and in vein deposits, where pyrite is usually predominant, cobalt is present in excess of nickel. Landergren (1948) found that nickel exceeds cobalt in titaniferous iron ores and that both are enriched in this type of early segregation.

The bulk of the cobalt and nickel in igneous rocks is, however, taken up in silicate minerals and is most abundant in the iron-magnesium-rich rocks and much less abundant in the silicic rocks. The ionic radius of Mg^{2+} is 0.66; for Ni^{2+}, 0.69; for Fe^{2+}, 0.74; and for Co^{2+}, 0.72; from this it would be expected that nickel would follow magnesium closely and cobalt would follow ferrous iron. It has been found however that there is no direct linear relationship in the amount of nickel with magnesium but that there is a direct relationship between cobalt and magnesium. It appears that ionic size is not the dominant factor in controlling this relationship in the silicate minerals. Much of the nickel and cobalt is taken up in early formed olivine and hypersthene in igneous rocks and in these minerals the nickel content does vary directly as the magnesium content. In silicic igneous rocks both elements are much less abundant and cobalt is three or four times as abundant as nickel. In weathering, nickel tends to remain as a residual in hydrosilicate minerals including some clay minerals.

When Ni^{2+} is taken into solution in nature it is stable and is transported in streams, but only very little of the nickel in the source area is taken into solution. Cobalt is usually separated from nickel during weathering. Some is taken into solution at that time and some is transported as colloidal particles. Cobalt is commonly abundant in manganese oxide deposits and in bog iron deposits. It is enriched in laterite deposits but not to the same extent that nickel is enriched. Landergren (1948) indicated that cobalt exceeds nickel in amount in the banded iron rocks of Great Britain and in most of the other sedimentary iron-formations investigated. Cobalt is more abundant than nickel in the various skarn iron ores, but the reverse relationship was determined in apatite-rich ores.

Only limited study has been made on the distribution of the ferride elements in iron ores of Canada and most of this information is included in the discussions of iron ore genesis. Landergren (1948) gave considerable detail on the relationship of iron and ferride elements in his work on the geochemistry of Swedish iron ores.

Chapter III

MINERALOGY OF IRON ORES

The composition, physical structure, and amenability to concentration of an iron ore are properties that depend on the mineral content of the ore. Moreover the geological processes and genetic factors that led to natural concentration of iron in ore deposits can to some extent be deduced from the minerals present in the ore and gangue. The mineralogy of most iron ores is relatively simple, the iron being present in one or more of only five major minerals. Iron, an abundant element, occurs in small amounts in a great many minerals, most of which are of no importance as iron ore. Numerous gangue minerals are commonly associated in iron ores, and some of these contain deleterious constituents that may be of greater significance in evaluating an ore than the iron minerals themselves.

A simple classification of iron ore based on mineralogy alone would include hematite, magnetite, magnetite-hematite, hematite-goethite, hematite-siderite-chamosite, siderite-pyrite, and other less common mineralogical types. Some of these minerals, such as hematite, occur in various physical forms ranging from earthy fine-grained aggregates to coarsely crystalline masses, and the mineral habit and texture of an iron ore are usually as important as composition in determining ore qualities for smelting. Iron ores are accordingly classified, for commercial purposes, on the basis of chemical composition and structural character. Classifications based on mineralogy and geological features are useful for descriptive purposes and for studies on the origin of occurrences.

As a basis for discussion, the physical and chemical properties and mode of occurrence of the common iron minerals listed in Table III, are reviewed below.

Native Iron

Native iron, which is rare in terrestrial deposits, is a common constituent in meteorites. Iron crystallizes in the isometric system forming cubes or octahedrons, but is commonly present in nature as blebs or irregular masses. Twinning is reported on (111) and (112) planes and a cleavage on (001). Native iron is malleable, has a hardness of 4, specific gravity 7.3-7.87, metallic lustre, steel-grey to black, and is opaque and magnetic. Terrestrial varieties usually have less than 3 per cent nickel, meteoric varieties have more. Other elements including Co, Cu, Mn, C, P, S, and Si are usually present in small amounts.

Table III

Common Iron Minerals

(Principal ore minerals listed in **bold** type)

Mineral	Composition	% Iron	Specific Gravity	Hardness	Colour
NATIVE IRON	Fe	100	7.3-7.87	—	Steel-grey
OXIDES					
Hematite (alpha)	α-Fe_2O_3	69.9	5-6	6.5 –	Steel-grey to red
Martite	(hematite pseudomorphous after magnetite)				
Turgite	$2Fe_2O_3 \cdot H_2O$	66.1 ±	4.2-4.7	6.5 –	Brown to red
Goethite (Limonite)	$HFeO_2$	62 ±	3.3-4.3	5.5 –	Brown to yellow
Lepidocrocite	$FeO(OH)$	62 ±	3.3-4.3	5	Brown, reddish brown
Ilmenite	$FeTiO_3$	29.4	4.72	5-6	Black, brown-black
Maghemite (gamma)	γ-Fe_2O_3	69.9	4.88	5	Brown
Magnetite	Fe_3O_4	72.4	5.17	5.5-6.5	Black, blue or brown-black
CARBONATES					
Siderite	$FeCO_3$	48.2	3.96	3.5-4	Ash-grey to brown
Dolomite	$CaMg(CO_3)_2$	26	2.8-3.1	3.5-4	White, grey, brown
Ankerite	$CaFe(CO_3)_2$				
Kutnahorite	$CaMn(CO_3)_2$				
SULPHIDES					
Pyrite	FeS_2	46.5	5.01	6-6.5	Brass-yellow
Marcasite	FeS_2	46.5	4.8	6-6.5	Light brass-yellow
Pyrrhotite	$Fe_{(1-x)}S$	63-61	4.5-4.6	3.5-4.5	Bronze-yellow
SILICATES					
Chamosite	$(Mg, Fe, Al)_6(Si, Al)_4O_{14}(OH)_8$	33-42 ±	3-3.5	—	Green to light yellow
Glauconite	$K(Fe, Mg, Al)_2(Si_4O_{10})(OH)_2$	varies	2.5-2.8	—	Dark green, black, earthy
Greenalite	$(OH)_{12}Fe''_9Fe'''_2Si_8O_{22} \cdot 2H_2O$	varies	3	—	Olive-green, light greenish
Minnesotaite	$Fe_3(OH)_2Si_4O_{10}$	varies	3-3.2	—	Pale green, yellow-green
Stilpnomelane	$(OH)_4(K, Na, Ca)_{0-1}(Fe, Mg, Al)_{7-8}Si_8O_{23-24} \cdot 24H_2O$	varies	2.7-3	—	Dark greenish
Grunerite	$Fe_7Si_8O_{22}(OH)_2$	varies	3.2-3.5	—	Light green-brown
Cummingtonite Series	$(Fe, Mg)_7Si_8O_{22}(OH)_2$				
Anthophyllite	$(Mg, Fe)_7Si_8O_{22}(OH)_2$	varies	2.9-3.4	—	White, grey-brown
Hypersthene	$(Fe, Mg)SiO_3$	varies	3.4-3.5	5-6	Dark green-brown

According to Dana (1947), native iron occurs in igneous rocks especially in basalts, as in Greenland, and in carbonaceous sediments mixed with limonite and organic matter, as in Missouri. In igneous rocks it is thought to have been reduced when carbonaceous material was assimilated. Native iron, which occurs at a number of places around the world, is reported in Canada as minute spherules in feldspar at Cameron township, Nipissing district, Ontario, and as a thin crust on quartzite on the north shore of St. Joseph's Island, Lake Huron, Ontario.

Reference: Palache, Charles, *et al.*, Dana's *System of Mineralogy*, vol. I, 1944.

Iron Oxide Minerals

Hematite (α-Fe_2O_3)

Hematite, the most abundant iron ore mineral, is present in a variety of geological conditions. Where fine grained it is deep red, bluish red, or brownish red and may be soft and earthy to unctuous, ocherous, compact or highly porous to friable, or granular, or may form dense hard lumps. Considerable siliceous or argillaceous impurities are common. Fine-grained red hematite may occur in smooth reniform masses (kidney ore), in botryoidal or stalactitic shapes, or be columnar, fibrous, radiating, or platy. Coarser crystalline hematite is steel-grey with a bright metallic to dull grey lustre and occasionally coarse crystals have a deep bluish to purplish iridescent surface. Coarse-grained hematite is known as specularite or specular hematite, and may form blocky or platy crystals with a strong micaceous parting. The red streak is difficult to observe on this variety. Earthy varieties are very soft but crystals have a hardness of 5 to 6 and a specific gravity of 5.256. Hematite melts between 1,350° and 1,360°C, is paramagnetic with a relative magnetic attractability of 1.32 compared to 100 for metallic iron, or a magnetic susceptibility of 0.0011 compared to 3.07 for magnetite.

Crystals are rhombohedral with an R type lattice and 32m symmetry. The unit cell dimensions are $a : C = 1 : 2.7307$ and each unit cell contains 6 (Fe_2O_3) with oxygen atoms arranged approximately in hexagonal close packing with these close-packed layers parallel to (0001). Cations lie between these layers each coordinated to six oxygen atoms but only two thirds of these positions are filled and each cation is in octahedral coordination with three oxygen atoms above and three below. Hematite does not have cleavage but has good parting parallel to (0001) and occasionally on (0112). Penetration twins occur on (0001) and lamellar twins on (0112). Crystals of hematite may form cube-like rhombohedrons with striatons on rhombohedral faces, or they may be tabular to platy and grouped in rosette-like clusters. The basal or *c* face may be striated parallel with (0112) face edges. Fractures may be subconchoidal to uneven.

Ideally hematite contains 69.94 per cent iron and 30.06 per cent oxygen. Small amounts of Ti may be present in the lattice, and it has a limited miscibility with ilmenite ($FeTiO_3$). This mineral may show deficiencies in oxygen according to the formula or contain some ferrous iron.

In nature, hematite occurs under many different circumstances, but its stability depends on the presence of oxygen (Edstrom, 1953). It is most commonly found in zones of weathering and oxidation but small quantities occur in various kinds of igneous rocks. In anorthositic and gabbroic rocks it is a common associate with ilmenite and occurs as exsolution lamellae. It is present in igneous rocks where there is an apparent deficiency in ferrous oxide (in this environment any ferrous oxide leads to the formation of magnetite), and is a common product of fumarolic activity associated with volcanism. In zones of weathering hematite may be derived from siderite, magnetite, or iron-silicate minerals or be deposited by meteoric solutions. Specular hematite is most common and widely distributed in metamorphosed sedimentary iron-formations. Hematite is a primary mineral in sediments being precipitated in some environments in granules and oölites mixed with siderite, chamosite, or silica.

References: Berry, L. G., and Mason, Brian, 1959; Edstrom, J. O., 1953; Edwards, A. B., 1949; Flaschen, S. S., and Osborn, E. F., 1957; Geith, M. A., 1952; Greig, J. W., *et al.*, 1935; Huber, N. King, 1958; Mason, Brian, 1943; Palache, Charles, *et al.*, Dana's *System of Mineralogy*, vol. I, 1944.

Martite

The name martite is now commonly used for hematite that is pseudomorphous after magnetite and, rarely, after pyrite. It is usually iron-black with a submetallic lustre, or red to brownish red; both have a reddish to purplish streak. Martite, which may be feebly magnetic due to residual magnetite inclusions, occurs in many different environments but mainly in strongly oxidized zones overlying magnetite occurrences. On the Lake Superior iron ranges and around Schefferville-Knob Lake in Quebec and Labrador, it is a common constituent in the hematite-goethite ores and altered zones adjacent to these ore deposits. In this report the writer prefers to restrict the name martite to material in weathered and oxidized zones near the earth's surface. It is not a desirable name for blocky specular hematite associated with magnetite in the high rank metamorphic rocks.

References: Gruner, J. W., 1922, 1926; Weinert, F., 1933.

Turgite ($2Fe_2O_3 \cdot H_2O$)

Turgite has physical properties similar to earthy red hematite and a specific gravity of 4.2-4.7. It gives an X-ray pattern identical to hematite and is considered to be finely divided hematite with some adsorbed and capillary water.

Goethite $HFeO_2$ (Limonite)

Goethite is one of the most common iron ore minerals, certainly the most abundant of the hydrous ferric oxide group. It is usually brown to ocherous yellow but may be black or dark brown to reddish brown. Most of the brown iron oxides of indefinite composition and water content, referred to as limonite, are goethite with variable amounts of adsorbed or capillary water. The descriptive information given here pertains to goethite as well as to most of the limonitic material.

Goethite has a brown to brownish yellow streak, may be dense and massive or porous, earthy, and friable, and may contain siliceous and argillaceous impurities or be intimately mixed with hematite. It occurs in dense, botryoidal, reniform, stalactitic, pisolitic, or oölitic masses which have internal radiating or acicular structures and an external metallic lustre. Massive material has a hardness of 5-5½, a specific gravity between 3.3 and 4.3. The relative magnetic attractability of limonite as given by Taggart (1948) is 0.84 compared to 100 for iron and the magnetic susceptibility is 0.0008 compared to 3.07 for magnetite.

Goethite crystals are orthorhombic with symmetry designated 2/m 2/m 2/m. Crystals may be striated parallel with the C axis or may be tabular scaly or acicular and have perfect cleavage along (010) planes, and less perfect along (001), with uneven fractures.

Most of the excess water is driven off at temperatures below 100°C and goethite decomposes in pure water at 130° ± 5°C. The differential thermal curve for goethite has a single endothermic peak at 395°C ± 10°C; this peak temperature is lower for very finely divided material. Most analyses indicate 89 to 90 per cent Fe_2O_3, and about 10 to 10.25 per cent H_2O. Small amounts of manganese are usually present and up to 5 per cent Mn_2O_3 may substitute for iron in the lattice. Silica impurities are found in most samples.

Goethite forms under oxidizing conditions at ordinary temperatures from the weathering of many different iron minerals but most easily from siderite, pyrite, iron-bearing sulphide minerals, magnetite, and fine-grained iron-bearing silicates. It is abundant in various gossans and cappings overlying sulphide bodies, in laterite, and in residual deposits formed by weathering processes. It is deposited in bogs, where spring waters come in contact with the atmosphere or groundwaters meet an oxygen-rich environment. It is also deposited in marine sediments as a primary constituent by inorganic and biogenic processes. It is especially important as an ore mineral in residual ores derived from iron-formation through oxidation of iron minerals and leaching of silica, and in laterites and various residual deposits. In bog ores it is the main constituent, and commonly replaces plant debris and wood fragments. In many bog deposits, concretions, or veins it is associated with manganese oxides, clays, carbonate, and silica. Much of the ferric iron oxide transported in colloidal suspension changes to goethite after deposition and aging. At room temperatures, X-ray patterns are obtained from artificial preparations after months of aging, but lines are obtained after an hour when the temperature is raised to 100°C. By dehydration goethite changes to hematite.

A number of substances that are actually goethite are mentioned in the literature under various mineral names. Some of these are xanthosiderite ($Fe_2O_3 \cdot 2H_2O$), ehrenwerthite (colloidal variety), hydrogoethite, brown hematite, and yellow ochre. Others of less definite composition or containing considerable impurity are limonite, kaliphite, hypoxanthite, melanosiderite, avasite, and esmeraldaite.

References: Geith, M. A., 1953; Kulp, Laurence J., and Trites, Albert F., 1951; Smith, F. Gordon, and Kidd, Donald J., 1949; Taggart, A. F., 1948.

Lepidocrocite FeO(OH)

Lepidocrocite has the same composition as goethite, practically the same physical characteristics, and occurs in the same environment but is much less common. It is usually deeper brown than goethite and commonly reddish brown, and has a dull orange streak. Crystals are orthorhombic with a 2/m 2/m 2/m symmetry and are usually platy or flattened and may be bladed, fibrous, acicular, or micaceous forming rosettes, plumose or concentric structures. The crystal structure is different from that of goethite, the hydrogen atoms being in discrete OH groups whereas in goethite hydrogen atoms act as cations between oxygen atoms in twofold coordination. The arrangement in goethite is more stable and as a result in thermal decomposition, lepidocrocite has an endothermic peak at about 357°C whereas goethite has a corresponding peak at about 395°C. On decomposition lepidocrocite yields γ-Fe_2O_3, maghemite.

References: Bragg, W. L., 1950; Kulp, Laurence J., and Trites, Albert F., 1951.

Ilmenite $FeTiO_3$

Ilmenite belongs to the same group of minerals as hematite and corundum but has a lower rhombohedral symmetry and forms a separate series of minerals. There is more or less complete miscibility between Fe and Mg members and between Fe and Mn members. The other two minerals of the group tolerate very little ionic substitution compared to the ilmenite series. Ilmenite is commonly associated with magnetite, less frequently with hematite.

Crystals may be thick tabular to blocky rhombohedral, but are usually found in compact massive irregular grains. Thin lenticular microscopic blades are oriented parallel with crystal planes in magnetite. Twinning occurs on (0001) and on (1011) planes. Fractures are conchoidal to subconchoidal. Cleavage is absent in ilmenite but parting is parallel with the twinning planes; it is present in the magnesium and manganese-rich members of this series. The specific gravity is 4.72 but less in varieties containing manganese and magnesium. Ilmenite is black to brownish black with a metallic to submetallic lustre and a black to brownish red streak. It is uniaxial negative, transmits only red light, and appears greyish white in polished section with weak pleochroism. Exsolved inclusions of magnetite, hematite, or rutile are oriented parallel with certain crystallographic planes.

As much as 54 atomic per cent of Mg may substitute for iron with gradation to geikielite in which the Mg:Fe ratio is 8:1. In some pyrophonite ($MnTiO_3$) members the Mn:Fe ratio is 5:7. Not more than 6 per cent Fe_2O_3 is likely to remain in solid solution in ilmenite at ordinary temperatures.

Ilmenite, which is widely distributed in black sand deposits, is present as an accessory mineral in igneous rocks of widely different compositions but is most abundant in gabbro and gabbroic anorthosite rocks where it is associated with magnetite and hematite. Most frequent occurrences are in the large injections, disseminated and vein deposits in anorthosite rocks and are well known in the Grenville Province of the Canadian Shield. It is associated with other iron and titanium oxides and with sulphide minerals, spinel, apatite, and biotite in plagioclase-pyroxene rocks.

Under conditions of chemical weathering it is relatively stable, but may be concentrated during weathering and development of residual deposits. It alters to white, yellowish, or brown, opaque masses that are metamict and composed of titanium oxides such as rutile, and anatase (TiO_2) or sphene ($CaTiSiO_5$).

References: Palache, Charles, *et al.*, Dana's *System of Mineralogy*, vol. I, 1944; Katsura, Takashi, and Kushiro, Ikuo, 1961.

Maghemite (γ-Fe_2O_3)

Maghemite is one of the polymorphic forms of Fe_2O_3 that occurs in nature as a metastable form, α-Fe_2O_3, hematite, being the more stable form. γ-Fe_2O_3 can be changed to α-Fe_2O_3 by heating, but the temperature at which this change takes place varies between 200° and 700°C and depends on the history of the maghemite and the presence or absence of impurities. Relatively pure maghemite, which is brownish in colour and has a brown streak, is similar to magnetite in many properties. It has a hardness of 5, a specific gravity of 4.88, and magnetic properties like magnetite but with a Curie point of 675°C.

The crystal structure is similar in type to that of magnetite giving essentially the same X-ray pattern but the cell dimension is 8.32 Å. The difference in these isostructural minerals is due largely to the fact that some iron positions of both the tetrahedral and octahedral type in the maghemite lattice are vacant giving maghemite a defect type of spinel structure. As a result, the ferrous-ferric iron ratio in maghemite may vary considerably and the cell dimension is affected by this ratio (Basta, 1959).

Maghemite forms by oxidation of magnetite or by dehydration of lepidocrocite ($FeO(OH)$) but does not form from hematite (α-Fe_2O_3). It converts to hematite under suitable temperature and pressure conditions and is not present as an original mineral in medium to high temperature deposits (Mason, 1943). Maghemite is found in nature in zones of oxidation and weathering and in some lavas where it may have formed from low temperature hydrothermal solutions. The presence of impurities apparently has a stabilizing effect on maghemite, retarding the oxidation and lattice adjustment to stable hematite. It is not to be confused with martite, which is hematite that is pseudomorphous after magnetite. Hematite tends to develop along grain boundaries, parting planes, or crystallographic directions in magnetite, whereas maghemite develops in magnetite independently of crystallographic form.

Ferrous oxide is commonly referred to in discussing iron minerals, but wüstite (FeO) is not known to occur as a natural mineral. This artificial phase is miscible with magnetite and is unstable below 570°C.

References: Basta, E. Z., 1959; Berry, L. G., and Mason, Brian, 1959; Geith, M. A., 1952; Högg, G., 1935; Lepp, Henry, 1957; Mason, Brian, 1943; Newhouse, W. H., and Glass, J. P., 1936; Palache, Charles, *et al.*, Dana's *System of Mineralogy*, vol. I, 1944; Schneiderhöhn, H., and Ramdohr, P., 1931; Sosman, R. B., and Posnjak, E., 1925; Verwey, E. J. W., 1935.

Magnetite Fe_3O_4

Magnetite is the most common species in the magnetite series of the spinel mineral group. Minerals of the spinel group have the generalized formula of AB_2X_4 where A is one or more of the bivalent metals, B is one or more trivalent metals and X is oxygen. The other series in this group consists of the spinel series composed of Al combined with Mg, Fe, Zn, Mn, and Ni and the chromite series composed of Cr combined naturally with Mg and Fe. The magnetite series consists of magnesioferrite ($MgFe_2O_4$), magnetite ($FeFe_2O_4$), franklinite ($ZnFe_2O_4$), jacobsite ($MnFe_2O_4$), and trevorite ($NiFe_2O_4$).

Magnetite is black to brownish black, strongly magnetic, and a good electrical conductor; it has a metallic to semimetallic lustre and a black streak. Hardness ranges from 5.5 to 6.5; observed specific gravity is 5.175; calculated specific gravity for pure magnetite is 5.20. Magnetic attractability is 40.18 compared to 100 for pure iron, and some varieties (lodestone) show polarity.

Magnetite occurs as fine- or coarse-grained masses or in octahedral or less commonly dodecahedral crystals which may be twinned on (111) faces with spinel or lamellar twins. Parting is good on (111) planes and may take place on (001), (011), and (138) planes. Crystal class is 4/m 3 2/m, the oxygen atoms are in cubic close packing arrangement with an F or face-centred lattice, and the unit cell dimension is 8.391 Å. The unit cell containing 8 (AB_2X_4) atoms has 8, A type or ferrous iron atoms in fourfold coordination in a tetrahedral group of oxygen atoms and the 16 B type or ferric iron atoms are in sixfold coordination in an octahedral group of oxygen atoms.

According to the ideal formula, magnetite contains 72.4 per cent iron and 27.6 per cent oxygen but iron atoms may not be present in all of the appropriate lattice positions and the ratio of $Fe^{2+}:Fe^{3+}$ may change from 1:2 to approach the ideal composition of maghemite γ-Fe_2O_3. When the composition of magnetite is intermediate between Fe_3O_4 and γ-Fe_2O_3, the cell edge varies uniformly according to composition between 8.391 and 8.32 Å, but because of other substitutions the composition of magnetite cannot be determined on the basis of cell edge alone. There is a possible solid solution between FeO and Fe_3O_4 with not more than 3 per cent by weight of FeO in magnetite. Up to 7 per cent Fe_2O_3 may be present in solid solution in magnetite but a small amount of FeO present with Fe_2O_3 would lead to the formation of magnetite (Mason, 1943).

Magnetite may contain other cations which may substitute for either the bivalent or trivalent iron and the cell dimensions are altered when this takes place. Table IV gives an indication of the approximate amounts of various elements that may substitute for iron and the greatest change in cell dimension to be expected.

The cell dimension of magnetite is increased when Mn^{2+} or Ti^{4+} replaces iron but is decreased when other elements substitute. Constituents such as ZnO, NiO, CoO, or Cr_2O_3 are usually present in such small amounts that they do not affect the cell size to any appreciable extent. Al^{3+}, Cr^{3+}, and V^{3+} replace ferric iron and other elements replace ferrous iron. Titanium substitutes for both ferrous and ferric

Table IV

Elements Substituting for Iron in Magnetite[1]

Approximate amounts of various elements that substitute for iron in magnetite crystal structures and changes in cell dimension that may be caused by these substitutions

Element	Commonly not more than %	Reported Maximum %		Cell Dimension
Titanium	7.5	32.5	increases to	8.50 Å
Manganese	3	8.46	" "	8.51 Å
Magnesium	2	8	decreases to	8.37
Alumina	1.5	7	" "	8.12
Chromium	0.5	1.42	" "	8.34
Vanadium	2	4.84	" "	?
Nickel	—	1.76	" "	8.36
Zinc	—	0.2		—
Cobalt	trace	—		8.38
Calcium	1	—		—

[1] Data from Basta, E. Z., 1957; Palache, Charles, *et al.*, Dana's *System of Mineralogy*, vol. I, 1944.

iron but due to the intimate natural mixtures of magnetite and ilmenite ($FeTiO_3$) or ulvospinel (Fe_2TiO_4) it has not been possible to determine accurately the amount of titanium present in solid solution in magnetite and that present as inclusions of ilmenite. Titaniferous magnetite commonly has small blades and lenticular inclusions of ilmenite that have exsolved from the magnetite along (111) or other crystal planes during its cooling history. These blades and inclusions of ilmenite may vary in size from submicroscopic to millimetres in dimension or ilmenite may be associated as discrete grains. There appears to be a continuous solid solution between magnetite and ulvospinel which may in some circumstances give segregations of ulvospinel and magnetite. Schwellnus and Wellemse (1943) report an antipathetic relation between V_2O_5 and TiO_2 in the magnetic iron ores of the Bushveld Complex and the presence of V_2O_5 in the magnetite rather than in the ilmenite.

Magnetite is ubiquitous in both igneous and sedimentary rocks, and is a common accessory mineral in all types of igneous rocks. Very large masses occur as magmatic segregations associated with apatite and pyroxene. It is commonly titaniferous in gabbroic and anorthositic rocks and is associated with titanium minerals. Irregular masses and disseminations occur in contact metasomatic deposits where magnetite is associated with garnet, diopside, amphiboles, and skarn minerals in border zones between limestone or sedimentary-volcanic complexes and intrusive rocks. Magnetite is present with sulphide minerals in a great variety of vein and replacement deposits and occurs in many pegmatite dykes. It forms extensive disseminations in alkaline ring complexes often associated with columbium, tantalum, rare earths, and radioactive minerals, and is a product of fumarolic activity. Detrital magnetite is present in many kinds of sedimentary rocks and very appreciable quantities are concentrated in placer deposits both ancient and modern. The most widely distributed concentrations occur in Precambrian iron-formations where magnetite is associated with hematite, siderite, iron silicate minerals, and chert or quartz. Some of the magnetite in iron-formations is primary (Huber, 1958), formed from chemical precipitates, but much of it is of metamorphic origin.

Magnetite forms in a wide range of environmental conditions which are not fully defined or understood. It melts to a liquid at 1,591° ± 5° under low oxygen pressure; a decrease in melting temperature is noted as the oxygen pressure is increased. With rising temperature it changes from ferromagnetic to paramagnetic at 590°C, the Curie point. In most natural environments magnetite has evidently formed under elevated temperatures and was regarded by many as a typical 'high temperature' mineral. It can however be synthesized at ordinary temperatures by the oxidation of alkaline solutions of $Fe(OH)_2$ and by the precipitation by alkalies of mixed solutions of ferric and ferrous salts, by the oxidation of iron under elevated temperatures in air or steam, or by heating Fe_2O_3 and $FeCO_3$ in a reducing environment. Laboratory tests have shown that finely divided magnetite can be formed by two types of reagents, very dilute nitric acid or dilute aqueous solutions of alkali metal chlorides such as dilute sodium chloride solution or dilute sea water (Lindley and Brown, 1958). According to Huber (1958) from studies of Eh-pH conditions, magnetite and siderite are stable under intermediate to mildly reducing conditions but magnetite forms in preference to siderite under alkaline pH conditions. He further points out that magnetite is more important as a primary or diagenetic mineral in sedimentary rocks than has previously been assumed. Occurrences of authigenic magnetite, supergene magnetite, and magnetite in the form of oölites and pisolites in laterite deposits have been recorded in geological literature for some time. What is apparently most significant about conditions favourable for the formation of magnetite is maintenance of appropriate Eh and pH conditions and the relative concentration of other ions that may combine with the iron. Under higher temperature conditions in igneous rock metals, the ratio of ferrous oxides to ferric oxides and the partial pressure of oxygen appear to be dominant factors in determining if magnetite will form.

References: Brown, John S., 1943; Flaschen, S. S., and Osborn, E. F., 1957; Hawley, J. E., and Robinson, S. C., 1948; Holser, William T., and Schneer, Cecil J., 1961; Huber, N. King, 1958; Lindley, A. L. G., and Brown, F. H. S., 1958; Mason, Brian, 1943; Ramdohr, P., 1953; Schmidt, E. R., and Vermaas, F. H. S., 1955; Schwellnus, C. M., and Wellemse, J., 1943; Sosman, R. B., and Hostatter, J. C., 1918; Spiroff, Kiril, 1938.

Iron Carbonate Minerals

Siderite $FeCO_3$

Siderite is the principal carbonate mineral that is important as an ore, although some other carbonates are valuable as fluxing constituents in treating ores. It is yellowish, buff, ash-grey, or greyish brown to reddish brown, or occasionally greenish grey; has a white streak and is translucent to subtranslucent; generally fine grained and dull to earthy, but crystalline varieties may have a waxy lustre. It occurs in coarse to fine granular masses, in botryoidal, globular, or oölitic forms with argillaceous or siliceous impurities. Crystals have good rhombohedral cleavage and fractures are uneven to conchoidal; hardness is about 4; and specific gravity is 3.96 or less where other elements substitute for iron. The magnetic attractability is 1.82 compared to 100 for iron, being higher than hematite; magnetic susceptibility is less than limonite and similar to pyrite (Taggart, 1948).

Siderite and calcite are isostructural in the hexagonal system and form hexagonal-scalenohedrals with 3 2/m symmetry, or thin to thick rhombohedral or prismatic forms. Lamellar twinning on (0112) directions is common.

Pure siderite contains 48.2 per cent iron but a number of elements may substitute for iron in the lattice. There is a complete series from $FeCO_3$ to $MgCO_3$; Gruner (1946) noted that siderite in Mesabi Range taconites contains from 10 to 15 per cent $MgCO_3$. The compositions of a number of carbonates in the Gogebic Range, Michigan, are given by Huber (1959).

A complete series also exists between siderite and rhodochrosite ($MnCO_3$) but in reality siderite rarely has more than a few per cent MnO. Limited substitution for iron by calcium exists up to a maximum of Ca:Fe = 1:3.4 and in ferroan calcite the Fe:Ca ratio is up to 1:4.5. Cobalt may substitute for iron at least to the ratio of Co:Fe = 1:1.67. Sr, Ba, Pb may be present in small amounts. A linear relationship between refractive indices and composition of the various carbonates exists with both indices of siderite being above the corresponding indices of other species. The various carbonate minerals have characteristic temperatures of decomposition when heated. Kulp, Kent, and Kerr (1951) indicate that it is possible by differential thermal analyses to estimate quantitatively the concentrations of each kind of carbonate in a natural aggregate and to estimate the amount of substitution in each phase. Siderite decomposition starts around 510° ± 80°C and reaches a peak around 570° ± 60°C depending on the amount of impurity and grain size of the sample. Siderite decomposes to α-Fe_2O_3 and γ-Fe_2O_3 but the γ variety changes over to α-Fe_2O_3 (hematite) very readily in an air or oxygen environment.

Siderite, although found in a wide variety of circumstances, is most common in sedimentary rocks formed under neutral to low pH conditions and an Eh of 0-0.2, or a reducing environment. It is present in hydrothermal veins and in many gangue mineral assemblages formed at elevated temperatures. In sedimentary rocks it is most abundant especially as the principal carbonate in silicate-carbonate facies of iron-formation. Oölitic siderite beds are common or occurrences may be intimately associated with iron silicate minerals such as greenalite, minnesotaite, chamosite, or glauconite. Siderite cement is present in some sandstones and siderite lenses, nodules, and concretions occur in many sediments especially where carbon is present or evidence of a reducing environment is found. Siderite beds are mined in the Michipicoten area of Ontario and it is an important ore constituent in Wabana ores in Newfoundland and in the bedded ironstones or clay ironstones mined in Europe. Siderite alters readily by oxidation to form hematite or goethite and weathered surfaces usually turn deep chocolate-brown in contrast to ferruginous dolomite or ankerite which turn medium brown or buff on weathering.

The range of stability in metamorphic facies is not well defined but it is apparently unstable in conditions of the amphibolite zone.

Names used for siderite are chalybite, spathic iron, brown spar, and a number of others where impurities are found in abundance.

References: Bragg, W. L., 1950; Frederickson, A. F., 1948; Gruner, J. W., 1946; Huber, N. King, 1958, 1959; Kulp, Laurence J., *et al.*, 1951; Palache, Charles, *et al.*, Dana's *System of Mineralogy*, vol. II, 1951; Rowland, Richards A., and Jonas, Edward C., 1949; Taggart, A. F., 1948.

Dolomite Group

Dolomite $CaMg(CO_3)_2$
Ankerite $CaFe(CO_3)_2$
Kutnahorite $CaMn(CO_3)_2$

The dolomite group of carbonates is represented by these three end members, dolomite being the most common. There is a limited substitution between Mg and Ca and between Mg, Fe'' and Mn'', but the pure manganese end member is rare and the iron end member is unknown. The structure of the dolomite group is similar to that of calcite except that alternate Ca positions are occupied by Mg, Fe, or Mn and the crystal symmetry is lowered to 3. The iron-magnesium series is arbitrarily divided at Mg:Fe = 1:1 with Mg> Fe referred to as ferrodolomite and Fe> Mg as ankerite. The whole carbonate class of minerals has complex substitution relationships varying within certain limits that are difficult to define without elaborate diagrams, and some are still not fully defined. Fe'', Mn'', Co'', Pb, and Zn may substitute for Mg, and Fe'' and Mn'' can substitute to some extent for Ca. Ca may replace Mg up to about Ca:Mg = 1:5 in the Mg positions, and Mg replace Ca up to about Mg:Ca = 1:20 in the Ca positions. In the dolomite-ankerite series the ratio may be up to Mg:(Fe, Mn) = 1:2.6 and perhaps higher in Fe. In the Mn-rich carbonates the ratios may be up to Mn:(Mg, Fe) = 1:3.2. This substitution phenomenon is particularly significant in iron-formations where much of the manganese is present in the carbonates or in the relatively pure limestones where iron and manganese are present in intermixed dolomites. Analyses show up to 26 per cent FeO in dolomite and up to 23.4 per cent MnO, but either is usually under 15 per cent.

Many of the physical properties of dolomites are similar to those of siderite. Hardness varies from 3½ to 4 and specific gravity from 2.85 to 3.12. Dolomites form rhombohedral and less often tabular or prismatic crystals which may be striated or have curved faces, or the crystals may be arranged in subparallel clusters to form saddle-shaped masses. Twinning on (0001) and also on (1010), (1120) or lamellar twinning on (0221) is common. Colour varies from white, to creamy grey, buff, greenish, or brownish and with weathering buff to brownish depending on the amount of iron present. Ankerite variety is common in hydrothermal veins formed at higher temperatures and in replacement deposits, and dolomite is ubiquitous in its occurrence with calcite or other carbonates. Dolomite is most abundant in thick successions of carbonate sediments of Palaeozoic age and is especially prominent in Precambrian carbonate successions. Ankerite and dolomite are nearly always present to some extent in all facies of iron-formation. The origin of dolomite rocks, which is beyond the scope of this summary of properties and mode of occurrence, has been extensively discussed in the literature and the problems are reviewed by Goldsmith (1959).

Differential thermal curves show that dolomites dissociate over a broad temperature range with the peak temperature varying from 810°C for $CaMg (CO_3)_2$ to near 550°C for $CaFe (CO_3)_2$ rich members. Dolomite apparently does not precipitate directly on the sea bottom as calcite or aragonite does but Mg ions may substitute for ions during diagenesis to form dolomite. Dolomite forms readily under elevated pressures and temperatures in metamorphic processes.

References: Garrels, Robert M., 1960; Goldsmith, Julian R., 1959; Graf, D. L., and Lamar, J. E., 1955; Kulp, *et al.*, 1951; Palache, Charles, *et al.*, Dana's *System of Mineralogy*, vol. II, 1951; Pettijohn, F. J., 1949; Twenhofel, W. H., 1950; Winchell, A. N., 1947.

Iron Sulphide Minerals

Iron sulphide minerals are the principal iron-bearing constituents in some of the by-product ores from base metal mines in Canada, but their sulphur content is deleterious in most ores consisting mainly of iron oxides. The sulphides have marked effects in the weathering of iron-bearing rocks and play an important part in the formation of residual iron deposits. Pyrite, pyrrhotite, and marcasite are those most frequently encountered.

Pyrite FeS_2

Pyrite is pale brass-yellow, gives a greenish black or brownish black streak, has a metallic to splendent lustre, and a conchoidal fracture; is brittle with a hardness of 6 to 6½ and a specific gravity of 5.018. It has both positive and negative thermoelectricity. The magnetic attractability of pyrite is 0.23 compared to 100 for iron; the magnetic susceptibility is 0.00002 compared to 3.07 for magnetite. Pyrite crystallizes in the isometric crystal system with 2/m 3 symmetry, cell dimension of 5.417, and the iron atoms occupy face-centred cubic lattice positions with the sulphur atoms in pairs spaced along the trigonal axes and in sixfold coordination with the iron atoms. Pyrite forms cubic pyritohedral or octahedral crystals, and the crystal faces on cubes or pyritohedrons may be striated. One poor cleavage is developed on (100) and interpenetrating twins form on the (110) axis.

When pure, pyrite contains 46.6 per cent iron and 53.4 per cent sulphur. Nickel may substitute for iron up to the ratio of Ni:Fe = 1:1.84 and cobalt up to Co:Fe = 1:2.5. Trace amounts of V, Mo, Cr, W, Ti may be present. A complete series may exist between pyrite and bravoite (NiS_2) and the unit cell dimension increases with the Ni content.

Being one of the most common and widely recognized minerals, pyrite occurs in practically all kinds of rocks and mineral deposits. A reducing environment and neutral to higher pH conditions are favourable for its formation but at higher temperatures it is stable where pH values are lower. A great deal of experimental work has been done on the stability fields of pyrite in solutions and in rocks; these complex systems are discussed by Garrels (1960) and by Huber (1958).

Pyrite is abundant in most massive sulphide deposits bearing Ni, Cu, Pb and Zn and in most hydrothermal vein and replacement deposits, and is disseminated in most types of igneous rock either as an accessory mineral or as a magmatic segregation product. In many carbonaceous sedimentary rocks it is abundant in well-formed cubes; it may be finely disseminated in dense greenish black layers in the rock and be easily overlooked. Oölitic pyrite is present in some of the Clinton-type iron beds and pyrite is the main iron constituent in the sulphide facies of iron-formations where it is associated with carbonate minerals and fine-grained carbon-bearing clastic sediments.

Pyrite oxidizes readily to form goethite and the sulphate solutions generated play an important part in processes of supergene enrichment of other metals and in the leaching and solution of gangue mineral constituents.

References: Ault, Wayne V., 1959; Berry, L. G., and Mason, Brian, 1959; Garrels, Robert M., 1960; Huber, N. King, 1958; Kullerud, G., and Yoder, H. S., 1959; Palache, Charles, *et al.*, Dana's *System of Mineralogy*, vol. I, 1944.

Marcasite FeS_2

Marcasite although similar to pyrite is most easily distinguished by its crystal form. Marcasite is orthorhombic, and twinning repeated on (101) gives spear-shaped and cockscomb crystalline masses; hardness is 6 to 6½ and specific gravity is 4.89. Its electrical properties differ from those of pyrite. Marcasite differs little in composition from and may convert to pyrite. It is found in near-surface deposits where it generally forms under low temperature conditions from more acidic solutions than does pyrite; it decomposes more readily than pyrite under oxidizing conditions.

Pyrrhotite $Fe_{(1-x)}S$

Pyrrhotite has the highest iron content of the sulphide minerals and is a common associate of magnetite. It is bronze-yellow to copper-brown, has a metallic lustre, tarnishes readily on exposure, often becoming iridescent, and has a dark greyish black streak. Usually it is massive to granular. Magnetic attractability is 6.69 relative to 100 for iron and its magnetic susceptibility is 0.00575 compared to 3.07 for magnetite. Magnetism is lost on heating to 348°C. Fractures are uneven to subconchoidal, hardness is 3½ to 4½, and specific gravity varies from 4.58 to 4.65, or 4.79 for troilite the nodular variety.

Crystals are hexagonal with 6/m 2/m 2/m symmetry, pyramidal with horizontally striated faces or flattened and rosette-like. Twinning on (1012) may occur and parting on (0001). Metal atoms are in closely packed hexagonal layers in its crystal structure and alternate with closely packed hexagonal layers of sulphur. The iron atoms lie between 3 sulphur atoms below and 3 above; each sulphur atom lies between 6 iron atoms in a similar arrangement. Not all of the iron positions are filled as the formula $Fe_{(1-x)}S$ suggests and x may vary from 0 to 0.2. The approximate composition by weight per cent is sulphur 39.6 and iron 60.4. Several per cent of nickel may substitute for iron and small amounts of Co and Cu are reported, which may be present to a large extent in inclusions.

In massive sulphide deposits and in most deposits associated with igneous and metamorphic rocks that formed at elevated temperatures, pyrrhotite is a prominent mineral. Its association with pentlandite $(Fe, Ni)_9 S_8$ in basic igneous rocks around Sudbury is well known. A number of pyrrhotite-rich iron-formations occur in the Canadian Shield where mixtures of pyrrhotite and magnetite are interbanded with chert and tuff. It may form at low temperatures and is present in many unmetamorphosed sedimentary rocks. The stability range for FeS, pyrrhotite, in solutions is shown in the literature cited for pyrite and its formation apparently takes place in neutral to alkaline conditions in a strongly reducing environment where there is a high H_2S concentration or sulphur phase.

Because of its magnetic properties it may be troublesome in the concentration of magnetic iron ores, but in by-product ores its high iron content compared to that of pyrite may be advantageous.

Laboratory experiments show that H_2S reacts with magnetite and converts it to pyrrhotite at temperatures as low as 400°C and iron silicates may be converted in part to pyrrhotite at somewhat higher temperatures (Kilburn and Wilson, 1955).

References: Arnold, R. G., 1958; Arnold, R. G., and Ruchen, Laura, 1959; Hawley, J. E., and Haw, V. A., 1957; Kilburn, L. C., and Wilson, H. D. B., 1955; Kullerud, G., and Yoder, H. S., 1959.

Iron-Silicate Minerals

Many silicate minerals contain appreciable amounts of iron, but the following descriptions are restricted to those that are major constituents in iron-formations and that have iron as a principal cation. Iron-silicate minerals are quantitatively unimportant as ore minerals and are undesirable in most ores because of their silica content, but they are important as a primary source of iron that may become concentrated through oxidation and natural enrichment. Many other iron-bearing minerals in igneous and metamorphic rocks contribute to the formation of residual and laterite deposits but these are well-known rock-forming constituents and are not described here.

Chamosite $(Mg, Fe, Al)_6(Si, Al)_4O_{14}(OH)_8$

Chamosite is an iron-magnesium-aluminum silicate of the chlorite type but its structure differs somewhat from that of the common chlorite group. It is a greenish grey to yellowish or brownish grey platy mineral that resembles fine mica in some respects or to the naked eye, chlorite; is commonly extremely fine grained and can only be identified microscopically or by X-ray analyses; is weakly pleochroic with x yellowish or colourless and y and z pale green; generally appears nearly isotropic, 2v is negative and very small; is weakly birefringent with a mean index of about 1.64. The hardness is listed as 3, specific gravity is 3 to 3.5, crystals are short, platy with a good basal and a fair prismatic cleavage. The composition is highly variable with FeO content ranging from 33 to 42 per cent, Fe_2O_3 from 1 to 6.66 per cent, Al_2O_3 from 15 to 20 per cent, MgO from 1 to 4 per cent, SiO_2 from 23 to 27 per cent, and analyses show small amounts of MnO, CaO, and TiO_2.

Chamosite is a primary mineral in oölitic iron-formations associated with siderite, hematite, carbonates, chlorites, clay minerals and rarely with minnesotaite. In the Wabana, Newfoundland beds, in the bedded iron ores of Great Britain, and in some of the late Precambrian cherty iron-formations in Canada, it is an important ore mineral. Microscopic plates of chamosite are oriented tangentially or radially in concentric rings in oölites with alternate rings of siderite and hematite and also as patchy intergrowths in siderite or chert. Under low-rank metamorphic conditions it changes over to a common variety of chlorite or other iron silicate minerals.

References: Dana, Edward Salisbury, and Ford, William E., 1947; Goodwin, A. M., 1956; Hallimond, A. F., 1925, 1938; Winchell, A. N., 1947.

Glauconite $K(Fe, Mg, Al)_2(Si_4O_{10})(OH)_2$

Glauconite is one of the mica group of minerals, rich in iron, that forms in marine sedimentary rocks. It occurs in small green to black earthy granules and pellets that have a hardness of 2 and a density of 2.5 to 2.8. The pellets have a greenish streak and consist of clusters of concentric flakes. Glauconite is an authigenic mineral in many sediments; it may form from the alteration of biotite, or directly on the ocean floor in areas that are relatively free of clastic sediments. The mode of formation may be similar to that of greenalite or chamosite. Celadonite is a glauconite rich in ferric iron that occurs in vesicular cavities in basalt.

Greenalite $(OH)_{12}\ Fe_9''\ Fe_2'''\ Si_8O_{22}\cdot 2H_2O$

Greenalite is a primary mineral, known to occur only in relatively unmetamorphosed late Precambrian silicate-carbonate iron-formations. It was first described in material from the Mesabi Range but also occurs in the Gunflint Range and in the central part of the Labrador geosyncline. A member of the serpentine group, it is the ferrous analog of antigorite. Greenalite is usually cryptocrystalline, forming vague masses in granules or disseminations in chert and is commonly intimately associated with minnesotaite, stilpnomelane, and siderite. Patches of greenalite are pale olive-green to yellowish but may be brownish to very dark to opaque if altered or oxidized. Much of it appears isotropic in thin section and can be positively identified only by X-ray analyses. Granules of greenalite have a blotchy to velvety appearance, and it is found where the iron-formation shows little or no evidence of recrystallization. Small amounts of ferric iron in the lattice are sufficient to give it a pale green colour.

Crystals are monoclinic, platy or short lath shaped, and discernible with the electron microscope. The index of refraction is about 1.67, the specific gravity approximately 3, but the properties have not been accurately determined because of its extremely fine crystalline nature. An analysis reported by Gruner (1946) shows 22.95 per cent Fe_2O_3, 29.15 per cent FeO, 1.01 per cent Al_2O_3, 5.34 per cent MgO. 32.02 per cent SiO_2, and 9.52 per cent H_2O^+.

From both alloy and alloy-salt materials, greenalite has been synthesized at temperatures below 470°C. Flaschen and Osborn (1957) found in their investigation of the FeO-Fe_3O_4-SiO_2-H_2O system that greenalite and magnetite coexist at temperatures below 250°C as the decomposition products of fayalite. Fayalite is not stable in an aqueous phase below this temperature. Greenalite-minnesotaite-water forms a stable phase at temperatures below 470°C if the oxygen partial pressures are low, but magnetite appears in the system with these minerals where higher oxygen pressures are present. At 470°C greenalite decomposes to fayalite, minnesotaite, and water.

References: Flaschen, S. S., and Osborn, E. F., 1957; Gruner, J. W., 1936, 1946; Jolliffe, Fred, 1935; Winchell, A. N., and Winchell, H., 1951.

Minnesotaite $Fe_3(OH)_2Si_4O_{10}$

Minnesotaite is a ferrous iron silicate with a layered structure that is analogous to the magnesium silicate talc. In thin section it typically appears as colourless to pale greenish or yellowish flakes or needle-like crystals grouped in sheaf-like clusters or rosettes in random orientation in chert granules, as intergrowths in granules, or parallel with the schistosity. These platy crystals have a hardness of about 2.5, specific gravity about 3.0 to 3.2, and a waxy lustre. It is usually too fine grained to be positively identified in hand specimens but minnesotaite-rich bands in the Labrador geosyncline weather brown or a distinctive deep orange-brown. Fresh surfaces of the rock are olive-green to light khaki coloured. Under the microscope it usually shows weak pleochroism with $x =$ colourless to pale yellow, and $y = z =$ pale green. The indices of refraction are $x = 1.586$ and $y = ?$, $z = 1.618$, 2v is less than 5 degrees and negative, and crystals show positive elongation and near parallel extinction.

Crystals are monoclinic with a good basal cleavage. Gruner (1946) gave an analysis of: SiO_2 — 51.29 per cent, Al_2O_3 — 0.61 per cent, Fe_2O_3 — 2.0 per cent, FeO — 33.66 per cent, MgO — 6.26 per cent, MnO — 0.12 per cent, CaO — none, Na_2O — 0.08 per cent, K_2O — 0.03 per cent, TiO_2 — 0.04 per cent, H_2O^+ — 5.54 per cent, H_2O^- — 0.24 per cent. Magnesium substitutes for ferrous iron in natural specimens and the Fe^{++}/Mg ratio is usually less than 5.4 but is reported as high as 10.5, which indicates a limited solid solution at least with the magnesian end member, talc.

Minnesotaite has been synthesized at low oxygen partial pressure and high water pressure and has been found to be stable up to 480°C where it dissociates to give quartz + fayalite + water. Magnetite may be present if the oxygen partial pressure is sufficiently high. Minnesotaite is thought to be a primary mineral in some iron-formations but this has not been established beyond doubt. There is general agreement that it forms under diagenetic conditions from greenalite with which it is usually intimately associated and that it forms under very low rank metamorphic conditions from amorphous silicate material and greenalite by the reaction of siderite and carbonates with cherty silica, and from iron oxides and silica. It may also form as a retrograde metamorphic mineral from iron amphiboles, pyroxenes, or fayalite. It is abundant in the silicate carbonate facies of late Precambrian iron-formations that have not been highly metamorphosed. According to Tyler (1949) it plays an important role in the formation of soft iron ores in the Lake Superior ranges, because silica is leached from this iron-silicate more readily than from chert. Minnesotaite-rich beds in the Labrador geosyncline alter readily to give residual ore.

References: Flaschen S. S., and Osborn, E. F., 1957; Gruner, J. W., 1944, 1946; Huber, N. King, 1959; Tyler, S. A., 1949; Winchell, A. N., and Winchell, H., 1951.

Stilpnomelane $(OH)_4(K, Na, Ca)_{0-1}(Fe, Mg, Al)_{7-8}Si_8O_{23-24}\cdot 24H_2O$

Stilpnomelane, commonly associated with minnesotaite, is chemically related to chlorite and optically similar to biotite. Occurring in platy to fibrous or needle-like crystals in sheaves or clusters, it is greenish brown, olive-green to light amber, or khaki coloured. In thin sections it usually has darker colours, higher birefringence and stronger pleochroism than minnesotaite and resembles biotite except that the basal cleavage is not so well developed, it lacks the mottled effect at near extinction positions seen in biotite, has a strong golden-yellow tint, and the flakes are more brittle. The hardness is 1.5, and the specific gravity from 2.7 to 3.0, (—)2v is 0° or very small, x nearly $\perp$ 001, and because of its highly variable composition the indices of refraction vary greatly with x from 1.54 to 1.62 and z from 1.576 to 1.73.

The formula suggests considerable variation in composition, the proportions of ferrous to ferric iron may change from 34.4 per cent FeO and 0.35 per cent Fe_2O_3 to 3.86 per cent FeO and 31.67 per cent Fe_2O_3, Al_2O_3 may range from 4 to 7 per cent, MgO from 2 to 9 per cent, and up to 3.3 per cent K_2O and 2 per cent CaO may be present. MnO, Na_2O, TiO_2 are reported in quantities of 0.60 per cent or less.

It is most often recognized in veins or shear zones, but is a common constituent in iron-formations associated with magnetite, minnesotaite, grunerite, and cummingtonite and with epidote, albite, and chlorite or green amphiboles. Textural relationships show that it is a metamorphic mineral in practically all occurrences. Because some of its properties are like those of minnesotaite and some like those of biotite and because of its intimate relationship with other iron silicates, its presence is frequently overlooked. It is present in some chlorite schists and in many rocks other than iron-formations.

References: Gruner, J. W., 1944, 1946; Huber, N. King, 1959; Hutton, C. O., 1938; Winchell, A. N., and Winchell, H., 1951.

Iron Amphiboles

Practically all amphibole species contain some iron and several substitution series in the amphibole group permit very complex variations in composition. The general formula $(w, x, y)_{7-8} (z_4O_{11})_2(OH)_2$ with w signifying Ca, Na, K; x signifying Mg, Fe^{2+}, Mn; y for Fe^{3+}, Ti, Al; and z for Si and Al illustrates the possible range in composition. Two of these series, the anthophyllites and the cummingtonites, are of special significance in the study of iron deposits because they are major constituents in many metamorphosed iron-formations. Other species of amphiboles are bresent where the bulk composition of the rock is appropriate for their development.

Cummingtonite Series: Cummingtonite-Grunerite

This is essentially an iron-magnesium-rich series of the monoclinic amphibole group which ranges from $Fe_7Si_8O_{22}(OH)_2$ for the pure grunerite end member to a natural cummingtonite end member with the ratio of $(Fe_{25}:Mg_{75})$. Cummingtonites are usually light buff to straw coloured or greenish for the magnesian-rich members, and grunerite varieties are darker amber to khaki or brown to greenish brown. They form silky to vitreous aggregates and sheaves of acicular, fibrous, or needle-like crystals, elongated prismatic forms, rosette-like clusters or felted crystalline masses. The hardness is about 6, specific gravity 3.2 to 3.5, good (110) cleavage, and lamellar twinning may be discernible in hand specimen and is distinctive in thin sections. The indices of refraction vary because of the compositional difference, with x ranging from 1.627 to 1.686 and z from 1.646 to 1.729, strong birefringence is characteristic, $z \wedge c$ from 10° to 19.5°, 2v is very large, and pleochroism in yellow and brown to greenish brown is distinct. The systematic variation of these properties with composition are given by Winchell (1951). Al_2O_3 and CaO and TiO_2 are commonly present in small amounts and up to at least 14 per cent manganese may replace iron and magnesium and up to 10 per cent ZnO may be present.

Cummingtonite and grunerite occur in metamorphosed iron-formations from the upper greenschist facies through the amphibolite facies and are associated with magnetite, hypersthene, fayalite, actinolite, other amphiboles and pyroxenes, garnet, carbonates, quartz, talc, and minnesotaite. Grunerite may develop from minnesotaite, stilpnomelane, or chamosite and in some specimens its relationship with hypersthene suggests that it formed as an alteration of this mineral. It appears in some places to have formed from siderite and quartz. Cummingtonites occur in calcium-poor and iron-rich rocks and the bulk composition of the rock is apparently of greatest importance in determining where it forms. Where Ca is present actinolite forms as an associated amphibole. Flaschen and Osborn (1957) succeeded in synthesizing grunerite as an interim phase in the decomposition of minnesotaite to fayalite and water at 700°C and 17,500 p.s.i. pressure and Boyd (1959) reported synthesizing cummingtonite. The stability ranges are however not known.

References: Berry, L. G., and Mason, Brian, 1959; Boyd, Francis R., 1959; Flaschen, S. S., and Osborn, E. F., 1957; Gruner, J. W., 1946; Mueller, Robert F., 1960; Winchell, A. N., and Winchell, H., 1951.

Anthophyllite $(Mg, Fe)_7Si_8O_{22}(OH)_2$

Anthophyllite is similar to cummingtonite in appearance and general habit and can be positively identified only by microscopic or X-ray studies. It is usually rich in magnesium but may have as much as 60 per cent iron replacing magnesium; this series is miscible in all proportions with gedrite $(Mg, Fe)_5Al_2(OH)_2Si_6Al_2O_{22}$. Anthophyllite may be white, grey or brown, and silky to fibrous material is more common than in the cummingtonite series. The hardness is 5.5 to 6 and the specific gravity 2.9 to 3.4 is somewhat less than that of cummingtonite.

Anthophyllites differ from cummingtonite mainly in that they are orthorhombic and the birefringence is not as high. Anthophyllite has been synthesized as a metastable product from the breakdown of talc and it eventually changed over to enstatite and quartz. It and grunerite are not usually present in the same beds or layers in metamorphosed iron-formation and the bulk composition of the respective layers seems to have determined to a large extent which amphibole formed. Anthophyllites occur in a greater variety of metamorphic rocks than cummingtonites and form an important constituent of metamorphosed iron-formations. More than 2 per cent manganese is present in some of the anthophyllite in iron-formation.

References: Berry, L. G., and Mason, Brian, 1959; Bowen, J. L., and Tuttle, O. F., 1949; Boyd, F. R., 1959; Peacock, M. A., 1928; Winchell, A. N., and Winchell, H., 1951.

Other Amphiboles

Other iron amphiboles found occasionally in iron deposits are:

Common hornblendes
    Actinolite $Ca_2(Mg, Fe)_5(OH)_2Si_8O_{22}$
    Ferrotremolite $Ca_2Fe_5(OH)_2Si_8O_{22}$
  All mainly green to dark olive-green or black.
Oxy-hornblende subspecies
    Ferritremolite $Ca_2Fe''_3\ Fe'''_2O_2Si_8O_{22}$
    Ferriedenite $NaCa_2Fe''_3Fe'''_2O_2Si_7AlO_{22}$
  Usually dark brown.
Soda-hornblende
    Glaucophane-riebeckite $Na_2(Mg, Fe, Al)_5(OH)_2Si_8O_{22}$
    Riebeckite $Na_2Fe''_3Fe'''_2(OH)_2Si_8O_{22}$
    Arfvedsonite $Na_3(Fe'', Mg)_4Fe'''\ (OH)_2Si_8O_{22}$

These vary from light blue-green to deep purplish blue. The asbestisform of riebeckite is known as crocidolite.

Iron Pyroxenes

Only one of the pyroxene series is of particular interest in the iron-formations although many different species occur in highly metamorphosed beds. Hypersthene (Fe, Mg)SiO_2 is a common constituent in iron silicate-magnetite-carbonate facies of iron-formation found in the albite-epidote-amphibolite to amphibolite metamorphic zones. It is the iron-rich member of the enstatite series and the iron-magnesium ratio is arbitrarily set at 10 to 90 on the enstatite end and natural hypersthenes have a ratio up to 90 to 10. This orthorhombic pyroxene is amber-brown, greenish brown to dark bronze. It occurs in blocky grains but commonly is intimately mixed with coarse grunerite and can only be identified in this association by very careful optical or X-ray study.

Most hypersthene in iron-formations appears to have formed from grunerite under high-rank metamorphic conditions and occurs as irregular patches in large grunerite grains. Many X-ray patterns from picked grunerite grains have hypersthene lines, indicating very intimate mixtures of the two minerals.

Chapter IV

GENERAL CHARACTERISTICS AND EVALUATION OF IRON ORE DEPOSITS

Of the many different kinds of iron deposits concentrated in the earth's crust some are suitable for iron ore in their natural condition, others require concentration and processing before they are usable, and many others are of little economic interest.

Many complex factors must be considered in selecting an iron deposit for mining. Some are geological such as the location, size, distribution, configuration, mineralogy, grade of iron, presence of deleterious constituents, texture, structure, and character of the gangue minerals; others are cost and method of mining, of beneficiation, of transportation, etc. Many of these factors are subject to change from time to time, making frequent revision of estimated ore reserves necessary. Changes in ore specifications, technology, and market conditions may make deposits that are of no economic importance today valuable ores tomorrow and profitable operations today uneconomical later on.

Definition of Iron Ore

According to conventional practice, an iron ore may be defined as any natural aggregate of minerals from which metallic iron can be extracted at a profit. In many respects this definition is too restricted to cover all circumstances under which iron is won from the earth or to include much of the material regarded as potential ore and considered on properties to be a speculative asset. The size of a mineable deposit depends on a number of variables, but principally on the grade and quality of the ore, the transportation and development costs, and the availability of a market. The nature of acceptable ore may depend on the type of steel-making plants and technological methods already established in the area or it may change with new technological developments. Low-grade taconite deposits on this continent received little attention as potential ore so long as there was an abundant supply of direct-shipping ore and competition from foreign steel sources was limited. With the technological ability to produce high-grade ore concentrate from this material at a competitive price and the advantages of these concentrates in furnace performance over lower grade direct-shipping ores, the concentrates not only came into great demand but now give the principal competition to suppliers of direct-shipping ores. A mineral deposit may have the same grade and technological properties as an orebody that is currently being mined but may remain untouched for years because it is not required by established companies who have previously organized different channels of supply.

An important part of the definition stated hinges on whether metal can be extracted at a profit. Iron can, for example, be extracted from titaniferous iron ores by using special smelting processes where titanium oxide is also recovered, but large-scale production of iron from this material has not been economical to date. In a competitive industrial system the final criterion for an ore is whether its use will enable a company to operate at a profit. Even this criterion is not entirely satisfactory because the profit motive may receive different emphasis from time to time and ores may be mined because of certain financial subsidies, indirect benefits through tariffs, or through requirements caused by an imbalance in foreign trade.

A definition of iron ore must be kept flexible to satisfy many changing factors and conditions and must not be too restrictive if it is to be useful in the study of natural occurrences of iron. For this reason and in keeping with practices adopted in International works (*see* Percival, United Nations, 1955) a broader definition is given. Iron ore may be regarded as any merchantable or beneficiated mineral aggregate, that occurs in sufficient quantity to be a sustained supply of raw material to industry, and from which iron can be satisfactorily recovered under existing economic conditions or which may be usable in anticipated economic circumstances. In other words, both ore as used in the conventional sense and potential ore are considered as 'ore material' in this treatment of iron deposits. This broader definition seems to be more suitable to the economic geologist, as a large part of his work has to do with the evaluation of mineral occurrences and the procurement of suitable material for future industrial supply. Recognition of potential ore or material of potential value is not to be confused with the established terms of measured, indicated, and inferred ore which are used to designate the occurrence or possible occurrence of quantities of material of a specified grade and quality that meet existing economic requirements.

Concentrations of iron that satisfy this broader definition are called deposits, and concentrations of iron of mainly scientific interest are called occurrences. The term ore deposit lends itself to general descriptive use for a great variety of shapes and sizes of iron-rich zones, but the term orebody seems to be better suited to designate specific well-defined masses of material.

Classification of Iron Deposits

Deposits that are geologically similar and that furnish ores of similar physical, chemical, and mineralogical character should be grouped together in a systematic and orderly fashion. One of the difficulties in establishing such a classification is the diverse interests of the exploration geologist, the mining engineer, the metallurgist, and the mineral economist. These interests cannot all be met in one classification, but iron deposits can be grouped in such a manner that certain important features of practical interest to the exploration geologist and mining engineer will be emphasized. The classification outlined below is intended to be practical and objective and is based on empiric rather than on genetic factors. Genesis, with all its controversial aspects, is largely disregarded.

If the mode of origin of a deposit is known it can be classified according to many of the systems based on genesis. Unfortunately the origin or genesis of some orebodies either is not well understood or is controversial. A sound hypothesis for the genesis of a type of ore supported by documented fact is necessary for guiding exploration, nevertheless mode of genesis is not a satisfactory basis for classification when information is limited.

Ore deposits may be classified according to their mineral content, but while mineralogy is indeed important by itself it is not sufficient basis for classification. Texture, grain size, porosity, and structure of deposits are also important but are lost sight of if, for instance, soft earthy hematite ore and oölitic hematite-chamosite-siderite beds are both classified simply as hematite deposits.

Probably the most useful classification is one based on objective description of physical and chemical features and the relation of such features to the geological setting. The mineralogy, ore texture, particle size, nature and kinds of impurities, and grade can all be described objectively, and by relating these features to factors in the geological setting, such as kind of host rock or protore and kinds of associated rocks and their structure, a clear picture of many important features of the orebody can be presented. A fairly elementary description generally is all that is necessary to identify a newly discovered orebody with an established type of deposit. Once this identification is made, experience and information relating to the type can be applied intelligently in the development and exploitation of the new deposit.

The following classification (Table V), although similar in some respects to others already in use, is an attempt to make the main basis descriptive features and the relation and kinds of associated rocks, using well-known distinctive orebodies as types. This classification also illustrates the great variety of conditions in which iron deposits occur.

Table V

Classification of Iron Deposits

Group I HEMATITE-GOETHITE DEPOSITS IN IRON-FORMATION[1]. Direct-shipping ores and wash ores formed in iron-formation by leaching of silica and concentration of iron by natural processes.

Type A) i) Lake Superior type in Minnesota and Knob Lake type in Quebec and Labrador. Concentrations of hematite, goethite, and other iron minerals in cherty iron-formation. Cherty iron silicate, carbonate, and cherty iron oxide iron-formations are the common kinds of protore for this type. ii) Semitaconites, partly leached iron-formations associated with Type A) i) that provide "wash ore". These are considered as a special subtype in this group.

Group I – (cont'd)

Type B) The Old Helen mine deposit, Michipicoten district, Ontario. Concentrations of hematite and goethite of direct-shipping quality in siderite beds.

Group II IRON-FORMATIONS. Various kinds or types of iron-formation that may be utilized without concentration or beneficiation, or that are suitable for beneficiation.

Type A) Clinton type of iron-formation, Wabana deposits, Newfoundland. Hematite-siderite-chamosite beds with distinctive oölitic texture.

Type B) Minette or Lorraine oölitic siderite-chamosite iron-formations, mined in the Northampton district of England.

Type C) Helen mine and Sir James mine, Michipicoten district, Ontario. Siderite iron-formation consisting of stratigraphic units of fairly massive siderite.

Type D) Taconite iron-formation, Moose Mountain mine, Ontario, and Mesabi Range taconites, in Minnesota. Included here are all iron-formations that must be ground finer than 100 mesh to be concentrated. Common types include iron oxide, iron-silicate, iron-carbonate, and iron silicate-iron oxide types with a banded cherty matrix.

Type E) Metataconite iron-formation, Wabush Lake, Labrador, Newfoundland, and Lac Jeannine, Quebec, or the itabirite deposits in Minas Gerais, Brazil. Included here are metamorphosed iron-formations that have undergone extensive recrystallization and can be concentrated without grinding finer than 100 mesh.

Type F) Pyrite-rich iron-formations of potential ore value.

Type G) Other types of iron-formation of distinctive character found in local areas.

Group III RESIDUAL DEPOSITS, AND CHEMICALLY AND MECHANICALLY TRANSPORTED SURFACE DEPOSITS

Type A) Steep Rock Lake type, Steep Rock Range, Ontario. Buried residual accumulations of iron minerals on regoliths, with subsequent alteration of iron minerals and redistribution of iron. Mainly massive to crudely banded red and brown hard rubbly hematite and goethite ores of direct-shipping quality.

Type B) Residual deposits, Londonderry area, Nova Scotia; Bilbao, Spain.

Type C) Laterite deposits developed by surficial oxidation and supergene enrichment of rocks other than iron-formation, Conakry, French Guinea.

Type D) Bog iron deposits, St. Maurice, Quebec. Iron is transported chemically and precipitated near or on the surface.

Type E) Placer deposits and deposits formed by mechanical concentration of iron minerals. Iron sands of the North Shore of the St. Lawrence River, or the Charleson hematite-gravel deposit south of Steep Rock Lake.

Group IV DEPOSITS DIRECTLY ASSOCIATED WITH PLUTONIC ROCKS

Type A) The St. Charles deposit, Bourget township, Quebec; or Chaffey mine at Newboro Lake, Ontario. Layered, disseminated, interstitial, and injected magnetite with minor hematite and ilmenite in basic and ultrabasic rocks.

Type B) Allard Lake deposits, Quebec; or Taberg type of deposit in Sweden. Layered, disseminated, interstitial, and injected ilmenite with magnetite and hematite in anorthosite rocks.

Type C) Stephenville, Newfoundland. Deposits in acid intermediate rocks or in alkalic rocks. Massive and disseminated ores in these rock types. Further subdivision may be based on types of host rock.

Group V SKARN, CONTACT METASOMATIC, VEINS, AND STRUCTURALLY CONTROLLED EMPLACEMENTS OF IRON MINERALS

Type A) Texada Island, British Columbia; Marmora deposit, Ontario. Skarn, contact metasomatic deposits, or high temperature replacement deposits.

Type B) Forsyth deposit, near Hull, Quebec. Veins with minor replacement, deposited under structural control.

Type C) Disseminated magnetite, iron oxides, and iron minerals that are probably of replacement origin and deposited under structural control in shear zones and other structural features.

Type D) Noranda deposits, Quebec. Massive and disseminated iron sulphide bodies, may contain base metals and other ore minerals.

Group VI OTHER TYPES OF DEPOSITS

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[1] For definition of iron-formation *see* Chapter V.

Iron deposits may be arranged in five major groups. In Groups I and II fall the varieties of sedimentary iron-formation or their metamorphic derivatives. Group III comprises residual and laterite deposits developed on rocks other than iron-formation and mechanically and chemically transported surface deposits. Most of these are unconsolidated surface material but some may be buried and may then be regarded as special kinds of iron-formation. Deposits in Group IV are directly associated with plutonic rocks and consist mainly of layered, disseminated, interstitial, or injected masses of magnetite with minor hematite or ilmenite in various kinds of igneous rocks. Deposits in Group V may be indirectly related to igneous rocks or igneous activity or there may be no evident relationship to igneous processes at all. Skarn, contact metasomatic, replacement, and structurally controlled replacement deposits are all listed in this group because they have features common and are difficult to categorize specifically because of inadequate definition or genetic implications in some of the terms used to describe them.

Estimates have been made which indicate that about two thirds of the world iron ore comes from sedimentary deposits included here in the first two groups, about 10 per cent comes from deposits of direct or indirect igneous affiliation as included in Groups IV and V, and a quarter from deposits of Group III.

Characteristic features of the various types of iron ore deposits are described briefly in the following sections of this chapter. Because sedimentary iron deposits provide the most important sources of iron ore, a classification of iron-formations and detailed discussion of their origin is given in Chapter V. The genetic relationship between anorthosite intrusive rocks and iron-titanium deposits is considered in the following sections along with the descriptions of deposits in Group IV associated with plutonic rocks. The origin of other kinds of deposits, particularly the replacement and contact metasomatic deposits, is not always clearly defined and generalizations are apt to be ambiguous or misleading. Their origin is therefore considered in later volumes where specific examples from various parts of Canada are described.

Group I — Natural Enrichments of Iron-formations

Direct-shipping Hematite-Goethite Ore

Direct-shipping ores of the Lake Superior or Knob Lake type are composed of porous, friable to rubbly, red, blue-black, or brownish yellow masses of hematite and goethite within bands of cherty oxide or silicate-carbonate iron-formation. Material of direct-shipping quality usually contains from 51.5 to 65 per cent iron, up to 12 per cent silica and 0.045 per cent phosphorus, about 1.5 per cent alumina, variable amounts of manganese, and small amounts of calcium and magnesium. The structure and texture of the ore range from fairly hard, porous, rubbly material to relatively fine, earthy and unctuous masses, and the character of the ore usually varies with the type of iron-formation from which it was derived.

The oxidized iron minerals are considered to have formed from the iron-formation by the leaching of silica and carbonate by natural processes, leaving a residual concentrate of iron. Iron minerals in the very porous friable zones are oxidized but no appreciable amount of iron seems to have been disturbed or transported from its original position in the rock. Considerable redistribution, introduction, and enrichment of iron has evidently taken place in zones of denser ore.

Banded structures in relatively undisturbed deposits can be matched with the thin bedding in adjacent iron-formations. Very delicate structures and relict textures of iron-formation are commonly discernible in the oxidized ore. The ore deposits tend to be elongated to tabular masses which follow favourable bands in the protore iron-formation. This type commonly occurs where the iron-formation is fractured and highly deformed by folding and faulting, and many large deposits occur in gently dipping monoclines. In the central part of the Labrador geosyncline deposits of this kind contain up to 50 million tons of ore and on the Mesabi Range in the Lake Superior area individual deposits of this type may contain several 100 million tons of ore.

Semitaconite and Wash Ores

The boundary between direct-shipping ore and unaltered iron-formation may be abrupt in parts of some deposits but more often there is a gradation from high-grade material, to partly leached, to unoxidized iron-formation. These broad gradational zones of partly leached material are known as wash ores or semitaconites, which are utilized as ore that is being used in increasing quantity. The iron can generally be concentrated by simple washing and screening, gravity separation methods, flotation, or other means to give a high quality ore.

As a result of intergranular leaching only part of the silica is removed from this material by natural processes and the iron beds are converted to loose, porous, friable, and sandy material with iron minerals evenly distributed in small particles or as nodular fragments or lenses. Different facies of iron-formation respond in different ways to weathering and decomposition. The oxide facies usually form granular to sandy bands which concentrate readily by gravity methods. Silicate and silicate-carbonate facies give a greater variety of material, and finely divided iron particles in some of these have a tendency to form slimes and suspensions in washing processes and are consequently difficult to separate from the silica and silicate minerals. Deposits of the direct-shipping type of ore have been the principal source of iron on the North American Continent for the past 50 years but with the depletion of reserves of this type of ore in the Lake Superior area, increased use will be made of high-grade ore concentrates in the future.

Hematite-Goethite Ores in Siderite Beds

The third type of deposit in this group is exemplified by the Old Helen mine in the Michipicoten district, Ontario. The ore consists of a rubbly mixture of hematite and goethite developed in beds of siderite and siderite-pyrite iron-formation. The main difference between this type of deposit and the Lake Superior type is the kind of protore iron-formation in which the ore deposit formed.

Group II — Unenriched Iron-formation

Included in the second principal group are the iron-formations of various types that are merchantable as ore without elaborate concentration and others that constitute low grade ore but require beneficiation. Deposits in this group are particularly important because of the very large tonnage of ore exposed at the surface and mineable from open pits. Principles of stratigraphy and sedimentation can be fully utilized in the exploration and exploitation of these bedded ores, and as stratigraphic units they display a degree of uniformity in grade and distribution that is rarely found in other types of deposits.

Clinton Type Iron-formation

The Clinton type of iron ore represented by the extensive deposits at Wabana, Newfoundland, is a particular type of oölitic-textured iron-formation rich enough in iron and of suitable composition to be used with little or no concentration. These deposits contain about 51.5 per cent iron, 11 to 12 per cent silica, 3 to 5 per cent of both alumina and calcium, and 0.8 to 0.9 per cent phosphorus. This distinctive type of iron-formation is most extensively developed in lower Palaeozoic rocks and consists of compact massive beds of red hematite-siderite-chamosite oölites, one millimetre or less in size, formed around fossil fragments or clastic debris. The ore beds are commonly interbedded with limestones or shelf-type sediments and with dark green to black shale and argillite.

Siderite-Chamosite Ironstones

The siderite-chamosite ironstones in Mesozoic rocks in Europe have oölitic to granular textures, contain 30 to 40 per cent iron, and include various lithological types. Because of economic conditions in Europe this lower grade material may be used as ore with little or no special treatment. There are many subdivisions of this type of iron-formation, as indicated by mineralogical and petrological descriptions. Oölitic siderite beds in Cretaceous rocks in the Peace River area of Alberta belong to this general category.

Siderite and Siderite-Pyrite Formations

Siderite and siderite-pyrite iron-formations in the Michipicoten iron ranges of Ontario are considered to be a separate facies or type of iron-formation. They are associated with banded siderite chert and cherty siderite-magnetite iron-formations but the ore deposits consist of massive to crudely banded buff siderite with lesser amounts of pyrite distributed in the upper part of the bed. They are recognized as stratigraphic units occurring consistently throughout the region at one horizon in the stratigraphic succession. Siderite ore of this type is mined at the Helen and Sir James mines and is sintered to provide a high quality iron ore.

Taconite Ores

Iron-formations that require concentration of the iron and removal of undesirable constituents before they can be utilized as ore are grouped into two main subdivisions. This has been done arbitrarily on the basis of the grain size of the iron minerals and consequently on the fineness of grinding required to liberate the iron minerals and permit their concentration. Although the name taconite was originally used for iron-formation on the Mesabi Range and has recently been applied to other fine-grained iron-formations suitable for beneficiation, it is used here for any iron-formation that must be ground finer than 100 mesh screen size to liberate the bulk of the iron minerals. This type constitutes the first subdivision. Metamorphosed iron-formations that can be concentrated at coarser sizes are called metataconite and form the second subdivision. The actual composition of the iron-formation does not determine the subdivision into which it falls. The thin-banded magnetite-quartz rocks mined at Moose Mountain near Capreol, Ontario, and the cherty magnetite-silicate-carbonate formations on the east end of the Mesabi Range are both considered as varieties of taconite.

Metataconite Ores

Coarser grained metataconite iron-formations form very large deposits in the Wabush Lake-Mount Wright area and in the Ungava Bay area in the Labrador geosyncline. They consist of coarse-grained (+ 0.5 mm), banded magnetite-hematite-quartz or magnetite-grunerite-quartz iron-formations that have been recrystallized during regional metamorphism. Although most of the primary sedimentary feaures in the iron-formations and adjacent metasedimentary rocks have been destroyed or masked by subsequent recrystallization, relict forms of the original oölitic to granular texture can be seen in some beds. Some iron-formation deposits of this type are very highly deformed structurally, and determining their size, shape, and regional extent may involve many complex problems in stratigraphy and structural geology. These stratigraphic units generally contain between 30 and 40 per cent iron which may be recovered by relatively simple gravity or magnetic separation methods. Very large reserves of this type of ore have been proven in Labrador and Quebec and will no doubt be the major source of iron in Canada for some time.

Pyrite Iron-formations

Pyrite-rich formations have been developed for the recovery of iron on the Goudreau range in the Michipicoten area, Ontario. Many other beds have been discovered in the early Precambrian iron-formations that might be used if the sulphur can be either utilized or disposed of satisfactorily.

Group III — Residual and Transported Surface Deposits

A very interesting and diversified group of ores is included in the third major category of this classification. In general they are residual ores, derived from ferruginous rocks other than iron-formation, and from various kinds of chemically and mechanically transported surface deposits. In Canada, except for the Steep Rock type, deposits of this group are relatively small and of minor importance.

Canadian Iron Deposits – Geology

Steep Rock Type

The hematite-goethite ore from Steep Rock Range forms tabular to lenticular masses of fairly hard, porous, direct-shipping quality material. It is associated with dolomitic and soft altered argillaceous rocks on one side and with basic tuffs and volcanic rocks on the other. Opinions differ as to the origin of these orebodies but all lie along one stratigraphic horizon and recent work indicates that they are residual accumulations of iron oxides on an old erosion surface. Some alteration and redistribution of iron minerals has taken place subsequent to deposition and structural deformation. The iron content is 52 per cent or better, silica content is generally less than 7 per cent, alumina up to 2 per cent, phosphorus ranges from 0.01 to 0.03, and calcium and magnesium is low. Minor impurities including some pyrite are present locally. The tabular masses of ore dip steeply and the depth of ore below the underground workings is not known.

Residual Deposits

The best examples of this type of deposit in Canada are in the Londonderry area of Colchester county, Nova Scotia. Up to 1908, ore mined from these deposits consisted of goethite and hematite masses in roughly parallel lenses with very irregular borders. It varied in texture from loose, porous, yellowish brown limonite-goethite to dense hard lumps of brown goethite and mixtures of dense hematite and goethite. Small irregular siderite and carbonate masses were present in the iron oxide material and some zones contained specular hematite. Shallow lenses of iron-oxide are found in all ferruginous carbonate rocks containing 10 to 12 per cent iron. The residual iron accumulated in pockets as a result of deep weathering processes. Much of the ore contained over 50 per cent iron but material bearing as little as 35 per cent was also used. Some of the specular hematite is partly altered to goethite and was obviously present in these carbonate zones prior to weathering and supergene enrichment of the carbonate rock and is not related to the main ore-forming processes.

Weathering and supergene enrichment of iron from any kind of ferruginous rock may lead to the accumulation of a residual iron deposit but ferruginous carbonate is one of the most common types of protore.

Laterite Deposits

These deposits are common in tropical areas but are not known to occur in Canada. Some laterites are of suitable composition to be used as iron ore but most contain too much alumina or titania and may bear too much nickel, chromium, or other elements. Where laterites are worked for other ferride elements good quality iron concentrates may be produced as a second product. Laterites form a blanket type of deposit overlying various types of bedrock. These deposits are a product of deep weathering of rocks in tropical and subtropical climates and are the residual accumulation of oxidized and relatively insoluble rock constituents. They may be as much as 60 feet thick but are usually less than 20 feet. They consist of nodular red, yellow, or brownish hematite and goethite which may contain 20 per cent alumina in clay constituents. The thicker iron-rich blanket commonly forms over ultrabasic rocks that have a high primary iron content but nickel, chromium, and titanium are usually intimately mixed with the iron minerals. Laterite deposits are well known in Cuba, as well as in British Guiana, French Guinea, Indonesia, and the Philippines and other islands in the South Pacific.

Bog Iron Deposits

Most bog iron deposits are small, rather heterogeneous in grade and structure of material, and are therefore of minor interest. Included in this category is chemically precipitated iron, usually goethite but occasionally some siderite, that has accumulated around the discharge area of springs and seepage water either in low ground or on slopes. The iron minerals accumulated in lens-shaped surface coatings, occasionally as fans or bars in stream beds, and typically in the bottom levels of bogs and marshes. Organic debris and carbonaceous material are commonly replaced and fossilized by ferric oxide compounds or in some bogs by siderite. Manganese oxides may be abundant, and all gradations exist from bog iron to bog manganese deposits. Bog ores may be loose and friable, and when drained are very light and dusty. Other deposits consist of porous rubbly nodular or concretionary material. Bog iron deposits are commonly an indication of other types of iron occurrences in an area, such as sulphide deposits, pyritiferous shale, or carbonate beds.

The first iron ore smelted in Canada came from bog iron deposits near St. Maurice, Quebec, and this type of iron occurrence is found in all terrains. A number of rather inaccessible occurrences in British Columbia have been described.

Placer Iron Deposits

Deposits of this type vary from the black sand accumulations on beaches, riverbars, and lenses in glacial deposits, to deposits of hematite and goethite bearing gravels mined south of the Steep Rock iron range. The black sands formed as a result of mechanical sorting of stream and beach sediments and the iron mineral of importance is, generally, magnetite. The magnetite in most of the larger sand deposits in Canada, such as those along the north shore of the St. Lawrence River, bears titanium, and good quality iron ore concentrates are difficult to prepare.

The gravel deposits south of Steep Rock iron range bear fragments of hematite and goethite ranging from a few millimetres to inches in size. These are separated from the gravel by jigging and gravity separation methods. The gravel bears up to 15 per cent iron and a high grade iron concentrate as well as a by-product of sized gravel for concrete aggregate and road material is produced. The hematite-goethite fragments were obviously derived from the upper parts of the Steep Rock orebodies and removed by glacial erosion.

Group IV — Deposits Directly Associated with Plutonic Rocks

Iron deposits that occur within plutonic rocks are included in Group IV, and are subdivided mainly on the kind of igneous host rock and on the mineralogy of the deposit. A few of the more common types are mentioned here, but other deposits could be classified more explicitly where detailed information is available. Deposits in this group have various shapes or forms and the iron minerals may occur in layers, or as disseminated, interstitial, or injected masses; more than one form may be present in a single occurrence. Many of these deposits are formed syngenetically but the genesis of the discordant bodies is difficult to determine.

Magnetite in Basic or Ultrabasic Rocks

This type of deposit is characterized by the presence of accessory titanium, often in an amount too large to satisfy quality standards for good blast furnace operation. Many deposits of this type were mined in the past on a small scale but with developments in metallurgical practice and new steel-making processes most of them are now regarded as reserves for possible future production.

The best defined type of deposit in this group and the most common in Canada consists of magnetite with lesser amounts of ilmenite distributed in gabbroic and gabbro-anorthosite rock assemblages. Many of the gabbroic complexes are banded or crudely laminated and magnetite is present in various proportions in the bands as interstitial grains with pyroxene, olivine, and plagioclase feldspars or as aggregates in which magnetite and ilmenite form major constituents. Much of the ilmenite is intimately associated with the magnetite as fine blades or plates developed along crystal directions in the magnetite grains. The size of these blades may range from a fraction of a millimetre to submicroscopic dimensions but in some occurrences both magnetite and ilmenite form discrete grains with only a small amount of ilmenite present. Many gabbro intrusions are not banded or laminated and large parts, especially the central area, are massive. Nonetheless irregular zones of disseminated magnetite and ilmenite are common in these areas. Magnetite with ilmenite also occurs in these rocks as irregular-shaped discordant bodies that appear to have been injected into the gabbro or to form vein-like masses within it. Discordant stringers or shoots of massive magnetite may cut across banded or foliated gabbro carrying abundant disseminated magnetite. In some gabbro masses the magnetite forms definite patterns but in others its distribution is erratic and unpredictable. Where the enclosing igneous rocks are diorite or intermediate in composition, titanium is usually less abundant and the magnetite is relatively free of other impurities. Considerable apatite is found in some deposits of this type.

In some deposits, especially those within ultrabasic rocks, magnetite occurs in the manner described above but is associated with chromite instead of titanium.

The Chaffey mine and associated deposits near Newboro Lake, Ontario, and the St. Charles deposit in Bourget township, Quebec, are examples of magnetite-ilmenite deposits in gabbro. Deposits of this type commonly occur in the basic rocks of the Grenville geological province, and are well recognized throughout the world in all ages of intrusive rocks.

Many occurrences of magnetite in amphibolite rocks show the same distribution of the iron and titanium and probably had the same origin as those just described. Some are obviously metamorphosed gabbros or basic rocks that originally contained concentrations of magnetite.

The shape of these deposits and the habit of distribution of the iron minerals vary greatly in detail but the mineral content as well as the geochemistry is directly related to that of the igneous host rocks. Classification, therefore, on the basis of mineralogy and kind of host rock is not only convenient but especially significant.

Ilmenite-Magnetite-Hematite Deposits in Anorthositic Rocks

Iron deposits in anorthositic rocks are similar in mode of occurrence to deposits in gabbroic rocks just described and may consist of ilmenite with minor amounts of hematite or magnetite or of magnetite with small amounts of ilmenite. Injected masses of iron and titanium oxides appear to be much more abundant in anorthosite than in gabbro but layered, disseminated, and interstitial iron and titanium oxides still make up much of the deposits. Where ilmenite is the major constituent, hematite is usually intimately associated, on a microscopic scale, with ilmenite grains as exsolved bars and plates along crystallographic planes much like the magnetite-ilmenite intergrowths. Deposits in anorthosite rocks commonly contain considerable vanadium and other ferride elements.

Deposits of this type are found in nearly all anorthosite rocks in Canada, the Allard Lake deposits in Quebec, which consist mainly of ilmenite, being outstanding examples. These are mined and processed for the recovery of titanium dioxide, although metallic iron is produced as a second product in the processing of the ore.

The Genetic Relationship between Anorthosites and Iron-titanium Deposits

The origin of anorthosites. Two main types of anorthosite are recognized: the first forms layers in stratified basic intrusions, such as the Bushveld or Stillwater complexes; and the second forms domical masses of andesine-labradorite rock with associated hypersthene gabbro, norite, diorite, and syenite intrusions. Bodies of the second type, which are found only in Precambrian terrains where they are associated with gneisses, granite, and charnockitic rocks, are common in the Grenville Province.

Concepts regarding the origin of these rocks differ considerably. Arguments have been raised against the possible existence near the earth's surface of magmas of the composition of anorthosite because of the extremely high temperature required to maintain melts of this composition and also the fact that lavas of anorthosite composition are not known. An alternative hypothesis suggests that early formed plagioclase crystals settled from a melt of gabbroic to dioritic composition and that the residual magma was squeezed out to crystallize elsewhere, which would result in the formation of separate intrusions of complementary composition. A re-evaluation of this topic, with the benefit of more physical-chemical data, has been made by Buddington (1960), who found several lines of evidence compatible with the theory that the large domical anorthositic bodies and related sheets are differentiated facies developed under the control of gravity from a magma equivalent in composition to gabbroic anorthosite. He noted that evidence from phase petrology, plagioclase twinning, exsolved micro-intergrowths of clinopyroxene in orthopyroxene, and the composition of titaniferous magnetite are all consistent with inferred high temperatures essential for an origin from anorthositic phases of a magma. Work by Yoder (1958) is cited which shows that the role of water in lowering the melting interval of plagioclase and shifting the cotectic ratio of plagioclase and pyroxene towards the plagioclase component gives additional support for accepting a magma of gabbroic anorthosite composition from which relatively pure anorthite could be differentiated. From this work it is seen that where the water pressure is 5,000 bars, which can conceivably develop in a magma at a depth of 10 kilometres, the melting range of plagioclase, An_{60}, is about 820° to 1,100°C and the water content would be about 10 per cent, the composition of the diopside-anorthite eutectic is 73 per cent anorthite and its melting temperature is 1,095°C (Turner and Verhoogen, 1960). The possible magma temperatures indicated by this work are certainly reasonable and compatible with some of the metamorphic effects observed in the rocks adjacent to anorthosite masses.

From this information one can visualize the early separation of plagioclase crystals of the composition found in the anorthosite masses and the gradual enrichment of the remaining magma in ferromagnesian, alkali, and volatile constituents. With settling out by gravity of the early formed plagioclase, the remaining upper part of the magma would become more gabbroic or dioritic, and eventually rocks of this type would form in the upper and peripheral parts of anorthosite complexes. Any structural disturbance causing rupture of the enclosing rocks or lithified shell around the magma would permit the intrusion of dykes and small masses of gabbro and diorite. Because of the nature of the diopside-anorthite eutectic for the magma composition postulated, a higher ferromagnesian mineral content would be expected in the cooler border zones. The differentiation process would also give rise to high iron and titanium end fractions which would eventually crystallize as disseminated oxide grains in zones of anorthosite and gabbro. Some of the iron would be injected prior to crystallization as an oxide melt and some transported in the volatiles and emanations concentrated in this cooling process.

This concept of the origin of anorthosite offers a satisfactory explanation for the size and shape of the anorthosite masses and the distribution of gabbro and diorite intrusions around their margins. Injection phenomena and metamorphic effects around the borders of the anorthosite masses are also in accord with this mode of origin. The gabbro, norite, diorite, and syenite bodies occurring in and around these masses are evidently differentiates of the postulated gabbroic magma. Their composition may be greatly modified in places, however, by assimilation of other rock types.

The presence of volatiles in the parent magma can be inferred from some of the metamorphic effects in rocks adjacent to the gabbro-anorthosite masses and from possible deuteric effects imposed on mafic minerals in the intrusive mass. The abundance of amphibole and biotite in the metamorphic zones and within the intrusion, the development of chlorite and the uralitized pyroxenes in the intrusion, and the presence of considerable apatite and in places scapolite or other minerals bearing volatile elements, are all evidence of the presence of appreciable water and volatile constituents in the magma. Weak HCl solutions escaping from a cooling magma along structural channels and openings would be capable of carrying iron and other constituents in solution in considerable quantity (Holser and Schneer, 1961). The iron and associated constituents would eventually be precipitated in skarn and replacement deposits of magnetite or in veins of magnetite and apatite, similar to those found near gabbroic anorthosite masses.

The iron-titanium deposits and gabbro-anorthosite rocks are obviously phases of the same igneous process, but opinions differ as to the way in which the oxides and silicates are separated and the oxides are emplaced. As crystallization proceeded the residual magma evidently became richer in ferromagnesian constituents, iron, titanium, and ferride oxides. Whether or not there is immiscibility between oxide and silicate fractions in the melt is not clear but the ultimate effects led to a segregation of ferromagnesian silicates and iron-titanium oxides. Carstens (1957) considered that iron and titanium were concentrated at two main stages during the cooling of the magma. The first appreciable concentration, according to him, would be during the middle part of the cooling period when an iron-rich magma becomes sufficiently oxidized to produce iron-titanium minerals. This oxide material is apparently removed by such processes as filter pressing, and injected into fractures in the enclosing gabbro or anorthosite rock, or forced into tabular layers and irregular masses in partly lithified gabbro-anorthosite. A second concentration of ferride constituents is believed by Carstens to take place after slow progressive crystallization of magma at depth. This produces residual solutions rich in TiO_2, P_2O_5, and Fe, and the apatite-bearing, titaniferous iron deposits and magnetite-apatite veins may have been deposited by solutions of this kind. Carstens also proposed a hypothesis whereby iron and titanium may be expelled from the lattice of pyroxene minerals during late magmatic stages.

Iron-titanium minerals in anorthosite and gabbro. The relative amounts of iron and titanium oxides vary in deposits in different gabbro-anorthosite masses, and the mineralogy of these oxides is complex. Two main aspects of these mineral complexes have been investigated at length in recent years, both of which have economic and scientific significance. Attempt has been made to find out (1) what physical-chemical factors determine the mineral assemblage and character of mineral intergrowths formed by these oxides; (2) why some iron and titanium oxide masses have reverse remnant magnetism and give negative magnetic anomalies.

A great variety of mineral intergrowths are found that consist of ilmenite with blades of hematite, and magnetite; ilmenite with minor hematite; ilmenite as blades and platy intergrowths along crystal directions in magnetite; mixtures of discrete ilmenite grains and magnetite with blades of 'exsolved' ilmenite; ilmenite blades in magnetite that bears ulvospinel in very fine micro-intergrowths; and a number of other complex intergrowths composed of magnetite, ilmenite, hematite, ulvospinel, rutile, and anatase. The mineral assemblages are considered to have formed from oxide melts and their textures to have developed during cooling by exsolution of separate mineral phases from solid solutions. Titanium-iron spinel minerals found in some lavas are high in titanium and have compositions comparable to some of the titanium-iron oxide intergrowths in anorthosites, but these spinels are homogeneous and are solid solutions of oxides in which little or no exsolution has taken place. When some of the mineral intergrowths in anorthosites are heated to high temperatures the different mineral phases disappear and a single solid solution phase is formed, suggesting that many of the fine-grained mineral intergrowths were formed by exsolution of solid solution phases.

The mineralogy and texture of the iron-titanium oxides were apparently determined by the bulk composition of iron-titanium oxide melts, the temperatures that prevailed in the melts, the rate at which the oxides crystallized and cooled, the oxidation-reduction potential in the melts, and the abundance of volatile constituents. Much has been done to evaluate the effects of these different factors, and the very extensive results cannot be fully reviewed here. Buddington, *et al.* (1955) found that the amount of TiO_2 in titaniferous magnetite varied with the nature of the host rock and that it was lowest in rocks of granitic composition and highest in gabbros and olivine basalts. Because the temperature ranges in which these rocks crystallize also vary, with basalts having the highest temperatures, they suggested that the per cent TiO_2 in the magnetite is a function of temperature. They further suggested that, provided ilmenite is present as a separate mineral, the composition of the titaniferous magnetite may be used for geological thermometry. Wright (1961) concluded, from studies of solid solution relationships in titaniferous iron oxide ores in the Skaergaard Intrusion and in anorthosites at Lac la Blache, Saguenay county, Quebec, that granular aggregates of titaniferous magnetite and ilmenite have originated by exsolution rather than by simultaneous crystallization. He showed that Fe:Ti ratios are constant for granular aggregates of ilmenite and titaniferous magnetite that formed at any one stage in the cooling history of the containing rock. He suggested that granular aggregates of ilmenite and titaniferous magnetite in rocks lacking separately crystallized ilmenite would always have Fe:Ti ratios greater than 3. If the ratio is less than 3 there would be sufficient titanium present for ilmenite to crystallize as a separate primary phase. It is important to note that these statements refer to granular aggregates and that ilmenite may be present as exsolved blades in magnetite when the ratio is much higher than 3.

The paragenesis of titaniferous magnetite, ilmenite, and ulvospinel (Fe_2TiO_4) has been investigated in the Skaergaard Intrusion by Vincent (1960), and natural and synthetic titano-magnetites of Fe_3O_4-Fe_2TiO_4-$FeTiO_3$ system by Basta (1960). It appears that where the composition of the oxide melt is appropriate, a titanium-rich spinel mineral may form first at high temperature and ilmenite may exsolve along crystal directions from this spinel. If the temperature decreases slowly ulvospinel may exsolve in a fine grid pattern from the spinel mineral, magnetite. Later oxidation of this ulvospinel may lead to exsolution of a second age of ilmenite in the form of rods or blades in the ulvospinel-magnetite mineral grains. The studies of both Vincent and Basta show that the compositions of these mineral complexes may change or adjust if the oxidation state or temperature of their environment alters.

Because the nature of the iron-titanium mineral intergrowths is controlled by a number of variables, the cooling history or genesis of each occurrence must be inferred from its individual mineralogical and textural features. A satisfactory model has not been devised to represent the wide range of physical-chemical conditions in which these minerals crystallize or to explain their diverse features.

Magnetite and Hematite Deposits in Intermediate to Acidic Rocks

A syngenetic relationship of iron minerals and host rock is not as evident or widely recognized for deposits in intermediate to acidic rocks as it is for many deposits in gabbro and other basic rocks. Many iron deposits of this type are considered to be the product of magmatic differentiation at depth and may or may not have close genetic affinity with the enclosing igneous host rocks. With conclusive evidence and criteria lacking for determining the genetic affinity of many of these iron deposits in intermediate rocks, deposits are necessarily grouped according to their specific kind of host rock.

In Canada, deposits of this general type are not so common as those in more basic rocks, and no large occurrences are known. In the Indian Head area near Stephenville, Newfoundland, small lenses, disseminated zones, and patchy masses of hematite and magnetite occur in syenite and soda granite-gneiss. The zones rich in massive to disseminated magnetite and grey hematite are generally tabular to lenticular and are concordant with the foliated rocks in which they occur, although some discordant veins and replacement bodies are also present. The iron deposits in intermediate rocks in this area are relatively free of titanium, but other deposits in the same area associated with norite and basic rocks contain considerable titanium and some vanadium.

The best known iron deposit in intermediate plutonic rocks is the Kiruna deposit in Sweden. This large, steeply dipping tabular body of magnetite lies between a syenite porphyry foot-wall and a quartz porphyry hanging-wall. The orebody consists of massive magnetite which appears to have been injected into the surrounding rock. Over a billion and a half tons of magnetite ore has been proven by drilling to depths of 1,000 metres. The phosphorus content, present as apatite, in this ore varies and may be as high as 1.8 per cent but sulphur and titanium are low or insignificant.

Magnetite Deposits in Alkalic Rocks

A number of interesting magnetite occurrences in alkalic rock complexes in northern Ontario form a special type of potential iron ore. The alkaline rock complexes, which form circular structures half a mile to 5 miles in diameter, are composed of various rock types ranging from syenite, nepheline syenite, ijolite, augite syenites, lamprophyre dyke rocks, biotite-pyroxene-rich rocks, and various hybrid rocks as well as carbonate-rich zones. The various rock types when mapped appear as ring-like or crescent-shaped zones arranged concentrically around an intrusive core. Magnetite is present in massive bodies and as disseminations in certain rock zones in these complexes or it may be present in the matrix of breccia rocks and in fractures and fault zones. Apatite is closely associated with magnetite in many occurrences and the proportions of magnetite to apatite may be about 3 to 1. These deposits are particularly interesting because of the great variety of elements present. They have been examined as a potential source of niobium (columbium) and tantalum but contain a number of rare earth elements and appreciable titanium, zirconium, manganese, uranium, and thorium. Their potential as a source of iron ore depends mainly on the incidental recovery of a magnetite concentrate if the ore is milled for other constituents.

One group occurs along a line passing from Kapuskasing to Chapleau, and deposits near Nemegos and Nemegosenda Lake have been studied in some detail. Other deposits are present in Clay and Howells townships north of Kapuskasing, but their internal structure is not well known.

Group V — Skarn, Contact Metasomatic Deposits, Veins, and Structurally Controlled Emplacements of Iron Minerals

Many commercially important ore deposits and innumerable iron occurrences are included in this group, although the amount of potential indicated ore in these deposits appears small when compared with that in iron-formations or in basic rocks. Most of the Group V deposits are considered to be related genetically to igneous activity, and this relationship is fairly obvious in the contact metasomatic deposits. The source of the iron in most veins and replacement deposits and the way in which it was transported and deposited are not well understood. The formation of these deposits has been attributed to igneous and metamorphic processes or, in some veins, to the action of groundwater. Deposits have been considered separately in many of the classifications based on genetic factors, and are especially prominent in classifications that emphasize the formation of ore deposits by hydrothermal solutions.

Contact Metasomatic, Skarn, and Replacement Deposits

Deposits with many common features and characteristics may be referred to in the literature by different names, as the above title suggests. Those on Texada Island, for example, are typical of a large number of skarn or contact metasomatic deposits in the coastal area of British Columbia, and the Marmora deposit is typical of other similar occurrences in Ontario and southern Quebec. In general, this type develops in contact metamorphic aureoles developed in the metasedimentary and volcanic rocks adjacent to syenites, porphyries, and intermediate to acidic intrusions. These border rocks are silicified and altered to a complex assemblage of silicate minerals or skarn consisting of epidote, garnet, pyroxenes, and amphiboles. Magnetite, which is the predominant ore mineral, is distributed in these skarn rocks and to a lesser extent in the intrusive rocks in an erratic manner, generally as massive pods, shoots, veins, and stringers or disseminated in poorly defined patches. Beds of crystalline limestone are nearly always present in the succession of metasedimentary rocks and are usually the most favourable host rock for ore emplacement and skarn development. In some deposits bands of tuff or greenstone may be selectively mineralized and altered to skarn, but limestone is present in most deposits and evidently plays an important chemical role in the deposition of magnetite and skarn minerals. The continuity of many of the mineralized zones is disrupted by both pre- and post-ore dykes which vary in composition from alaskite, lamprophyre, and syenite to diabase. The patchy, irregular or erratic distribution of the magnetite is typical of these deposits and accurate estimates of the grade and quantity of ore are generally difficult to make even when considerable detailed information is available. Although many high-grade pods of massive magnetite may be present in these contact zones, rarely is any one large enough to be mined individually. Generally the whole mineralized zone is mined, and after elimination of as much waste rock as possible in the pit the crude ore is crushed and ground, and the magnetite concentrated magnetically. When this general procedure is followed the crude ore from most deposits runs between 40 and 45 per cent iron and concentrates about 65 per cent. Fairly coarse grained apatite is associated with the magnetite in many deposits and sulphides, including pyrite, pyrrhotite, and chalcopyrite, may be present in sufficient quantity to require fairly elaborate processing for their elimination.

Veins, Open-Space Fillings, Replacement Deposits

The shape, size, and distribution of these deposits are controlled mainly by structural factors and only a few are large enough to be of economic interest. Vein deposits in a local area may have many characteristics in common but generally the mineralogy and structural environment of both vein and replacement deposits vary widely. The Atikokan iron mine in Hutchison township, Ontario, falls into this group and was located in a magnetite vein that followed a fault zone. The Forsyth deposit near Hull, Quebec, is another example of a magnetite vein deposit.

Disseminated Deposits

Iron minerals, probably of replacement origin, are commonly disseminated along shear zones or fracture and joint systems. Magnetite is also present in appreciable quantity in the altered wall-rocks adjacent to some massive sulphide bodies. Some of these mineralized zones may be suitable for beneficiation.

Sulphide Deposits

Iron in the form of pyrite, pyrrhotite, and magnetite is a major constituent in many sulphide deposits mined primarily for the nickel, copper, base metal, and gold. Considerable iron is being recovered in this manner from the Noranda deposits, Quebec, the nickel deposits in the Sudbury area, Ontario, and from pyrrhotite tailings at Kimberley, British Columbia.

Evaluation of Iron Deposits

Iron ore is a bulk product of relatively low value compared with many other mineral commodities, and much of its cost on delivery at a steel plant consists of transportation, handling, or processing charges. It is abundant on most continents, but many of the large deposits of good grade ore occur far from the main areas of ore and steel consumption. Several types of ore of different quality and specification are utilized, depending on the smelting equipment being used and the kind of iron or steel needed for the market. Technological developments in smelting methods introduce frequent changes in the specifications for chemical grade and structural quality of ores. These specifications are met either by supplying ore from different types of deposits or by processing ore from established mines.

Many technical and commercial factors must be considered in the appraisal of iron deposits. The size, grade, and location of deposits as well as transportation facilities, availability of power, market requirements, and many other factors must be favourably combined if successful ore production is to be achieved. Any one or a number of factors may be critical in evaluating an iron deposit in the highly competitive iron ore industry.

Commercial Classification of Iron Ores

A general classification is based on the nature of the commercial product, which, in turn, depends on the natural grade and quality of the ore and the treatment it receives. Further classification is based on physical structure. Finally a fairly elaborate grade classification based on chemical analyses of the ore is used for calculating the selling price of ores.

Direct-Shipping Ores

Ores of this type are acceptable for use in blast furnaces or other types of smelters in the condition in which they come from the mine. About the only treatment they receive is crushing and in some cases screening for size. At one time practically all the iron ore used in North America was of the direct-shipping type, but currently only about one third of the ore used is of this type and technological developments indicate that proportionately less will be used in the future.

Treated Ores

Concentrates

Much of the available ore requires some beneficiation by washing, jigging, drying, hi-density or gravity separation, flotation, magnetic, or other processes to increase the iron content and remove siliceous gangue or undesirable constituents.

Agglomerates

Further treatment of ore concentrates and of some natural high grade but fine-grained powdery ore is necessary to improve their structure and smelting quality. This may be by sintering, pelletizing, nodulizing, or briquetting. The structural qualities produced bring about so decided an increase in efficiency in blast furnace performance, particularly for pelletized ore, that processing of nearly all the available natural ore is now warranted.

By-product Ores

Iron is recovered from the treatment of some ores that are mined primarily for the recovery of other metals. Pig iron is a by-product in the smelting of titania-rich ores for titanium oxide or titanium metal. Iron sulphide concentrates and magnetite recovered from treatment of copper, lead, or zinc ores may be roasted, pelletized, or reduced to metal in electric furnaces.

Classification by Physical Characteristics

Further classification is based on the particle size and general physical features such as density, colour, friability, porosity, and grain size. The deposit name or the type from which the ore comes may indicate the nature of the material.

Ore from the Lake Superior district for example has been considered in two classes 'Old Range' and 'Mesabi'. Old Range was applied to ore with hard coarse structure mined on the Negaunee or Marquette ranges and was the first type to be mined; Mesabi type was soft, fine-grained, earthy, mined later on the Mesabi Range. Subsequently these names were applied elsewhere to similar ores.

Lump Ores

These are hard dense massive ores that break into resistant lumps and withstand considerable handling without appreciable fragmentation. Lump ores of suitable chemical composition for use in the open-hearth furnaces in refining steel are of premium quality and price. Less desirable lump ores are blended in blast furnace burdens to help to maintain uniform structure and permeability for the reducing gases. They are carefully sized to suit particular requirements.

Soft or Fine Ores

Many ores are fine grained, soft, friable earthy masses that break down easily when handled. A high proportion of the ore passes through quarter-inch diameter screen openings and only a little can be used directly in the furnace burden without choking gas circulation or being blown out with exhaust gases. Fine dusty fractions are usually sintered or agglomerated in specially prepared furnace burdens. Price penalties are commonly imposed on ores with a high proportion of fines.

Classification by Chemical Composition

Iron ore analyses usually list iron content on a natural basis and phosphorus, silica, and manganese on a dried basis.

Bessemer Ore

The phosphorus content of Bessemer ore must not exceed 0.045 per cent; premiums were paid in 1962 ranging from $0.0080 a gross ton for ore with 0.044 per cent phosphorus to $0.7955 for ore with 0.002 per cent phosphorus.

Non-Bessemer Ores

Ores containing more than 0.045 per cent and not more than 0.180 per cent phosphorus are known as low phosphorus, non-Bessemer ores. Those containing more than 0.180 per cent phosphorus are known as high phosphorus, non-Bessemer ores.

Manganiferous Ores

Ore containing 2 per cent (natural) or more manganese may be marketed as manganiferous ore and receive some extra price recognition. Ore with 5 per cent (natural) or more manganese is recognized as standard manganiferous ore and carries a premium for the manganese in excess of 5 per cent. The manganese content between 2 and 5 per cent is generally combined with the iron units and priced as 'Old Range', non-Bessemer ores.

Siliceous

Ores containing 18 per cent or more silica (natural) are usually subject to penalties as agreed in the contracts between buyer and supplier. Practically all the ore used in North America has less than 11 per cent silica and most pellets contain considerably less.

Aluminiferous

A small amount of ore used in the past few years contained 8.5 per cent alumina but most ores contain from 0.5 to 3 per cent.

Calculation of Iron Ore Prices

The price or value of iron ore is calculated according to a formula adopted in 1925 for the value of a gross ton of 2,240 pounds of ore delivered at rail of vessel at Lower Lake Ports (Lake Erie). The base price is quoted for an ore type containing 51.50 per cent natural iron. The value of ore containing more or less than 51.50 per cent iron is calculated on the base unit value, which is determined by dividing the base price by 51.50, the number of units in the base ore. One per cent of iron in an ore constitutes one iron unit.

For ores containing less than 51.50 per cent and not less than 50.00 per cent iron, deductions are made from the base price for each unit or fraction of a unit of iron less than 51.50 per cent iron at the rate of the base unit value.

For ores containing less than 50.00 per cent and not less than 49.00 per cent iron deductions are made from the price computed for 50.00 per cent iron for each unit or fraction of a unit of iron less than 50.00 per cent at the rate of one and one-half times the base unit value.

For ores containing less than 49.00 per cent iron deductions are made from the price computed for 49.00 per cent iron for each unit or fraction of a unit of iron less than 49.00 per cent iron at the rate of two times the base unit value.

For ores containing more than 51.50 per cent iron addition is made to the base price for each unit or fraction of a unit of iron more than 51.50 per cent at the rate of the base unit value.

The following examples will illustrate how this formula is used.

The 1962 base price for Mesabi non-Bessemer ore was \$10.65. The base unit value for this ore was \$10.65 divided by 51.50, or \$0.20679. Using these figures, the delivered prices at rail of vessel Lower Lake Ports for ores containing natural iron contents of 52.50 per cent, 49.50 per cent and 48.75 per cent are given:

52.50 per cent Iron (Natural)

Base price 51.50 per cent iron natural =			\$10.65
52.50 – 51.50 = 1.00 unit @ \$0.20679 per unit =		+	0.20679
Price delivered rail of vessel Lower Lake Ports			\$10.85

49.50 per cent Iron (Natural)

Base price 51.50 per cent iron natural =			\$10.65
51.50 – 50.00 = 1.50 units @ \$0.20679 =	\$0.310185		
50.00 – 49.50 = 0.50 unit @ \$0.310185 =	\$0.1550925	–	0.4652775
Price delivered rail of vessel Lower Lake Ports			\$10.18

48.75 per cent Iron (Natural)

Base price 51.50 per cent iron natural =			\$10.65
51.50 – 50.00 = 1.50 units @ \$0.20679 =	\$0.310185		
50.00 – 49.00 = 1.00 unit @ \$0.310185 =	\$0.310185		
49.00 – 48.75 = 0.25 unit @ \$0.41358 =	\$0.103395	–	0.723765
Price delivered rail of vessel Lower Lake Ports			\$ 9.93

Further adjustments are made in price for low phosphorus content or for high manganese content and for other elements or physical characteristics mentioned above according to the terms agreed in purchase contracts. For instance, in 1962 Lake Superior pellets were purchased at rail of vessel Lower Lake Ports for \$0.252 per unit of natural iron.

The price of iron ore sold for export is determined in individual contracts between the buyer and producer in which ore specifications and price premiums or penalties are defined.

Effect of Minor Constituents

Silica, which is present to some extent in all ores, ranges from 6 to 12 per cent in most, but many pellets, concentrates, and some natural ores have less than 3 per cent. It gives an acid reaction in the blast furnace and is separated from the iron and removed in the slag by fluxing with lime or magnesia which have basic reactions.

Calcium and magnesium are usually present as carbonate or silicate minerals and are desirable constituents. Ores with appreciable calcium and magnesium may be self-fluxing because of the reaction of these constituents with silica to form slag. About 2 pounds of lime are required in the blast furnace to flux one pound of silica and more coke is required when ores of high silica and low iron content are used.

Alumina is usually present as clay minerals. It reacts as an acid constituent in the furnace with calcium and magnesium and is removed in the slag.

Manganese is a desirable constituent in ores used in basic open hearth processes. It is used advantageously in the basic open hearth process for deoxidation and removal of sulphur but its content in ores for acid open hearth or Bessemer processes must be below 0.5 per cent, as it tends to cause a 'puffy' slag.

Only limited amounts of *phosphorus* are acceptable in ores for smelting and steel-making. Phosphorus is reduced when the ore is smelted and practically all remains with the pig iron, and most of it must then be removed when the pig iron is converted into steel. Much phosphorus in steel causes brittleness and 'hot shortness', and a tendency to crack when being rolled. The amount of phosphorus that can be accepted depends on the steel-making process being used. High phosphorus ores are unacceptable for the North American open hearth furnace processes but are required for the Thomas process which is widely used in Europe.

A high phosphorus content in pig iron on the other hand gives high fluidity and permits pouring of very intricate and complex shaped castings although this cast iron is very hard and brittle.

Titanium is an undesirable constituent and ores with more than 0.1 per cent are usually not acceptable. It reacts as an acid in the furnace and when fluxed tends to make the slag highly viscous and difficult to control. It may also form accretions in the furnace that are not melted at ordinary operating furnace temperatures. Titaniferous ores are blended with other iron ores in some places but generally their use is carefully avoided.

Sulphur is an undesirable impurity and more than 0.1 per cent is usually not acceptable in ore for use in blast furnaces. It causes objectionable properties in steel similar to those caused by phosphorus.

Arsenic is reduced in the furnace and practically all remains in the metal. It causes undesirable properties in steel like sulphur and phosphorus, and is objectionable in steels used for canning foods because it is a poison.

Other elements. Chromium, nickel, and copper are undesirable in an iron ore as they accumulate in the hot metal and even small amounts have adverse effects upon iron and steel making processes. Vanadium collects in the slag and may be recovered as a valuable by-product if sufficiently abundant. Zinc and lead are generally not troublesome if present in small amounts. Tin collects in the iron and is most undesirable as it causes 'hot shortness' and makes the steel brittle. The combined amounts of Cu, Pb, Zn, As, V, Cr, and Ni should not exceed 0.05 per cent. The moisture content in an ore should not exceed 10 per cent.

Composition of Iron Ore Mined in Canada in 1961

The analyses of iron ores shipped from various mines in Canada in 1961 are listed in Table VI, which gives the different types of ore and their grades and illustrates how the iron ore classifications are used. Range in composition and structure of ore are also indicated.

Factors to be Considered in Evaluating Iron Deposits

Ore Grade and Structure

The composition of the various types of ore and the acceptable amounts of minor ore constituents as given in the previous section are minimum standards for direct-shipping ore for use in blast furnaces. Preferred ores now contain 60 to 66 per cent iron, less than 6 per cent silica, less than one per cent lime, magnesia, or alumina, very low phosphorus or sulphur, with rigid minimal restrictions on other minor constituents. The economic advantages of using high-grade ore in blast furnace operation make it profitable to concentrate many ores containing as much as 50 to 60 per cent iron. Of greater significance are the many low-grade deposits, such as iron-formations, that can now be mined and beneficiated to give very high grade ore concentrates at a cost less than that of directly treating material of higher natural grade. Because of this a very large number of low-grade iron occurrences, especially in central Canada, are now recognized as potential ore, and some of these have been developed for ore concentrates, while others have been selected for early development. In the Quebec-Labrador iron belt, for instance, low-grade metamorphosed iron-formations have been developed for ore concentrates.

The increased efficiency in blast furnace performance in recent years is attributed more to the physical condition of the concentrate, brought about by pelletizing or sintering, than to the increased iron content. The advantages of using pelletized ore rather than unprocessed ore have been so well demonstrated that the almost exclusive use of processed ore is anticipated. Furthermore fluxing constituents may now be added to the ore concentrate to produce self-fluxing pellets or sinter — an advance that is likely to be continued. By maintaining a reducing atmosphere in the pellet roasting furnaces and by adding carbon to the pellets, an appreciable amount of iron oxide ore can be reduced to metallic iron in this preliminary process, and the amount of iron in this product is therefore decidedly greater than that in the pure ore minerals.

The evaluating of a mineral deposit is no longer simply a matter of confirming existing grade and texture; the amenability of an ore to treatment and the processing necessary to provide marketable pellets, sinter, or sponge iron are prime factors.

Amenability to Concentration

To determine if material is amenable to concentration and treatment may require an involved and rather extensive series of tests, ranging from simple laboratory tests to large pilot plant operation. Concentration and treatment of iron ores has become a very complex metallurgical science and only some of the significant features can be mentioned here. The first essential steps are to obtain an adequate mineralogical description of the ore and a detailed geological description of the deposit. The former must be based on enough specimens so that ore zones of different quality may be delineated and variations in mineralogy and texture within these zones determined. Without an appreciation of the type of mineral intergrowth and the grain size, texture, and mineralogy of the ore appropriate test procedures

Table VI

Analyses of Canadian Iron Ore Shipped in 1961

Province and Company	Shipments Long Tons	Natural Fe	Fe	P	SiO_2	Mn	Al_2O_3	CaO	MgO	S	Loss by Ignition	Moisture	Size Range
NEWFOUNDLAND													
Dominion Steel and Coal Corporation, Limited													
Wabana hematite	2,292,369	50.38	51.38	0.90	12.93	0.10	4.7	3.3	0.65	0.03	—	1.94	
Iron Ore Company of Canada													
Bessemer	170,695	57.688	—	0.039	6.69	0.171	0.32	—	—	—	—	7.62	
Non-Bessemer	4,057,772	54.753	—	0.087	6.14	0.826	0.86	—	—	—	—	9.26	
Manganiferous	219,265	49.120	—	0.152	4.65	3.973	1.41	—	—	—	—	11.38	
Company total	4,447,732	54.588	—	0.088	6.09	0.956	0.87	—	—	—	—	9.30	
QUEBEC													
Hilton Mines, Ltd.													
Hilton pellets	800,215	66.02	66.73	0.0048	2.02	0.08	0.34	0.27	1.25	—	—	1.22	−5/8″
Iron Ore Company of Canada													
Bessemer	73,683	56.362	—	0.038	5.38	0.367	.68	—	—	—	—	8.18	
Non-Bessemer	2,615,937	52.041	—	0.068	7.08	1.167	1.26	—	—	—	—	9.83	
Manganiferous	306,247	47.610	—	0.068	8.15	5.232	1.07	—	—	—	—	9.60	
Company total	2,995,867	51.694	—	0.067	7.15	1.563	1.23	—	—	—	—	9.77	
Quebec Cartier Mining Company													
Concentrates	1,239,572	64.41	65.83	0.021	4.87	.06	.71	—	—	—	—	2.16	10 mesh
ONTARIO													
Algoma/Ore Properties Division													
Algoma sinter	1,640,919	50.36	50.63	0.015	11.19	2.84	1.95	4.31	7.09	0.098	.98	.54	
Algoma crude	3,213	33.0											
Caland Ore Company Limited													
Caland lump	18,320	54.86	56.52	0.024	9.48	0.40	1.00	—	—	0.02	—	2.93	+4″, −8″
Waldo	154,932	52.43	57.53	0.030	7.14	0.27	1.85	—	—	0.02	—	8.87	−4″
Raymond	836,104	52.41	57.84	0.029	6.94	0.31	1.89	—	—	0.02	—	9.39	−4″
Company total	1,009,356	52.46	57.77										

Canadian Charleson, Limited													
Charleson concentrate......	17,889	55.76	57.25	0.031	6.16	0.62	0.11	—	—	—	—	2.61	
Lowphos Ore, Limited													
Moose Mt. concentrate.....	577,890	59.81	66.34	0.015	7.18	0.43	0.31	—	—	—	—	9.85	fines
Marmoraton Mining Company, Ltd.													
Marmora pellets...........	528,865	65.1	65.1	0.007	3.8	0.083	0.73	1.4	1.1	0.02	nil	nil	−1″, +6 mesh
Steep Rock Iron Mines Limited													
Seine River, unscreened.....	808,203	53.600	58.64	0.030	7.06	0.25	1.63	0.11	0.10	0.031	7.48	8.59	−4″
Seine River, coarse.........	220,548	56.719	59.70	0.032	6.65	0.21	1.14	0.11	0.09	0.033	7.24	5.00	−4″, +3/8″
Seine River, fines..........	184,921	54.544	59.37	0.031	6.76	0.22	1.45	0.10	0.09	0.030	7.46	8.14	−3/8″
Company total............	1,213,672	54.206	58.93	0.030	6.94	0.24	1.52	0.11	0.10	0.031	—	8.07	
BRITISH COLUMBIA													
Empire Development Company, Limited													
Magnetite concentrate......	264,818	57.84	58.98	0.039	6.81	—	2.68	—	—	0.481	—	1.942	−6″
Nimpkish Iron Mines Ltd.													
Magnetite concentrate (sinter feed).............	378,243	59.1	61.84	0.014	5.10	—	1.71	—	—	0.44	—	4.86	−1/4″
Texada Mines Ltd.													
Magnetite concentrate......	442,234	61.42	64.74	0.011	3.93	0.13	0.86	2.25	0.64	0.349	—	5.38	−10 mesh
BY-PRODUCT IRON ORE													
The Consolidated Mining and Smelting Company of Canada, Limited....................													
Sinter*	40,865	65.0	—	0.029	3.7	—	—	—	—	0.131	—	—	
The International Nickel Company of Canada, Limited													
INCO pellets..............	230,995	68.0	68.0	0.03	1.5	0.05	0.65	0.25	0.30	0.01	—	—	
Magnetite concentrate......	403												
Noranda Mines, Limited													
Noranda calcine**.........	57,000	64-66	—	0.025	2.0	0.02	0.5 to 0.9	0.7 to 2.0	0.15 to 0.40	0.1	—	—	75% −200m.

* Own consumption. ** Also: Cu–.08 to .10; Zn–.07 to .10. Source: Data supplied to Mineral Resources Division by each company.
(Reference — Elver, R. B., and Wittur, G. E., 1962)

cannot be selected. The next step is to determine the grain size and nature of the ore texture. Coarse-grained ores are more easily concentrated than fine-grained ores provided that the elements present are in separate grains and that each grain is free of inclusions or intimate intergrowths. If the grains of ore can be liberated from the gangue without excessively fine grinding, concentration of the iron can usually be done effectively by one of several means. Coarse-grained low-grade ores of this type are easily concentrated by spiral or gravity methods, and with all the iron present as magnetite the ores are particularly easy to concentrate.

In the very fine grained cherty hematite ores constituent grains can only be liberated by fine grinding, and concentration by flotation is of increasing importance.

Ores composed of mixtures of iron oxides, quartz, and iron silicate or carbonate minerals are usually difficult to concentrate. Often the coarser grained minerals contain poikilitic inclusions of silicates in the magnetite or magnetite in the carbonates and silicates. Both liberation and concentration are difficult, and the fine-grained cherty ores of this nature are particularly difficult to concentrate. It may be impossible to liberate their mineral components even with very fine grinding and much material must be discarded if a high quality concentrate is to be obtained.

Fine-grained ores with an earthy texture are also difficult to concentrate without rejecting a large proportion of the iron. The mineral constituents are present in extremely small, clay-sized particles or even as a molecular intermixture. Separation of the iron by mechanical methods is usually impossible and pyrometallurgical or hydrometallurgical methods are required. These are expensive as the ore must be roasted to convert the iron to magnetite which can then be separated by magnetic or chemical-leaching processes. Equally difficult and costly processing is necessary for ores that have undesirable constituents combined chemically in the iron minerals.

Location

Traditionally, iron and steel plants have been built close to the source of iron ore and coal, and major industrial areas of the world have been developed close to the steel producing centres. The Midlands of England, the Ruhr Valley of Germany, and the Great Lakes area of North America are important examples of this development policy. In recent years steel industries have been developed in population centres close to where the steel is marketed and used. This is because more efficient methods of transportation and handling of the raw materials have made it less expensive in many cases to move the coal and ore to areas of steel consumption than to transport the rolled steel products to market areas. As many countries must import most of their iron ore, integrated blast furnaces and steel plants are normally placed in market areas near to ocean ports.

The advantages of having the source of iron ore close to steel plants and industrial areas are obvious but location of ore is not necessarily a predominant factor in selecting a site for a steel plant. When ore transportation costs are small, money saved can be spent on beneficiation and the treating of poor quality ore. On the other hand, ore deposits close to existing steel plants can no longer supply increasing ore requirements — the quality of ore may be unsatisfactory for modern steel-making processes or for economical beneficiation. The increased demand in North America for high grade ore of prepared structure is such that major ore deposits are being worked in distant parts of the continent, and high grade ore is being imported from Africa and South America.

Transportation

Despite these considerations, charges for transportation and handling are a major consideration in the evaluation of an ore deposit, and special studies must be carried out for each deposit and mining area, as every proposed mine development presents a different set of circumstances. Transportation of bulk products has become a highly specialized part of industry. Methods are being constantly improved so that ore is now moved over remarkably long distances. In many mine areas costs for construction of railways, harbours, and ore storage facilities are greater on a per ton basis than costs for mining and beneficiation. Some features of the common modes of transportation are mentioned here.

Ore loaded on an ocean vessel can be transported very inexpensively and the shipping costs per mile decrease considerably as the haulage distance increases. Decidedly greater efficiency is being achieved in ocean transportation by using larger ore carriers, and vessels carrying up to 70,000 tons of ore are now in use and some with a capacity of 100,000 tons are being constructed. Much iron ore is now shipped between continents because of the low cost of ocean transportation.

Railways used for ore haulage must be designed to accommodate heavier cars and equipment than are normally used for other freight traffic, indeed existing railways may need reconstruction. Nevertheless an all-rail haul has some advantages. The cost of transfer from rail to ship and back may cost more than a longer rail haul without the extra handling. Also, with an all-rail haul, ore can be supplied from the mine to the steel plant as required, without having to stockpile ore near steel plants where space is limited.

Many new methods for bulk handling have been investigated in recent years. The use of conveyors and large trucks for trackless haulage has been increased for moving ore short distances. Furthermore, extensive research has been carried out on the possible use of pipe lines for transfer of ore concentrates with very encouraging results. In many northern areas where shipping and ore handling are restricted by seasonal conditions, transportation costs are higher, as facilities must be provided to move a year's supply of ore in six months and expensive equipment may be idle the rest of the year.

Markets

Most steel companies in North America own and operate their own mines. This assures an adequate and stable supply of desired quality ore at a steady and reasonable price. As ore from different deposits are blended into a product suitable to give uniform and satisfactory operation of the furnaces, most companies maintain a number of mines with different types of ore. Where sufficient ore of a particular type cannot be supplied from a company mine, it may be acquired by bartering with another supplier or by direct purchase.

An attempt is made to anticipate specific ore needs for tens of years in advance, and long range planning of ore and raw material supply is considered essential. New mines and properties may be acquired to maintain an adequate supply of the different varieties of material as ore needs vary depending on the production program of the individual steel manufacturer. Possibilities of increasing production from existing mines, or of obtaining ore from new sources all influence the plans for ore development and the way in which companies manage their reserves.

Mine ownership by the steel companies and the captive market so created make it most difficult for independent mines to operate unless long-term contracts can be secured. With heavy investments in the development of an ore source a company will naturally want to recover its investment and continue working a developed deposit as long as possible.

Marketing conditions and patterns may change considerably with the increased use of pellets and processed ore. Pelletized ore can be prepared with fairly uniform grade and structure and less and less blending is necessary for ores so treated. More companies in future will probably be interested in purchasing a uniform pelletized ore at the prevailing price without too much concern about its source. However, ore consumers will presumably maintain part ownership or financial control of ore sources in order to be assured of an adequate ore supply.

European companies have not participated in the ownership of their ore supply to the same extent as the North American companies. In Europe much more ore is purchased from foreign sources on a short-term contract basis. The supply of this ore is highly competitive and comes from many source areas.

Japanese companies obtain most of their ore by short-term purchase contracts but in recent years have loaned or invested money in mine developments for the guarantee of an ore supply.

Methods of Smelting and Reduction

About 90 per cent of the iron produced in North America is smelted in blast furnaces and most of the ore required must fit the specifications and requirements for this method. Nearly all the rest of the iron ore is smelted in electric furnaces of various types and designs. The molten iron or pig iron produced in blast furnaces is converted to steel by a number of different processes, including the Bessemer acid or basic processes, open-hearth converters, the direct oxygen steel-making method, or in electric furnaces.

Many other smelting methods different from standard blast furnace techniques have been developed and some are used on a small scale in various parts of the world. They are referred to as direct-reduction processes, and include methods of processing iron ore to provide metallic iron or partly reducing ore suitable for making steel or for iron castings. These direct-reduction processes have been developed to provide an economical method for making iron in areas where only small scale production is warranted and where other low cost sources of energy and reducing agents could be used instead of coking coal. They are intended to provide a substitute for scrap iron or a metal product that can be used with iron from blast furnaces or to provide small amounts of iron for making steel in electric furnaces or small batch processes.

Only some of these methods are used for commercial production of iron and then only in especially favourable circumstances. Although it is unlikely that any of these processes will compete seriously with blast furnace operation, they may be used more in the future to produce small amounts of iron in special circumstances. Some, for example, may be of use in producing very high grade ore concentrates containing a high proportion of metallic iron. Many kinds of crude ore can be utilized in direct-reduction processes and for some purposes ores of marginal quality can be used. They also provide a large variety of metallic iron products or sponge iron. Information on a number of the direct-reduction processes is summarized in Table VII.

Other Factors

When considering a proposed development project, a source or potential source of power is a necessity. Other major considerations are technical skills and labour to operate mines; favourable living environment for the mine staff; well-defined government policies regarding land tenure, property development, and taxation; and a secure political system to ensure continuance of established policies.

Trends in Technological Developments

What amounts to a technological revolution in blast furnace and steel-making practices has led to such sweeping revisions in the specifications for acceptable ore that future demands are difficult to predict. One trend is to concentrate the ore and reduce it to metal near the mines, and to process the metal and produce steel in the industrial market area. Another trend is to develop further the traditional practice of integrating ore refining, reduction, steel-making, and processing in the industrial market area. Many steel-makers who have to import iron ore may prefer to purchase iron metal or some type of sponge iron, and process it to meet the special needs of local markets. These possible trends in the iron industry are of fundamental significance with regard to the use of raw materials. Not only is metallic iron worth much more than natural ore as an export product but other resources in mine areas, such as power, fossil fuels, flux minerals, and manpower, could be put to use.

A brief review of trends in iron ore specifications over the past ten years gives some idea as to why a change to greater production of metallic iron in the mine areas may take place.

About ten years ago an increasing trend towards improving ore structure became apparent. Fines and dust-size particles were removed by screening so that

Table VII

A Summary of Information on Prominent Direct Reduction Processes

Process Name	Type of Equipment	Reductant (*)	Reductant Produced How	Ore	B or C	Final Product	Temperature	Pressure	Developer, Licenser, etc.
Dwight-Lloyd McWane	"Flying Saucer" pellitizer, D-L sintering machine, electric furnace	solid carbon fines	....	**	C	molten iron	ignition-about 2000°F	atm	McDowell Co
Esso Research Little	fluidized-bed reactor, 3-stage	carbon monoxide hydrogen	partial oxidation of nat. gas or oil with air	various -⅛″	C	sponge or briquettes 85 % reduced	1450-1600°F	1-4 atm	A. D. Little Inc.
Freeman	rotary kiln, co-current fired	solid carbon fines	....	mag. or hem. conc's ½-1″ pellets	C	sponge pellets, 91.5 % Fe°	up to 1850°F	atm	Freeman Corp.
H-Iron	fluidized-bed reactor, 3-stage	hydrogen	partial oxidation, water-gas shift. scrubbing	hi-grade mag. conc's millscale	C gas B sol	low-C iron powder, 98 % reduced	reactor, 1000°F; passivation 1200-1600°F	500 psig	Hydrocarbon Research Inc. Bethlehem Steel
Höganäs	sealed "saggers" in tunnel kiln	solid carbon fines	....	rich ore, sinter, pellets	B	sponge, as sintered porous cake	about 2100°F	about atm	Höganäs-Billesholms, Sweden
HyL	fixed-bed reactor	hydrogen, carbon monoxide	catalytic steam reforming of nat. gas	rich lump ore, sinter, pellets; up to 1½″ max. 20 %-¼″	B	sponge iron 85 % reduced	about 2000°F	about atm	Hojalata y Lamina; M. W. Kellog Co.
Ralling-Domnarvet	rotary furnace, internally fired	solid carbon fines	....	various	C	sponge iron	**	atm	Stora Kopparberg
Krupp-Renn	rotary kiln, countercurrent fired	solid carbon fines	....	hi-silica low-grade ores	C	"luppen" 92 % reduced iron may be hi-S	reduction 600-1100°C slag exists 1200-1300°C	atm	Fr. Krupp, Germany; Southwestern Eng. Co. U.S.A.
Nu-iron	fluidized-bed reactor, 2-stage	hydrogen	optional: partial oxid'n cat. steam ref'g etc.	hem. or mag. -10 mesh	C	powder of briquettes 90-95 % reduced	1300°F	15-50 psig	U.S. Steel Corp.

Orcarb	pre-heating kiln, balling drum, rotary kiln, electric furnace	coking coal fines	....	mag. conc's low P & S 50%-100 mesh	C	molten iron -2% carbon	1700-1900°F in reduction kiln	atm	Swindell-Dressler Corp.
R-N	rotary kiln, countercurrent and internally fired	solid carbon fines	....	various -1″ plus 20 mesh	C	briquettes or sponge, about 90% reduced	1830-1970°F	atm	R-N Corp. (Nat. Lead Co. & Repub. Steel Corp.)
Strategic-Udy	rotary kiln, countercurrent fired and elec. furnace	solid carbon fines	....	various -3/8″	C	molten iron 1-2.5% C.	solids leave kiln about 1700°F	atm	Strategic Mat'ls Corp. Koppers Co.
Wiberg Soderfers	fixed-bed shaft furnace	carbon monoxide hydrogen	electric-arc-heated carburetor from coke and hydrocarbons	mag's some hem's 2½-8 cm	C	sponge iron	max. about 1000°C	slightly atm	Stora Kopparberg, Sweden & N.Y.
ACAR (Allis-Chalmers)	travelling grate, dual concentric kiln-within-kiln	hydrogen, carbon monoxide	cracking of hydrocarbons w. steam or air within kiln	hi-grade conc. pellets ½-¾ in. dia.	C	iron pellets 81% Fe°	about 1920°F	atm	Allis-Chalmers Mfg. Co.
S-L	rotary kiln	natural gas	gas and pre-heated air, coal, fuel	conc. pellets	C	sponge iron	2000°F	atm	Steel Company of Canada
Direct Steel	travelling grate with reduction of ore and conversion to steel	CO, H_2, & gas mixture	separate source	very high purity iron oxide conc.	C	steel rods or wire	2100°F	....	Ontario Research Foundation
Jet Smelting	special furnace	natural gas	....	hi-grade granular concentrate	C	high carbon pig iron	3600°F	....	Ontario Research Foundation

Various ELECTRIC FURNACE methods for producing iron are not listed

B — batch

C — continuous

* — "Solid carbon fines" indicates in general coke breeze, char., etc., anthracite fines or bituminous coal of subcoking grade

** — Not specified in literature examined

From — McAneny, Colin C., 1960, with additional information the remaining coarse fragmental ore could be subjected to higher intensity hot air blasts in the furnace and thus increase the rate of the reduction reactions. The fine fraction of the ore was sintered to provide an ore with controlled structure and grade. These treatment practices were continued and the great benefits in blast furnace operation gained by improving ore structure were more fully appreciated. It was also realized that concentrating iron ore increased furnace efficiency and output tremendously, and that it was economically possible to spend a good deal on such processes. It further became obvious that there were more efficient ways of concentrating natural ore than by pyrometallurgical processes in the blast furnace.

A great deal of experimentation took place to determine if sinter was as satisfactory as pellets. Many kinds of natural iron ore could be mixed and blended in sinter whereas high-grade ore concentrates were needed for pellets; also very high grade concentrates could be produced from the metamorphosed iron-formation at low beneficiation costs. In Canada, with its vast reserves of metamorphosed iron-formation, much capital was invested to develop a supply of concentrate for markets in the United States. It was fully expected that iron ore in the form of hematite or magnetite sand or granular concentrates would be shipped for blending and sintering at the blast furnace sites. A number of steel plants consequently enlarged their facilities for sintering and an increased amount of ore was beneficiated before being smelted. Recently the debate as to whether pellets or sinter provided the most efficient furnace operation seems to have been decided in favour of pellets, and a large number of pelletizing plants have been built.

Not only is ore grade and structure in the final product being rigidly controlled, but research and pilot plant tests show that greater economy can be achieved by mixing dolomite and flux materials with the ore to make self-fluxing pellets or sinter. Recent work on pelletizing in a reducing roasting atmosphere shows that partly reduced pellets can be produced inexpensively and blast furnace operators in the near future will no doubt require pelletized ore that contains 80 per cent or more iron. This means that 10 to 15 per cent of the iron in the ore must be reduced prior to reaching the blast furnace.

Along with these changes in ore specifications for blast furnaces have come achievements in the development of small-scale direct-reduction processes. The possibility of using direct-reduction methods to supply metallic iron instead of scrap for use with hot metal from the blast furnace has received considerable attention. The price and supply of scrap iron fluctuates rapidly and direct-reduction methods could supply metallic iron from ore at a predictable price and in suitable quantities. Furthermore, direct iron processes could supply iron metal in which the quantity of trace and alloy elements is controlled and specifically known. The amount of these constituents in scrap iron supplied is rarely well known and generally difficult to manage. Furthermore, the cyclical use of scrap iron from sources such as automobile bodies or other machinery might lead to such an accumulation of undesirable minor elements that most scrap iron would be unacceptable.

Requirements for sponge iron are very limited at present, but advances being made in blast furnace operation could lead to radical changes in methods of steel production. Such changes are, however, unlikely to be speedily accepted as steel companies have huge investment in processing and pelletizing equipment and in furnaces designed to utilize these products. This investment must be recovered. It is probable therefore that a decided increase in the use of beneficiated ore, such as partly reduced pellets can be expected, as these do not require radical changes in equipment. Replacement of blast furnaces for the production of large quantities of iron in a central market area is unlikely, but increased use of direct-reduction processes in conjunction with existing blast furnaces and steel-making equipment is probable.

New methods and techniques for both ore processing and steel-making will probably be utilized in the design and installation of new steel plants and consequent changes in existing industrial patterns will take place. Saving in transportation costs can be realized by shipping iron metal rather than natural ore, and direct-reduction processes located near the mines could produce sponge iron or metal to substitute for scrap iron. These processes would provide a very high grade iron ore or superconcentrate, consisting of 90 per cent metallic iron.

Chapter V

IRON-FORMATIONS AND BEDDED IRON DEPOSITS

Stratigraphic units containing three to ten times as much iron as the average content in the earth's crust are found on all continents and in rocks of all ages. They vary considerably in their characteristics and obviously formed in different environments. Extensive study of these rocks during the last 50 years has provided a voluminous literature and much is known about their mode of occurrence and specific characteristics. Nonetheless, some aspects of their origin are not understood and disagreement regarding the source of the iron and mode of transportation of iron and silica still exists. Problems of this nature are particularly difficult to answer as sediments similar to most types of iron-formation are not known to be forming at present.

Ninety per cent or more of the potential iron ore in Canada either occurs within or is directly associated with iron-formations and bedded iron deposits.

Definition of Iron-Formation

Many different names have been used for bedded ferruginous rocks and many separate local names have been introduced. The use of these names, however, has not always been restricted to type-areas or to the same kind of material in other areas, and as a result many terms are used without general agreement as to where and how they should be applied.

Layered ferruginous rocks of Precambrian age are known as iron-formation, itabirite, taconite, banded iron ore, hematite or magnetite quartzites, banded ironstone, calico rock, jacutinga, specularite schist, and by a number of other names. Layered or bedded iron-rich rocks of Palaeozoic age are referred to as Clinton-type ore, oölitic hematite beds, flaxseed ore, bedded iron deposits, bedded ironstone, and banded iron-formation. Many younger beds are known as Jurassic ironstones, Minette ore, Lorraine ores, Salzgitter-type ore, ironstone, oölitic iron deposits and by other names that have locality connotations.

To most geologists terms such as iron-formation or Lake Superior banded iron-formation, Clinton-type ore, Lorraine- or Minette-type ore indicate fairly specific types of material with a limited range in mineral composition and texture.

These types are reasonably well established through consistent use of these terms. There is much to be said in favour of using names established for ore material in certain well known localities for similar material in other deposits, provided the type deposits are adequately described. However, the fact that several rather distinctive mineralogical and lithological facies are present in formations that are referred to by a single name such as Lake Superior banded iron-formation is often overlooked. The use of type names can thus be more misleading than helpful.

Much emphasis has been given to the belief that certain types of layered ferruginous rocks are restricted to specific periods in geological time and that therefore unique conditions prevailed during those periods. It is true that some types of iron-formations are much more abundant in rock successions of certain periods but they are by no means restricted to them. For example, thin-banded chert and iron oxide beds that resemble the iron-formations in early Precambrian rock belts are associated with volcanic rocks of Palaeozoic and Mesozoic age, the Clinton-type beds are present in rocks ranging in age from Precambrian to Mesozoic (O'Rourke, 1961). As more thorough investigations are carried out in different parts of the world a greater variety of ferruginous sediments are being found in all parts of the geological column.

Many insist that the term iron-formation must be used only for Precambrian rocks composed of alternating bands of silica- (chert or fine-grained quartz) rich material and bands rich in iron minerals, but such restriction presents several difficulties. For example, thick siderite or hematite beds that contain very little silica are intimately associated with cherty iron-formation in areas where the whole stratigraphic unit is referred to in the literature as iron-formation. James (1954) has pointed out some of the difficulties of trying to use the term iron-formation in a restricted sense and notes that many rocks intimately associated with cherty iron-formations in the Lake Superior region, especially in the Cayuna Range, do not have cherty layers. He concluded that "neither the presence of chert nor the mineralogy of the iron can enter into the definition, although both may be incorporated as modifying adjectives."

The problems of nomenclature are much too involved to be resolved here, but iron-formation is a useful general descriptive term. It is obvious that a concise lithological description of individual formations is necessary if a particular type of material is to be recognized. The term iron-formation therefore is used here in a general way to include all stratigraphic units of layered, bedded, or laminated rocks that contain 15 per cent or more iron, in which the iron minerals are commonly interbanded with quartz, chert, or carbonate, and where the banded structure of the ferruginous rocks conforms in pattern and attitude with the banded structure of the adjacent sedimentary, volcanic, or metasedimentary rocks (Gross, 1959). An iron-formation may therefore be a chemical or clastic sediment or layered rocks that have been replaced by iron minerals during diagenesis or during a later period of their history. Banded or layered iron-rich parts of intrusive rocks and discordant replacement masses are not considered to be stratigraphic units and are excluded by this definition.

Iron-Formation Facies

Sedimentary rocks reflect their environment of deposition by their composition, mineralogy, texture, and primary structures. Most sediments are composed of clastic material oriented and arranged in layers, and the degree of sorting and nature of the primary features are determined by factors such as their position in the sedimentary basin, depth of water, wave and current action, source areas, and so forth. The detrital grains in many sedimentary rocks of this type show no evidence of chemical reaction or alteration in response to the depositional environment and therefore give little information about the chemical environment in the sedimentary basin. The situation is very different in the chemically precipitated sediments that form in direct response to their environment. They, by their composition, mineralogy, and the oxidation state of the various elements, show much about the chemistry of their deposition.

Most of the iron-formations are chemical precipitates and are particularly sensitive indicators of their depositional environment. Iron occurs in different valence or oxidation states, ferrous and ferric ions, and combines readily with most of the other rock constituents. Although the chemical state of iron in most clastic sediments may be easily overlooked, it is a prominent feature in most iron-formations. The environment of deposition of an unmetamorphosed iron-formation therefore is suggested by the characteristics of the iron-formation itself and by the other types of associated rock. Because of this close relationship between the mineralogy and depositional environment for chemically precipitated ferruginous sediments the various lithological types can be satisfactorily classified as distinctive sedimentary facies related to chemical factors.

The depositional environment for chemical sediments has been studied in detail by Krumbein and Garrels (1952) who have demonstrated that the pH and Eh of the aqueous medium are major factors in controlling the kind and amount of chemical deposition. James (1954) applied these concepts in his study of iron-formations in the Iron River district of Michigan, and to other iron ranges in the Lake Superior region. He demonstrated four distinctive facies of iron-formation (the oxide, silicate, carbonate, and sulphide) that apparently formed in environmental conditions corresponding to fields of stability defined by Eh and pH factors outlined by Krumbein and Garrels (*see* Fig. 1). The iron-formation facies were concisely shown and described by James for Superior-type iron-formations, but they can be applied to practically all other chemically precipitated iron-formations and have been used as a fundamental classification for these sediments.

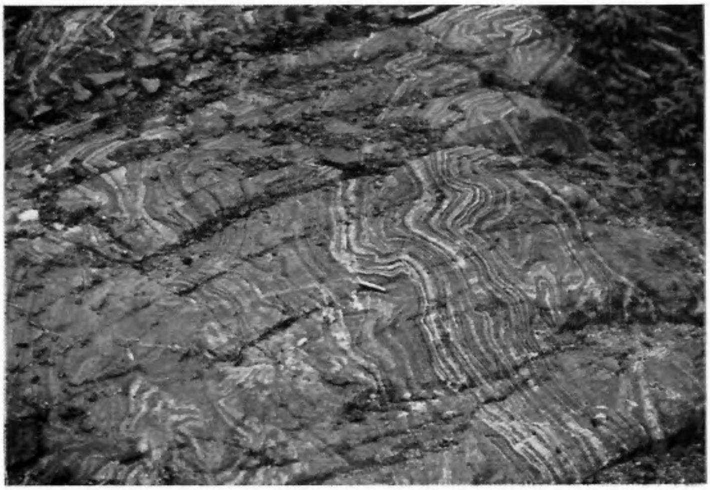
In applying the facies concept it is imperative that only primary mineral assemblages be considered in order to define the environment of deposition. Most Precambrian iron-formations show some evidence of metamorphic alteration so that complete definition of primary facies is not always possible. It is recognized, however, that sufficient evidence can usually be seen in the relict textures and mineral paragenesis of metamorphosed beds to indicate the type of material from which they were derived and much can be inferred about their mode of deposition.

PLATE II

TYPICAL FACIES OF ALGOMA-TYPE IRON-FORMATION

A. Magnetite-siderite-chert facies, Algoma District, Ontario.

B. Transition from magnetite-chert facies on left to iron-silicate chert facies on right, Timagami area, Ontario.



C. Highly deformed magnetite-chert facies, Boston township, Ontario.

Table VIII

Average Chemical Composition of Sedimentary Facies of Iron-Formation

	Oxide Facies					
	A Magnetite Banded Rocks	B Hematite Banded Rocks		Carbonate Facies	Silicate Facies	Sulphide Facies
SiO_2	48.515	40.10	44.008	34.508	51.302	36.670
Al_2O_3	0.965	0.80	0.883	1.090	2.833	6.900
Fe_2O_3	26.390	50.10	38.245	1.845	7.142	—
FeO	15.705	1.60	8.653	31.720	24.805	2.350
MgO	2.210	2.00	2.108	3.146	3.760	0.650
CaO	1.175	1.40	1.288	1.162	0.535	0.130
Na_2O	n.d.	n.d.	—	0.040	1.180	0.260
K_2O	n.d.	n.d.	—	0.200	2.090	1.810
H_2O-	—	—	—	—	0.442	0.550
H_2O+	—	—	—	—	3.960	1.250
H_2O	1,455	—	1.455	0.876	—	—
CO_2	3.910	2.6	2.600	22.960	4.906	—
P_2O_5	0.070	0.70	0.385	0.512	0.490	0.200
S	—	—	—	0.050	—	—
MnO	—	0.20	0.200	1.371	2.08	0.002
TiO_2	—	—	—	0.072	0.52	0.390
C	0.060	—	0.060	1.880	0.21	7.280
FeS_2	—	—	—	—	—	38.700
V_2O_5	—	—	—	—	—	0.150
SO_3	—	—	—	—	—	2.600
Organic matter	—	—	—	—	—	0.320
Total Fe	29.845	—	—	26.57	19.302*	—

* Total Fe + Mn

Oxide facies

A Average of five analyses of taconite specimens from different horizons of the Biwabik Formation, Mesabi Range.

B Analysis of shipment from Millie Iron Mountain, Mich. for 1936 (Lake Superior Iron Ore Association 59) recast into standard oxides (Gruner, 1946, pp. 56-58).

Carbonate facies — Seven analyses of carbonate iron-formation used and averages of each component extrapolated. Samples obtained from: a banded chert-carbonate iron-formation, Iron River district, Michigan (James, 1951a); a carbonate iron-formation in Marquette district (Van Hise and Bayley, 1897); a carbonate iron-formation in Gunflint district, Minnesota (Irving and Van Hise, 1892); a carbonate iron-formation in Gogebic Range (Irving and Van Hise, 1892); a carbonate-silicate iron-formation in the Crystal Falls district, Michigan (Pettijohn, 1947); and a "fine grained, light grey, siliceous iron carbonate rock from Kennedy Mine, near Cayuna, Minn." (Harder and Johnston, 1919).

Silicate facies — Seven analyses of silicate iron-formation presented by James (p. 271) were extrapolated and an average chemical analysis of the silicate facies determined. Specimens were taken from: greenalite rock (Leith, 1903) of Mesabi Range; taconite of the lower cherty division of the Biwabik iron-formation, Mesabi Range (Gruner, 1946); taconite of the lower slaty division of the Biwabik iron-formation, Mesabi Range (Gruner, 1946); greenalite rock containing other silicates and chert of the Mesabi Range (Jolliffe, 1935); magnetic ironstone of the Iron River district (Shapiro, L., U.S.G.S.); and laminated rock similar to the magnetic ironstone from a Hiawatha drill-hole, Iron River district (James, 1951).

Sulphide facies — From James (1951), a complete chemical analysis of the pyritic slate zone on the 10th level in Buck mine of the Iron River district, Michigan.

Oxide Facies

The predominant iron minerals in the oxide facies are hematite and magnetite, which may occur in separate bands or in various proportions in single bands. Depositional environments with oxidizing conditions, where the Eh values are positive and the pH ranges from weakly acid to alkaline, are most favourable for the precipitation of hematite or the hydrated ferric oxide goethite. Hematite may still form under conditions of low negative Eh value if pH is about 6 or greater. The formation of magnetite is favoured in an intermediate to mildly reducing environment where Eh values range from low positive to negative and where pH values range from weakly acid to alkaline. A low CO_2 partial pressure is considered necessary, otherwise siderite may form instead of magnetite in the stability field indicated. Where the Eh is close to zero the formation of magnetite may be dependent on the ratio of ferrous and ferric ions contributed to the depositional environment and with a low oxygen partial pressure in the system the oxidation of ferrous iron would be at a minimum.

The oxide facies of iron-formations is commonly separated into two types, the hematite and the magnetite. Both minerals occur as primary constituents and a complete transition from one end member to the other occurs.

The chemical and physical environment most suitable to account for the composition and texture of the facies appears to be that of fairly shallow water containing considerable oxygen, where there is sufficient wave or current action to promote the development of granules and oölites. The precipitation area would generally be free of clastic sediments although in many areas clastic deposition is imposed on the chemical facies to give a mixed rock. In other areas fine-grained clastic beds are interbedded with iron-formation. Most oxide facies in the late Precambrian and younger iron-formations have granular or oölite textures but early Precambrian formations apparently do not.

Information on the kind of carbonate present in hematite facies with a cherty matrix is still sparse, but calcite, dolomite, and ankerite rather than siderite are usually found and these are the carbonates expected to form under the conditions postulated for the primary deposition of hematite. Magnetite appears to form where waters contain less oxygen. It is commonly associated with ferrous iron silicates and siderite and generally, in transition zones from hematite to magnetite to silicate or siderite facies, very little hematite is associated with magnetite beyond the stage where silicate or carbonate minerals begin to appear.

In general the ferric and ferrous iron content vary as the amount of hematite and magnetite present (*see* Table VIII); Al_2O_3, MgO, and CaO are each usually under 3 per cent; MgO and CaO are present as carbonate; the manganese content is very low; the iron and manganese content is lowest in the facies with the highest ferric iron content; phosphorus, sulphur, and titanium are very low; and sodium and potassium are present in almost insignificant quantities.

Carbonate Facies

The most prominent carbonate mineral in this facies is siderite, and relatively pure siderite beds are common. In the Michipicoten area of Ontario thick bands of siderite, recognized as carbonate facies, are associated with siderite-pyrite material and this mixture is transitional across the section to thin-banded siderite and quartz and ultimately to siderite-magnetite-quartz iron-formation.

Garrels (1960) pointed out that the presence of siderite in a sediment indicates strongly reducing conditions and the presence of CO_2 in more than atmospheric amounts. He also indicated that fluctuations in deposition of siderite and magnetite may be brought about by fluctuations in pH without a change in oxidation conditions in a system closed to CO_2, but in an open system, where the partial pressure of CO_2 is constant, the change from siderite to magnetite cannot take place without a change in the partial pressure of oxygen.

Conditions favourable for the deposition of siderite would probably be found in deep and quiet water where there is little current or wave action to mix oxygenated surface waters. The area of carbonate deposition would be farther from shore than the area for oxide deposition and probably in cooler water, highly charged with CO_2, perhaps derived to a large extent from the decay of organisms and fine organic debris. Siderite may form over a broad enough set of chemical conditions to permit considerable variation in detail of basin and environment. James pointed out that a granular or oölitic texture is absent in the carbonate iron-formation he studied in the Lake Superior region, and he believed that this is evidence for quiet-water conditions undisturbed by waves or currents. Oölitic limestones are known to form in shallow oxygenated water and agitation of the water is considered essential to the development of oölites. Many iron-formations contain granules or even oölites of siderite that have abundant siderite in their matrix. In nearly all these considerable iron silicate, magnetite, and chert or fine clastic and carbonaceous material is present and the facies apparently formed in a transitional zone between oxide and siderite facies conditions.

Most of the iron present in the carbonate facies is in the ferrous state. Calcium and magnesium are somewhat higher than in the oxide facies (*see* Table VIII). One of the most significant features in carbonate iron-formations is a higher manganese content than that typically present in the oxide facies.

Silicate Facies

Silicate facies are recognized by the distinct green or olive-green to khaki colour imparted by the iron-silicate minerals. Of all the facies of iron-formation, the depositional environment for silicate minerals is least understood and the factors controlling their formation are poorly defined. This is mainly due to the number and complexity of these minerals and the fact that primary silicate minerals are difficult to distinguish from low-rank metamorphic silicates. Known to be of primary origin are greenalite in cherty iron-formations, some minnesotaite, and probably some stilpnomelane; chamosite and other iron-chlorites are recognized as primary in some facies; glauconite forms as a primary or a diagenetic mineral. Practically all other iron-silicates are of metamorphic origin but appear to be derived from primary iron-silicates.

Iron-silicate minerals are commonly associated with magnetite and siderite and the silicate facies apparently forms in an environment common to parts of the oxide and carbonate facies. Silicate minerals of high iron content in cherty iron-formations are commonly associated with magnetite in beds or layers but rarely directly with hematite, although adjacent layers may consist of mixtures of hematite and magnetite. This association, like the widespread association with siderite, indicates that a reducing environment is necessary for the formation of iron-silicate minerals. Most of the iron in these minerals is in the ferrous rather than the ferric state, which further testifies to a reducing environment. The CO_2 content is probably very critical in deciding whether iron-silicates or siderite forms, as a high CO_2 partial pressure would probably lead to precipitation of the iron as siderite rather than as a silicate.

Facies of iron-formation that contain abundant fine-grained clastic material, magnetite, and occasionally siderite may be regarded as a silicate facies if iron silicates are the most abundant mineral constituent. This facies commonly appears to be thin-banded, black magnetic shale but on microscopic examination is found to consist of a high proportion of iron-rich chlorite, clay minerals, or minnesotaite. The silicate minerals in some rocks of this facies occur in thin streaky layers that alternate with relatively pure chert layers, and in others they are disseminated through relatively thick cherty beds. Common are granular textures in which greenalite or fine minnesotaite forms dense, massive, greenish, cryptocrystalline to microcrystalline, oval-shaped granules in the chert layers. Oölites with concentric layers of greenalite and chert or carbonate are apparently absent and it is possible that the granules formed as gelatinous globules rather than through accretionary growth in the way normal oölites form. Chamosite and some of the iron-rich chlorite and clay minerals are present in the concentric rings of oölites and the silicate minerals may perhaps be sensitive indicators of subtle differences in environment.

Table VIII illustrates the predominance of ferrous over ferric iron and the typical higher magnesium content of the silicate facies. The manganese content also is usually greater than in adjacent oxide facies.

Sulphide Facies

The sulphide facies is the least abundant of all facies of iron-formation. Because of its lack of resistance to weathering it is usually poorly exposed and frequently overlooked, but it is found, even if in small amounts, with all major types of iron-formation regardless of age. Lithologically, sulphide facies occur in the form of disseminated pyrite in black carbonaceous shale; pyrite and pyrrhotite mixed with siderite and carbonate in banded chert or quartz; distinct pyrite and pyrrhotite beds with minor siderite; and oölitic-textured pyrite members in black shale.

The formation of a pyrite-rich facies is favoured by a strongly reducing environment with an abundant H_2S and a high ferrous iron content. The pH value is not considered to be critical although pyrite and pyrrhotite apparently form most readily in weakly acid to alkaline waters and marcasite under stronger acid conditions. Variables affecting the formation of pyrite and pyrrhotite include H_2S and CO_2 content, Eh and pH, iron content, abundance of organic carbonate, and other factors. Equilibrium conditions have been dealt with in considerable detail by Garrels (1960). Nearly all sulphide facies contain considerable carbon and commonly have more carbonate or fine clastic material than chert. Many siderite facies contain considerable sulphide material, and carbonate-rich sulphide facies are not uncommon and apparently formed in transitional zones where environmental conditions of the sulphide and carbonate facies overlapped or alternated.

Most sulphide-rich facies lack granular texture and other evidence of agitated water, although a distinctive thin oölitic pyrite bed is present in the rock succession above the hematite-chamosite-siderite beds at Wabana, Newfoundland. Apparently pyrite beds formed, in most areas, in the deeper parts of sedimentary basins where there was little agitation of the water, where strongly reducing conditions permitted preservation of the organic carbon, where H_2S was released by bacterial action, and where deposition of clastic material took place too slowly to mask or inhibit the accumulation of pyrite. Table VIII illustrates some of the typical chemical features of this facies, such as the high carbon and Al_2O_3 content, lack of ferric oxide, ferrous iron mostly present as FeS_2, the presence of SO_3, and the higher content of K_2O and TiO_2, probably absorbed in clay constituents.

General Classification of Iron-Formations

Nearly all iron-formations in the world can be classified in six major types, based on lithological features, geological setting, and associated sedimentary rocks, and each type is named after a locality or a dominant physical feature. Each type represents rather specific environmental conditions but these conditions are developed in many of the world's tectonic belts and are not restricted to a single geological period.

Although distinct differences separate the major types, some have features in common that permit grouping. Four of the six are primarily chemical precipitates and two are clastic sediments. Of the four chemical precipitates, two are predominantly siliceous and two predominantly aluminous (*see* Fig. 3). Each major type can be further subdivided into the facies described above. All four facies may be present in any type, but usually only one or two are common in any one occurrence.

Three of the chemical precipitate types were deposited in shallow-water environments and in these oölitic textures are characteristic. The fourth type occurs with volcanic rocks in a eugeosynclinal environment (*see* Fig. 3). Nearly all cherty iron-formations are closely associated with volcanic rocks but no such relationship is obvious for aluminous iron-formations. The diagram (Fig. 3) is presented to illustrate some of the prominent iron-formation types and their general relationships in depositional environment and composition.

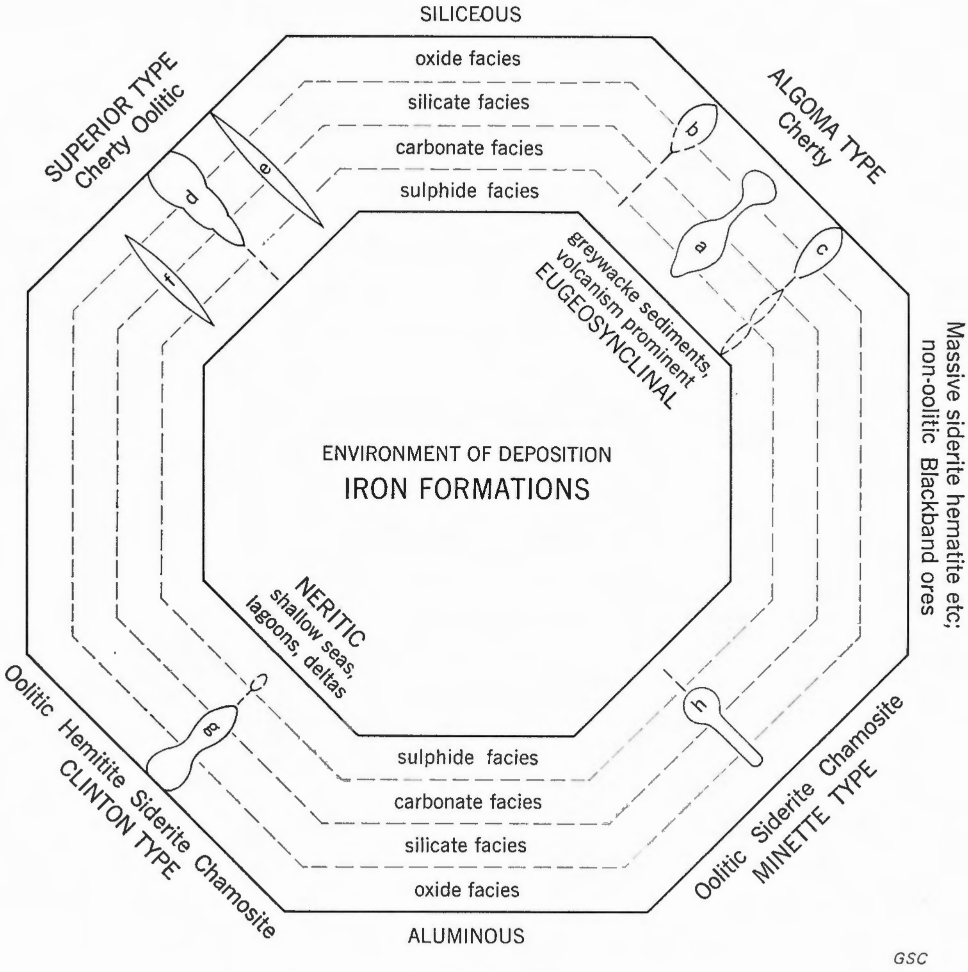


FIGURE 3.

Diagram showing major types of chemically precipitated iron-formation with their sedimentary facies and depositional environment. Examples are from: a, Michipicoten, Ontario; b, Moose Mountain, Ontario; c, Timagami, Ontario; d, Knob Lake, Labrador and Quebec; e, Iron River, Michigan; f, Gunflint and Biwabik iron-formation, Ontario and Minnesota; g, Wabana, Newfoundland; h, Clear Hills, Alberta.

Principal Types

Algoma Type

This type is present in nearly all early Precambrian belts of volcanic and sedimentary rocks in the Canadian Shield as well as in a few similar rock belts of Palaeozoic or Mesozoic age. It is characteristically thin banded or laminated with interbands of ferruginous grey or jasper chert and hematite and magnetite. Siderite and carbonate iron-silicate or sulphide facies are present locally or in small amounts. Occasionally, as in the Michipicoten area of Ontario, massive siderite and pyrite-pyrrhotite beds form part of the formation. Single iron-formations range from several hundred feet to inches thick and are rarely more than a few miles long. Usually a number of lenses of iron-formation are distributed *en échelon* within a volcanic belt. They are intimately associated with various volcanic rocks including pillowed andesites, tuffs, pyroclastic rocks, or rhyolitic flows and with greywacke, grey-green slate, or black carbonaceous slate. Tuff and fine-grained clastic beds or ferruginous cherts are interbedded in the iron-formation and detailed stratigraphic successions show heterogeneous lithological assemblages. These iron-formations have streaky lamination or layering, and oölitic or granular textures are absent or inconspicuous except in post-Precambrian rocks. The associated rocks indicate a eugeosynclinal environment for their formation and a close relationship in time and space to volcanic activity.

Superior Type

This type of iron-formation is characteristically thin-banded cherty rock with iron-rich layers that are typical of the various sedimentary facies. Granules and oölites composed of both chert and iron minerals are a typical textural feature and the rocks are practically free of clastic material, except in the transitional border zones or in distinct, well-defined members within the formation. The alternate or rhythmic banding of iron-rich and iron-poor cherty layers, which normally range in thickness from a fraction of an inch to several feet, is a prominent feature. Individual layers may pinch and swell to give a wavy banded member or the uniformity of the layering may be disrupted by nodular or stubby lenses of chert and jasper, by rare occurrences of crossbedding, or by cherty forms resembling in shape and structure "Collenia" or "Cryptozoan" growths in limestones formed by algae colonies. Tension, syneresis, and desiccation cracks are present in some chert granules and nodules, and stylolites are common. The textures and sedimentary features of this type of formation are remarkably alike in detail wherever examined although certain sedimentary features are more prominent in some formations.

The close association of this type of formation with quartzite and black carbonaceous shale, and commonly also with conglomerate, dolomite, massive chert, chert breccia, and argillite, are recognized throughout the world. Volcanic rocks, either tuffs or flows, are not always directly associated with Superior-type iron-formation but they are nearly always present somewhere in the succession. The sequence dolomite, quartzite, red and black ferruginous shale, iron-formation, black shale, and argillite, in order from bottom to top, is so common on all continents that some have been led to believe that it is invariable. However, stratigraphic studies have shown that, although there is a persistent association of these sedimentary rocks, the succession may differ in local areas; it does so for example in the Labrador geosyncline. Quartzite and red to black shale generally lies below the iron-formation and black carbonaceous shale above it, but the presence of other sedimentary rocks and their position in the stratigraphic succession may vary from place to place, even in a single range or rock belt.

Continuous layers of the Superior-type iron-formations commonly extend for hundreds of miles along the margin of a geosynclinal basin. The formations may thicken and thin from a few tens of feet to several hundred feet and occasionally to over 1,000 feet but their persistence is truly remarkable. The rock successions in which the iron-formations occur usually lie unconformably above highly metamorphosed gneisses, granites, or amphibolites, and the iron-formations are, as a rule, in the lower part of the succession. In some places they are separated from the basement rocks by only a few feet of quartzite, grit, and shale, or, as in certain parts of the Gunflint Range, they lie directly on the basement rocks; in most areas however they occur at least some hundreds of feet above the base.

The Superior-type iron-formations are present in late Precambrian rocks in nearly all parts of the world and also in some early Palaeozoic rocks (O'Rourke, 1961). They apparently formed in fairly shallow water on continental shelves or along the margins of continental shelves and miogeosynclinal basins, and consist of sediments derived from the adjacent landmass and also some material from the volcanic belts within the basin. It is still uncertain whether the iron and silica in this type of iron-formation were derived from a land or a volcanic source.

This type of siliceous formation is the protore or host rock for the rich hematite-goethite orebodies of the Lake Superior region, of Quebec-Labrador, and for many other major iron deposits of the world.

Clinton Type

These formations are typically deep red to purple, massive beds, with oölitic textures, and are composed of intimate mixtures of hematite, chamosite, and siderite. The oölites are formed around fossil fragments and debris, and in some facies around clastic grains of quartz. They differ from the Superior type by the absence of chert, the silica present being mainly in iron-silicate minerals and small amounts in clastic grains. They contain about 51 per cent iron and are typically higher in phosphorus than the Superior type. Some beds of this type, which form distinctive fossil ore, are composed of shell fragments and debris partly replaced and cemented by hematite or carbonate.

The Clinton-type iron-formations are associated with carbonaceous shale, sandy shales, dolomite, and limestone; they tend to be in lenticular beds usually 10 to 15 feet thick but never more than 40 feet. In marginal zones the iron-formations are commonly in thin layers interbanded with shale. In some areas they are in sharp contact with limestones or dolomites and are very rich in fossil debris. Individual beds may thicken and thin within short distances to form ore shoots that fill channels scoured in the underlying sediments by current and wave action. In places depositional bars and lenses may be composed of iron-rich oölites. Lenses of oölitic ferruginous material of this type occur along one horizon or within a restricted part of the stratigraphic succession for hundreds of miles in a tectonic belt although only locally may the lenses be thick enough to be mined. In Silurian rocks, for example, Clinton-type iron-formations are present in the State of Wisconsin, in southern Ontario, and almost continuously from northern New York State to Alabama.

Clinton-type iron-formations apparently formed along the margins of continents, on continental shelves or in shallow parts of miogeosynclines. They are common in rocks ranging in age from Cambrian to Devonian, but some of Precambrian age are present in the Transvaal of South Africa and one of the best known examples are the Ordovician beds at Wabana, Newfoundland.

Minette Type

This type is particularly important and widespread in Europe, but the only Canadian examples known are the iron-formations of the Peace River area, Alberta. The sideritic or chamositic ironstones of northern England, the minette ores of the Lorraine district of France and in Western Germany, the Salzgitter ores of Lower Saxony, and many other Jurassic beds in northern Europe all belong to this type. They are most abundant in Mesozoic and Tertiary rocks.

The oölitic-textured rocks of Minette-type iron-formations are composed mainly of siderite and iron silicates such as chamosite or iron chlorites, and of limonite-siderite and chamosite. Many beds contain a high proportion of fine-grained clastic material and are transitional to sideritic or chamositic mudstones or sandstones. Sand grains may form the nuclei of siderite oölites, as in the Salzgitter and Peace River beds; occasionally fossil fragments are present in the oölites. Silica, which is usually more than 20 per cent, generally is present in chamosite or iron-silicate minerals or in lesser amounts in clastic grains, rather than chert. The iron content of the Minette-type iron-formation is usually less than 40 per cent; the lime content from 5 to 20 per cent, which is high compared to other iron-formations; and the phosphorus content is higher than that in the cherty Superior and Algoma type.

Minette-type formations are closely associated with black carbonaceous shale, mudstones, and sandy shales, which apparently formed in marine or brackish water in shallow basins. Cyclic sedimentation with repetition of the sequence — calcareous siltstone, siderite siltstone, sideritic limestone, and calcareous limonitic (or chamositic) oölite — has been noted by Taylor (1951) in a section in southern England. He suggested that the limonitic facies may be altered sideritic or chamositic beds where oxidation was penecontemporaneous with deposition. Cyclic sedimentation in the Yorkshire Lias beds was also reported by Hemingway (1951).

The Minette-type formations have many features in common with the Clinton-type iron-formations and are characteristically thin lensy units. In general they have a higher silica and lower iron content, the ferric oxides are usually brown and limonitic rather than red hematite, siderite is more abundant, and they contain more lime, and most have more alumina. Carbonate and silicate facies are much more prevalent in Minette-type formations than in the other types and, like the Clinton type, facies are not well segregated but are mixed on a microscopic scale in some oölites.

Non-oölitic Type

A number of stratigraphically concordant but lithologically diverse iron deposits are grouped under this name. Most consist of lenticular beds composed of siderite or sideritic mudstones, siderite-hematite, or massive hematite, and limonite. Bedding as a rule is poorly developed or lacking and massive units are typical. Most formations of this type contain less than 25 per cent iron; they are siliceous and high in alumina or lime. Some beds contain up to 15 per cent manganese; the phosphorus content varies. Most are associated with dark shales or greywackes, limestones, and volcanic tuffs.

Many of these iron-formations are considered to be the product of volcanic exhalations of iron and silica but in some beds the iron may be derived from the alteration of glauconite or iron carbonates.

This type of formation is not known to occur in Canada but European examples include the blackband clay ironstones associated with coal beds, the hematite and goethite beds locally rich in manganese and silica in the Lahn and Dill River valleys of Hesse, Germany, and siderite-hematite beds near Vares, Yugoslavia.

Clastic Iron-Formations

These may be composed of various iron minerals, but magnetite is the most common. Many consist of sandstones carrying titaniferous magnetite or appreciable amounts of monazite, rare earth minerals, zircon, rutile, or other high gravity minerals. They are characteristically lensy beds varying in grade and quality, deposited as beach sands, placer bars, and fans. The iron-formation in the Belly River sandstone near Burmis, Alberta, is the best known example in Canada.

Table IX

Characteristic Features of the Different Types of Iron-Formation

1. *Algoma type*: Thin-banded chert or quartz and iron oxides, silicates, carbonates and sulphides, associated with volcanic rocks, and greywacke; usually in thin lenses with streaky banding; oölitic textures absent or inconspicuous; relatively rare but in rocks of all ages, especially early Precambrian; eugeosynclinal; e.g., Moose Mountain and Michipicoten area, Algoma district, Ontario.
2. *Superior type*: Interbanded chert or quartz and iron minerals with prominent granular or oölitic textures; associated with dolomite, quartzite, black carbon-bearing shales and slate, chert breccias and volcanic rocks; usually well-defined formations of broad regional extent; form in continental shelf and miogeosynclinal environments. Extensively developed in late Precambrian rock groups, e.g., Lake Superior and Labrador iron ranges.
3. *Clinton type*: Hematite-chamosite-siderite formations; interbedded with carbonaceous shale, carbonate rocks and sandstones; high in phosphorus compared to cherty formations; oölitic to granular texture; miogeosynclinal and shallow-basin environments; lenticular beds restricted to thin formations of broad regional extent as in the Appalachian belt. Most common in rock groups of Cambrian to Devonian age, e.g., Wabana, Newfoundland; Birmingham, Alabama.
4. *Minette type*: Siderite-chamosite-iron silicate, and goethite beds; usually high in phosphorus; oölitic textures; lenticular beds containing clastic material, sideritic mudstones, and concretionary siderite layers; associated with sandstones, carbonaceous shales, siltstone and limestone; cyclic sedimentation of sandy beds, shale and ironstones, and black shale, common; formed in transition zones between marine- and brackish-water environments. Most common in Mesozoic and younger rocks, e.g., Peace River area of Alberta, Canada; Jurassic beds in England; Minette ores of Lorraine, France, and in Luxembourg and Belgium.
5. *Non-oölitic*: A heterogeneous group of non-oölitic, lensy, ferruginous beds; vary in composition and type from the blackband sideritic carbonaceous claystones associated with coal; to the Lahn and Dill hematite and goethite beds locally rich in manganese and/or silica; to siderite-iron silicate-goethite lenses in shale. Associated with fine-grained clastic or volcanic rocks; occurring mainly in late Palaeozoic and younger rocks, e.g., Lahn and Dill River area in Hesse, Germany; Vares, Yugoslavia; Gross Ilsede and Peine in Lower Saxony.
6. *Clastic*: Placer deposits of black sands; mainly magnetite with siderite or hematite in sandstones, usually contain titanium and rare earth elements; thin beds of lensy distribution, varied grade, and quality; deposited along beaches or near shore marine environments, e.g., iron-formation near Burmis, Alberta, in the Belly River sandstone.

Origin of Iron-Formations

Differences between the various types of iron-formations described in the foregoing indicate diversity in mode of origin. All these types are chemical precipitates and have many features in common so that the fundamental problems regarding the source, mode of transport, and depositional environment of the iron and silica are common to all. Many accurate empirical predictions and inferences that can be based on the distribution, occurrence, characteristics, and variations in different types of iron-formations, are important guides in the development and use of these ferruginous beds as ore. The theoretical aspects of their origin, however, are more subject to debate. Except for the clastic beds, iron-formations are not known to be forming anywhere in the world today and, because of this, many have concluded that they formed under circumstances unique to certain periods of the past.

Much emphasis is given in the literature to the special nature of these beds and to their limited occurrence in rocks of certain ages. As iron is the fourth most abundant element, composing about 5 per cent of the earth's crust, whereas calcium composes about 3.63 per cent, it might be expected that concentrations of iron should be at least as abundant as those of calcium. However, differences in chemical properties between the two elements cause most of the iron to be disseminated through a wide range of sediments and calcium, mainly, to separate and be precipitated as limestones and lime oozes.

Relatively limited information is available about the geological setting, the nature and kind of associated sediments, and the general tectonic framework of the areas in which many iron-formations occur. An appreciation of these features is essential if the mode of origin of these rocks is to be understood, and indeed the answers to a number of pertinent problems may depend more on studies of broad features of the geological environment than on studies of the iron-formations themselves. So far investigations have mainly been centred on the chemistry of iron and silica in natural environments, on possible methods of transportation, and on factors that control the solution and deposition of these elements.

Iron-formations, except for a few relatively unimportant clastic beds, are chemical precipitates with negligible or at most small amounts of intermixed clastic sediments. As the iron content of iron-formations is three to ten times greater than the average content of 5 per cent computed for the crust of the earth, the fundamental question is, where did the extra iron come from and was the source similar for all types of iron-formations? In the cherty formations the question is even more complex, as the association of iron with abundant silica, although their chemical properties are so very different, must be explained. How were both iron and silica taken into solution, transported, and precipitated in the immense quantities present in the iron ranges? What factors caused the precipitation of abundant iron and silica and, at the same time, only small amounts of other common elements? Alumina is scarce in the cherty iron-formations but fairly plentiful in the other types. What are the environmental factors that controlled the formation of such primary features as rhythmic bedding, oölites, and granules, and why are these features

Table X

Composition of Iron-Formation Types

	I	IIa	IIb	IIc	III	IV	V	VI	VII	VIII
SiO_2	45.20	60.51	48.50	51.12	11.42	28.63	31.91	14.7	9.45	12.16
Al_2O_3	0.250	1.20	0.79	1.31	5.07	6.20	6.79	10.8	6.80	6.42
Fe	36.70	25.22	29.14	26.11	51.85	35.63	—	—	—	—
Fe_2O_3	—	—	41.60	10.81	61.83	—	35.45	3.3	1.97	1.69
FeO	—	—	5.46	23.86	11.00	—	—	35.3	41.01	39.38
CaO	1.06	0.67	0.01	0.75	3.32	0.70	3.79	5.4	6.15	2.26
MgO	1.59	0.52	0.60	3.26	0.63	0.71	6.94	3.8	4.17	3.89
Na_2O	—	—	0.03	trace	—	—	0.74	—	0.06	—
K_2O	—	—	0.01	trace	—	—	0.25	—	0.03	0.37
H_2O+	—	trace	2.46	3.43	1.94	7.74	—	3.1	4.50	1.41
H_2O-	—		0.09				—	—	—	0.68
TiO_2	—	—	—	—	0.015	0.27	4.35	0.3	0.40	—
P	—	—	—	—	0.863	0.422	—	—	—	—
P_2O_5	—	0.05	0.04	0.04	—	—	0.26	1.1	1.44	0.47
H_3PO_4	—	—	—	—	1.970	—	—	—	—	—
MnO	—	—	0.11	4.77	0.17	—	—	0.8	0.47	0.95
MnO_2	—	0.92	—	—	—	—	0.34	—	—	—
Mn	0.04	—	—	—	—	0.28	—	—	—	—
CO_2	—	—	0.06	—	2.15	—	—	20.6	23.00	29.38
C	—	—	0.07	—	—	0.63	—	0.2	0.30	0.54

S	0.024	0.59	0.01	—	0.023	0.009	0.137	0.2	0.06	trace
H_2SO_4	—	—	—	—	0.058	—	—	—	—	—
Cr_2O_3	—	—	—	—	—	—	—	—	0.03	—
V_2O_3	—	—	—	—	—	—	—	—	0.09	—
As	—	—	—	—	—	—	—	—	0.02	—
Cu and Ba	—	—	—	—	—	—	—	—	—	trace
Ignition loss	—	—	—	—	—	—	8.81	—	—	—

I Algoma type: average analysis of samples taken across No. 2 deposit of magnetite ore, Moose Mountain area.
Report of the Ont. Iron Ore Committee, 1923, p. 206, *Ont. Dept. Mines.*

II Superior type:
IIa Gunflint: average composition of a large part of the Gunflint iron-formation.
Broderick, T. M., 1920, Economic geology and stratigraphy of the Gunflint iron district, Minnesota; *Econ. Geol.*, vol. 15, No. 5, p. 445.
IIb Labrador: average analysis of seven iron-formation chip samples collected by G. A. Gross, *Geol. Surv. Can.* Analyst, G. Bender, GSC.
IIc Mesabi: computed average composition of the main Mesabi district.
Gruner, J. W., 1946, The mineralogy and geology of the taconites and iron ores of the Mesabi Range, Min.; *Iron Range Res. and Rehab. Commission*, St. Paul, Min., p. 127.

III Clinton type — Wabana: analysis of composite sample of Wabana ore, Dominion Wabana Ore Limited, W. H. Pearson and Company.

IV Minette type — Peace River: chemical analyses averaged from three ferruginous sandstones of southern Clear Hills deposits.
Kidd, D. J., 1959, Iron occurrence in the Peace River region, Alta.; *Res. Council Alta.*, Geol. Div., Prel. Rept. 59-3, p. 24, Nos. 13, 14, 15.

V Clastic type — Burmis, Alta.: Chemical analyses of nineteen Belly River iron-rich sandstones. Analyses by H. Wagenbauer, *Res. Council Alta.*, from —
Mellon, G. B., 1961, Sedimentary magnetite deposits of the Crowsnest Pass region, southwestern Alberta; Bull. 9, *Res. Council Alta.*

VI and VII Minette type — Yorkshire, England:
Dunham, K. C., 1959, Syngenetic and diagenetic mineralization in Yorkshire; *Proc. Yorkshire Geol. Soc.*, vol. 32, pt. 3, No. 11.

VIII Clay Ironstone — Yorkshire, England.
Dunham, K. C., 1959, Syngenetic and diagenetic mineralization in Yorkshire; *Proc. Yorkshire Geol. Soc.*, vol. 32, pt. 3, No. 11.

present in some iron-formations and absent in others? Does the predominance of certain iron-formations in rocks of a particular geological age indicate unique conditions not since repeated or were they formed in more common circumstances but fortuitously preserved?

The possible role played by organisms in the formation of iron-formations is a most tantalizing question. The extent to which organic compounds affected the solution, transportation, and deposition of iron and silica cannot be fully appraised until much more is known about the palaeoecology of depositional basins, but the abundance of carbon associated with these beds suggests that organisms and organic materials were abundant.

Little is known concerning the amount of time required for the deposition of beds of a given thickness. Positive recognition of primary, diagenetic, and metamorphic features in these rocks is a prerequisite for studies on genesis, yet criteria for recognizing such features are not well established.

Résumé of Existing Theories

Many writers have proposed many answers to the questions outlined above. The diversity of these answers is striking, but in part at least is due to the lack of a clear definition of the particular iron-formation being discussed. This is particularly true of the cherty Precambrian iron-formations where Algoma and Superior types have been considered together, and yet may have been formed by different processes.

Prior to the twentieth century most geologists assumed that the iron and silica in the iron-formations were derived from a landmass by weathering; their views have been summarized by Leith (1903).

In an exhaustive study of the Lake Superior iron-formations Van Hise and Leith (1911) expressed serious doubt as to whether the iron in these rocks was derived from the weathering of ordinary rocks. If it were, they argued, iron-formations should be found on the same scale in younger rocks deposited on Precambrian terrains similar to those underlying the Lake Superior formations. They were further concerned about the fact that rocks deficient in iron and silica are not apparently present and something should represent the residue of the material from which the vast amount of iron and silica present in the iron-formations was leached. They concluded that the iron came from a more direct or unusual source and suggested the hot solutions from basalt flows, hot or cold meteoric waters dissolving iron from basic extrusive rocks, or sea water, the iron being derived from submarine flows. Elaborating on this idea Leith (1934) suggested that the material in the iron-formation was first deposited as carbonate and greenalite interlayered with chert, and that with exposure to weathering and oxidation the iron in these minerals changed to hematite or limonite resulting in the formation of ferruginous cherts and jaspers. Only minor amounts of iron were thought to be deposited as iron oxide.

The ideas regarding the origin of the Lake Superior iron-formations developed by Van Hise and Leith were anticipated to some extent by Winchell (1900). He suggested that lavas reacting with the sea water contributed silica and iron, which was either precipitated directly as iron-formation or went into solutions that later replaced and silicified basalt flows or obsidians to form jaspilite.

A different view about the origin of this type of iron-formation was expressed by Harder and Chamberlain (1915) who suggested that the precipitation of ferric oxide might be due to

> . . . purely chemical reactions taking place in the sea, or perhaps to the operation of the well-known iron bacteria which cause the deposition of ferric hydroxide from waters containing ferrous carbonate in solution. These iron bacteria are said to possess the peculiar property of utilizing as food the carbon dioxide locked up in very dilute solutions of ferrous carbonate. Ferric hydroxide is left behind and is deposited as a sediment. The characteristic reaction may be written: $2FeCO_3 + 3H_2O + O = 2Fe(OH)_3 + 2CO_2$. This process is operative in very dilute solutions. Apparently only two or three parts of iron per million are needed to make certain types of the iron bacteria active.
>
> Not having much confidence in the hypothesis that the iron oxide was precipitated directly from sea-water by ordinary chemical means we prefer to turn to the iron bacteria as perhaps forming a better working hypothesis.

In a discussion of the nature and origin of the Biwabik iron-formation of the Mesabi Range, Grout (1919) described a number of primary features and made several inferences about the origin and diagenetic alterations of these beds. He concluded that:

> . . . the deposition of taconite probably occurred in shallow water mainly by precipitation through organic processes, resulting in lean ferruginous cherts with more or less iron-bearing carbonate, ferric oxide, and greenalite. Alternation with periods of precipitation came periods of solution, leaching, oxidation, and wave action, producing intraformational conglomerates, and granular rocks much higher in iron than the original precipitate, and some layers of pure ferric oxide.
>
> The richer deposits of magnetite occur so characteristically in the granule and conglomerate zones that it seems probable that the primary leaching was a determining factor in the development of primary iron concentrations from lean ferruginous cherts, and greenalite rock.
>
> When the sediments were covered with other layers, there may have been more or less concretionary rearrangement. Deep burial under slates, developed heat and pressure that recrystallized a good deal of the formation.

Very complex diagenetic changes in the iron-formations were reported by Grout and Broderick (1919). They indicated that the iron was precipitated as carbonate or silicate, altered to magnetite, and some later oxidized to hematite with the possible reduction of some of this in turn back to magnetite.

Gruner (1922) was concerned with the importance of decaying organic matter and the organic compounds in peat water in the solution and transportation of iron and silica. He conducted a long series of experiments on the chemical nature and solubility of iron and silica in solutions from decaying plants and found that they dissolved all oxides and carbonates of iron and most of the silicates but did not seem to attack pyrite appreciably. Ferric iron compounds are reduced to ferrous compounds in peat solutions in the absence of air but are not reduced in ordinary carbonic acid. When a solution of a neutral salt with a strong acid radicle comes into contact with natural organic colloids, acid is set free which dissolves much iron and silica. Among other results he found that the ratio of iron to silica in peat solutions was very like that in the undecomposed minerals. He concluded that large amounts of iron are carried as organic colloids or adsorbed by them and not in true solution as a bicarbonate, and that silica is transported in colloidal form protected by organic colloids. Electrolytes such as salts of sea water appeared to be ineffective as precipitants of the colloidal iron and silica where protective organic colloids were present. From the results of these experiments, Gruner suggested that the source of iron and silica in the Biwabik iron-formation was a large landmass covered mainly with greenstone lavas and basalts. He presumed the climate to be humid and probably tropical or subtropical, causing abundant land vegetation of a low form which resulted in rapid decay of the rocks and the solution of large quantities of iron and silica. He concluded that most of the iron was carried to the basin of deposition as organic colloids or adsorbed by organic colloids and was precipitated with colloidal silica by the action of algae and bacteria. In his opinion very little silica and iron were derived or precipitated by inorganic reactions and he marshalled strong evidence against the concept of Van Hise and Leith of the origin of these elements directly from hot lavas. He pointed out that hot springs containing 100 ppm iron and 300 ppm silica would have to contribute about 524,000 cubic miles of solution to supply the constituents for the Biwabik iron-formation. In contrast a river such as the Amazon carries this much iron and silica to the sea in solution in 176,000 years and the constituents could be derived from weathering to a depth of only a few feet in a river basin of its size.

Further interest in the behaviour of iron and silica in natural solutions prompted Moore and Maynard (1929) to conduct a number of experiments that were most revealing. They found that carbonated water is able to dissolve sufficient iron and silica from basic rocks to form a large sedimentary iron deposit and that organic matter if present also assists in dissolving much iron and silica. Indeed inorganic acids, under exceptional circumstances, can cause much iron and silica to go into solution. They concluded that the iron was transported principally as a ferric oxide hydrosol, stabilized by organic matter, and that the greater part of the silica was transported as a silica hydrosol. They found in their experiments that ferric oxide and silica hydrosols taken separately are stable but when mixed mutual precipitation occurs unless organic matter is present. They concluded that ferric hydroxide is precipitated in a few days when colloidal iron and silica solutions, stabilized by organic matter, come in contact with electrolytes of sea water, whereas silica is not completely coagulated even after several months. They suggested that thin banding in the iron-formations may result from differential settling of these constituents according to this mechanism, with newly supplied material causing repetition of the banding. In summarizing they say "... the results of laboratory work indicate how banding may be developed but they do not show anything which is adequate to explain the solution and transportation of such relatively large quantities of silica in cold waters. It seems necessary to invoke the direct aid of igneous activity on a large scale to account for the very siliceous chemical deposits. The other stratified iron deposits appear to be the result of normal sedimentary processes in which life has played an important part."

An explanation of the cause of banding in the silica iron oxide formations was offered by Boydell (1928) who said "... this banding is much easier explained by mutual reaction between oppositely-charged particles of iron oxide and silica in colloidal solution, supplemented by sedimentation, than in any other way."

Collins and others (1926) who studied the origin of associated siderite masses and banded cherty iron-formation interlayered with acid and basic rocks in the Michipicoten area of Ontario concluded that

> ... apparently the iron-formations were formed by ascending heated mineralized waters at many loci in a land area of great volcanic activity. These loci ranged in area from a few yards to 7 or 8 miles in major horizontal extent, and they are to be found throughout the vertical range of the volcanic complex. The heated waters were mineralized with carbon dioxide, iron, silica, and sulphur compounds at least. They permeated the volcanic rocks and converted them to a maximum depth of nearly 250 feet into carbonates and sulphides. In so doing they were protected from atmospheric oxidizing influences. At the surface they probably spread out in depressions and by evaporation and cooling precipitated their content of silica and such carbonate as was not deposited below ground. It is not absolutely certain that the banded silica deposits were formed above ground, but the stratified arrangements of the silica and iron compounds forming them, and their great horizontal extent as compared with their thickness, seem to require this postulate.... In the case of the Michipicoten iron-formations, therefore, the banded silica member, found at the surface, may be due, first, to extraction of large quantities of silica from the volcanic rocks by ascending hot, acid waters, and to subsequent precipitation of this silica at the surface, by evaporation, and second, to the solutions capacity for silica having become lessened by loss of carbonic acid and sulphur compounds.

A number of others, including Grout (1926), Moore (1931), Bruce (1940), Hawley (1942), and Tanton (1948), differed in their opinions regarding the nature and time of replacement of the volcanic rocks to iron-formation. Goodwin (1961) restudied these cherty carbonate-sulphide formations and concluded that they have a direct and intimate relationship to the volcanic development of the area. They are believed to have originated through chemical sedimentation of banded chert, pyrite, and carbonate members in a submarine volcanic environment as a result of large-scale hot spring and fumarolic activity.

Dunn (1941) studied the iron-formations in Singhbhum, India, and his view was "... that the beds were originally deposited as fine-bedded ferruginous tuffs (which became oxidized), and more normal ferruginous sediments, and that surface and sub-surface silicification of these more or less contemporaneous with their deposition, under the action of solutions partly magmatic in origin, gave rise to the banded jaspers and cherts. The cherty nature of the silica was determined by the low temperature conditions of deposition." Spencer and Percival (1952), on the other hand, believed (1) that these beds are of sedimentary origin and supported their belief with descriptions of a number of primary structures that indicate that the beds were deposited as colloidal precipitates; (2) that the iron and silica may have been taken into solution from basic rocks under the influence of algae and bacteria, carried in colloidal form, partly oxidized, and precipitated as a mixed colloid; (3) that the granular forms in the texture are inorganic spherulites rather than algae or bacteria forms; and (4) that iron- and silica-bearing volcanic solutions may have contributed directly to these sediments.

A very stimulating discussion on the origin of iron-formations was offered by Woolnough (1941) who suggested

> . . . the banded iron deposits of the older pre-Cambrian throughout the entire world represent epicontinental sediments formed as chemical precipitates from cold natural solutions in isolated closed basins on a land surface that had been reduced to the last limit of peneplanation.
>
> Analogies are drawn with the 'duricrust' of Australia, formed on a surface of somewhat less ideally perfect peneplanation during an era of low rainfall with marked seasonal incidence. The exceedingly small probability of repetition of periods of such exceptionally perfect base-leveling is considered to explain the uniqueness in character of deposits of this type.

MacGregor (1927) believed that the reason for the broad distribution and abundance of cherty iron-formations lay in the fact that the early Precambrian atmosphere was high in carbon dioxide and low in oxygen and that this favoured the solution of iron and silica on exposed land. When oxygen was produced by algae with chlorophyll it led to oxidation of the ferrous iron in solution to ferric iron and extensive precipitation, forming the jaspilite beds.

Bruce (1945) did not regard the cherty iron-formations as unique Precambrian rocks and pointed out that ". . . the Precambrian Iron Formations, restricting the term to those rocks that contain no clastic material, are so similar in many other respects to sediments of later epochs that it seems impossible that the processes under which they were formed were unique and occurred only in Precambrian times. It is more likely that accumulation and precipitation took place under some specially favourable combination of circumstances that occurs seldom, but that is not restricted to any age."

Tanton (1950) put forth the view that the banded iron-formations are of igneous origin and not sedimentary and says ". . . they meet the host rocks in which they occur as irruptive bodies or primary replacement deposits." He believed that the iron and silica separated as immiscible liquids.

Sakamoto (1950) has considered a possible mechanism by which iron and silica are derived from a landmass under conditions of mature weathering and laterization. He suggested that these constituents are deposited in rhythmic bands resulting from alternation of seasonal conditions between arid and humid in a monsoon-like climate.

> The cause of banding in the Pre-Cambrian banded iron ores is sought in a cyclical deposition of colloids due to a periodic change of reaction (pH) of lake water. The two component poor ores, with alternating bands of ferric iron oxides and silica, are the product of an

> acid-neutral semi-cycle; while the three-component poor ores, with alternating bands of the above two plus a third of iron silicates (often of iron carbonates), are the product of an acid-alkaline complete cycle. The interrelation between these two types of ores seems to be one of primary sedimentation and not of a later oxidation of the latter into the former.
>
> Both iron oxides and silica are the products of a mature weathering. . . . Iron migrates in acid environments and is precipitated in neutral and alkaline environments. Silica, on the contrary, migrates in alkaline and is precipitated in acid environments. . . . The soil solution above the water table, except in arid regions or seasons, is acidic, the acidity being highest in the cold zone in higher latitudes; whereas the solution below the water table, disregarding the climatic zones and seasons, is always alkaline (pH 8 ±). . . . Supposing that the annual rainfall was great and divided into periodically wet and dry seasons, shallow grounds above the water table became neutral or even alkaline in a dry season and the phreatic zone near the water table became neutral or temporarily even acidic in a wet season. (This is the condition which obtains under the present tropical monsoon.) In wet seasons, iron migrated in the acidic surface water, while in dry seasons, silica migrated in the alkaline ground water. . . .
>
> In a lake (basins and depressions separated from the open sea with only low barriers) where the reaction of water repeats regular cycles of acid-neutral-alkaline-neutral-acid, the conditions under which silica and iron are precipitated are inferred as follows:
>
> The iron in the form either of colloids or of true solution is supplied to the lake in wet seasons. Ferrous iron is oxidized into ferric iron due to aeration. Then, in arid seasons, when the influx of surface water is stopped and the pH of the lake water gradually rises to neutral, ferric hydroxide is precipitated very rapidly, because its isoelectric point is pH7. A little iron still remains in solution. Due to the seepage of underground water and evaporation, the reaction becomes alkaline. Then silica is supplied to the basin in colloidal solution, gradually increasing its concentration as the evaporation goes on. Under an increased alkaline condition, colloidal silica and iron hydroxides combine with each other to form iron silicates (greenalite, chamosite, etc.) and the rise of temperature accelerates the precipitation of carbonates of iron and other compounds. Again, upon the return of a wet season and influx of fresh water, the reaction returns through neutral back to acid again.

Sakamoto concluded that an environment suitable for this cyclical deposition occurred only once — in the Middle Precambrian — during the earth's history and is unlikely to occur again.

According to Percival (1954):

> N. Svitalski (1937), writing on the iron deposits of the Krivoi Rog district in the Ukraine, expressed the opinion that the iron and silica were derived from Archean granites, gneisses and greenstones by weathering analogous to laterite weathering, leading to the concentration of iron on the continental surface. Later, swamp conditions favoured colloidal solution, and streams transported the colloids to a marine or large lake basin where they were precipitated as suggested by E. S. Moore and J. E. Maynard. Reactions between the ferrous oxide and silica gave iron silicates. Small quantities of other components present gave rise to the chlorite schists and later some of the ore was precipitated as carbonates.

The studies of Precambrian banded iron ores by Alexandrov (1955) are particularly interesting: "It is suggested that the intermittent banding of silica and hematite in Precambrian banded ores is caused principally by the selective weathering of Precambrian soil. Due to the seasonal changes of temperature, amount of precipitation, and the alternately higher and lower pH range of the leaching solution, the Precambrian soil yielded alternately solutions carrying to the basin of deposition almost exclusively silica during the warm season and chiefly iron oxide during the cool period of the year." His discussion centred around the behaviour and possible role of humic acids as selective leaching agents. Higher temperatures, above 20°C, cause the breakdown of humic acids, which is accelerated by the activity of bacteria that cannot exist in a very acid environment, and the humus content in the soil is lowered. The pH of the soil rises, favouring increased leaching of silica and general laterization conditions prevail. When the temperature of the soil drops below 20°C in the cool season humic acids develop, humus content of the soil increases, and leaching of iron oxides by humic acids takes place with development of podzolization rather than laterization. The experimental evidence in support of this theory is most impressive and he postulated deposition of these elements in a freshwater basin where they can be widely distributed prior to precipitation. The salts of sea water would cause immediate precipitation of the colloids.

Hough (1958) described a possible freshwater lake environment that would satisfy the required chemical conditions for precipitation of iron and silica with banding ascribed to seasonal circulation of the lake water.

Ofterdahl (1958) discussed a theory of exhalative-sedimentary ores and has contributed considerable evidence and argument for the contribution of iron from volcanic emanations and gases that are discharged and precipitated in a marine environment. The association of acid pyroclastic rocks is emphasized and present-day fumarolic activity is discussed. Hegemann (1961) attributed the extrusive-sedimentary ore deposits of the Eastern Alps — banded chert and iron oxides in volcanic assemblages — to the same mode of origin.

Most concepts of the origin of iron-formations, although disagreeing as to the processes of transportation and precipitation of iron and silica, agree that the source of these elements was either a deeply weathered landmass, or volcanic emanations, or the reaction of sea water on hot extrusive rocks. In considering the origin of seams of iron ore below the North German Plain, Borchert (1960) proposed the idea that these elements were dissolved from deep ocean sediments under special conditions as follows:

> Since trivalent iron is practically insoluble in the presence of oxygen, it is contended that the mobilization of iron by weathering solutions in continental regions could not have contributed significantly to the formation of marine iron ore deposits in pre-Devonian times. The mobilization and precipitation of iron must then have been effected within the oceans themselves.
>
> Emphasis is placed on the existence of a CO_2 zone at moderate depths in the ocean wherein abundant iron is dissolved, later to be precipitated mainly in oxygenated shallow waters as limonite oöliths or clay ironstones, and partly in deep waters of the H_2O zone as chamosite.
>
> Distribution of the different facies of limonite, siderite, chamosite, and pyrite tends to parallel the coast line and is controlled by the direction of flow of sea currents. Changes in these currents, each with its own characteristic physico-chemical traits, account for the principal vertical and lateral variations in the iron-ore facies, small-scale variations being due to climatic changes.

Many of the authors mentioned here discuss problems of origin mainly of the cherty iron-formations so prevalent in the Precambrian. These discussions also apply in general to all iron-formations although the mechanism of precipitation may differ for the various types. The origin of the Minette type of formation was considered in papers by Hallimond *et al.* (1951), Dunham (1960), Hallimond (1925), Whitehead (1952), Cayeux (1922), Borchert (1960), and Bubenicek (1961).

This brief résumé of the writings of some prominent geologists on the origin of iron-formations indicates some of the concepts held on this fascinating subject. The excellent data that have been marshalled and the arguments presented cannot be summarized here in any detail and, in fairness to the authors, the reader is referred to their original papers. Only there can he gain an appreciation of the effort that has been made to understand the origin of iron-formations.

Sources of Iron

Solubility of Iron in Natural Waters

According to Hem and Cropper (1959), most natural waters exposed to the atmosphere have a pH within the range of 5.0 and 8.0 and an Eh between 0.3 and 0.5. Their work indicates that 100 ppm of ferrous iron could be present in a solution with an Eh of 0.3 and pH of 5.0. Insoluble ferric iron will however form in about two thirds of the field defined by this range of Eh and pH and that part of the field will have very little iron in solution. Only in waters with properties that fall within the range of the remaining third of the field will appreciable amounts of iron be found, and it will be in the ferrous state. They noted that water below the land surface and not in contact with the air has Eh values of 0.20 volts or less and a relatively low pH (less than 6) if enough carbon dioxide is present. Under these conditions ferrous solutions with up to 50 ppm iron may be stable. On exposure to air the ferrous iron in most solutions is oxidized to ferric, and in a pH range of 5 to 8 the total amount of ferric iron that can remain in solution in the form of Fe^{3+}, $FeOH^{2+}$, and $Fe(OH)_2^+$ ions is generally below 0.01 ppm. Iron detected in excess of this amount in solutions exposed to the air must be either in undissolved particles or colloids, or in complex ions.

Investigations by Oborn and Hem (1961) of microbiologic factors in the solution and transportation of iron show that distilled water may dissolve from 1 to 8 ppm of iron from mixtures of soil, sand, and organic matter after a period of incubation at 72° to 100°F. The solution of this iron is related to microbiologic activity, and organic complexes of iron aid in retaining the iron in solution made available by the microbiota. They note that aquatic vegetation is particularly important in adding iron to surface water or in removing dissolved or suspended iron from water. The iron from vegetable debris is likely to pass into groundwater, and iron combined with organic molecules can form relatively stable and soluble complexes. The effects of organic complexes in the solution and transportation of iron are apparently much more significant than those caused by changes in pH and Eh of the solutions. A very significant role in the cycle of iron in nature is played by anaerobic bacteria which reduce ferric iron in the absence of oxygen and return it to solution.

Moore and Maynard (1929) found that waters from a number of lakes and rivers in the Canadian Shield contain from 1.3 to 14.4 ppm ferric oxide and that the quantity of iron carried was dependent on the amount of organic matter in solution. After carbonated water, peat solutions were found to be the most effective solvent of iron and silica from basic rocks. They concluded that "Iron is not carried as bicarbonate in cold natural surface waters high in organic matter, but in all probability it is transported as a ferric oxide hydrosol stabilized by organic colloids. A relatively small amount may be carried as salts of organic acids adsorbed by organic colloids." According to their work, as much as 36 ppm ferric oxide, formed by the oxidation and hydrolysis of ferric bicarbonate, can be held in colloidal solution by 16 ppm organic matter.

The amount of iron carried in solution in streams in North America is usually less than one part per million; in tropical areas it is usually higher, ranging in many rivers between 2 and 7 ppm. This higher content is attributed mainly to the presence of organic acids that facilitate the solution of the iron from the source rocks and hold it in solution during transportation as organic compounds or by stabilizing ferric hydrosols. The average content of iron dissolved in sea water ranges from 0.002 to 0.02 ppm; the iron content, including particulate iron, may be slightly higher in some near-shore areas. As there is little evidence to indicate an appreciable increase in the iron content of sea water near the mouths of large rivers, the iron in the river waters must be precipitated almost immediately on reaching the sea. Apparently it is disseminated with the other sediment carried by the streams and dispersed in the muds with no appreciable concentration in the offshore basins. Even sediments in the delta and offshore areas from rivers, such as the Amazon, that carry a relatively large amount of iron do not have an unusually high iron content.

Derivation of Iron from Rocks by Weathering

Climatic conditions determine whether podzolization or laterization takes place during the weathering of a landmass. In areas with temperate to cool climate, appreciable rainfall, and accumulation of humus from decaying vegetation, the iron is generally leached during weathering and removed in the drainage, but silica leaching is relatively limited. Soils rich in silica but deficient in iron therefore develop under these conditions (podzolization). In contrast to this, in warm climates where there is a low humus content in the soil and high temperatures cause a rapid breakdown of humic acids, silica leaches rapidly but iron remains in the residual soils (laterization). Alexandrov (1955) drew special attention to the seasonal fluctuations in temperature that could alternately bring large quantities of iron and silica into solution during weathering and deposit them in thin banded iron-formations.

The complex factors affecting the solution and transportation of iron indicate that very special conditions must prevail in order to have large quantities of iron transported from a landmass and precipitated with only limited amounts of elements other than silica. First, chemical weathering must predominate over mechanical disintegration of the rocks in order to separate the iron from the elements combined with it in the common rock minerals. Isolation of the iron must involve either selective weathering and solution or a special environment where only iron and silica are precipitated. If the iron in these formations was derived from a landmass, best conditions for its removal would be expected on a landmass with low relief where chemical weathering predominated and mechanical weathering was at a minimum.

A temperate to warm climate with abundant rainfall would support a rich flora and develop humic acids and organic compounds to facilitate solution of the iron. Poor drainage which might be expected with low relief would favour the formation of swamps where aeration of the water would be minimized and the lack of oxygen would permit the accumulation of peat, humus, and organic compounds. Under these conditions of stagnant water, aquatic plants could flourish near and at the surface of the water but with a limited supply of oxygen in the deeper parts of the swamp the vegetation would accumulate to form peat or humus and supply an abundance of oganic compounds in solution. The pH of the swamp water would be lowered, strong reducing conditions would develop, and conditions ideal for the leaching of ferrous iron, reduction of ferric iron to the ferrous state, and the best environment for solution of iron by both organic and inorganic processes would prevail. Iron in solution in waters flowing into these swampy areas would probably be concentrated by aquatic plants and later be released in organic compounds when these plants decomposed. Iron carried out of these swampy basins in streams of low gradient would probably be in the form of organic compounds or of ferric colloids protected by organic constituents. Precipitation of this iron would take place when the river water became interspersed in the sea water and the protecting organic constituents decomposed. In shallow water areas where oxygen is mixed in the sea water by agitation much of the iron would be oxidized and precipitated as ferric oxides.

An explanation for the lack of appreciable clastic sediment in these formations may be either that iron was being eroded from a land surface of low relief and therefore little clastic material was being carried by the streams, or that it was deposited in a shallow basin formed by an offshore bar or some type of clastic trap. The iron in solution would be carried farther seaward before being precipitated.

Precipitation of the iron without extensive precipitation of calcium and magnesium carbonate is a more difficult problem. According to Krumbein and Garrels (1952), these carbonates precipitate when the pH of the solutions is above 7.8. As pH of sea water where circulation is unrestricted is about 8, the precipitation of iron (if it is to take place without calcium and magnesium carbonate) must take place where the pH is maintained below 7.8, that is before the solutions reach the open sea, and it must be in some type of depositional basin not open to the sea. River waters with a lower pH than sea water emptying into these basins could probably keep the pH too low to allow calcium and magnesium carbonate to deposit. James (1951) suggested that the Lake Superior iron-formations were deposited during a phase of development of the Huronian geosyncline when suitably restricted basins were formed by offshore swells or island arcs. A situation similar to this (to be described by the writer in a later volume) may have existed in the Labrador geosyncline.

Noting the association of volcanic rocks with many iron-formations of the world, Guild (1953) suggested that volcanic emanations aided in lowering the pH of sedimentary basins to the point where carbonate precipitation was inhibited. He believed however that at least part of the iron and silica was contributed by volcanic sources.

Organic compounds must play an essential role if iron is to be derived from a landmass, and it is doubtful whether sufficient iron could be leached and separated from other elements during weathering by inorganic processes alone. An interesting possibility for an abundant source of iron might be found in an area where a deep lateritic or residual accumulation of iron became flooded and covered by swamps. The biggest difficulty in accepting this simple hypothesis for the origin of iron is that iron-formation should be forming today where such conditions prevail. Nor can the problem of the deposition of iron be divorced from that of the simultaneous deposition of silica.

Derivation of Iron from Volcanic Sources

Except for Borchert's belief that iron is dissolved from sediment in the deep ocean basins, the two main schools of thought are that iron is derived by the weathering of rocks and from volcanic emanations. Many believe that the delicate balance of physical-chemical factors and very special conditions necessary for iron-formations to be derived from a landmass appear extraordinary and are too exceptional to be repeated in widely separated areas. It seems to such that the intimate and persistent relationship of the Algoma-type formations with volcanic rocks and the association of contemporaneous volcanism with practically all Superior-type formations cannot be fortuitous but indicate a genetic relationship. The sediments associated with many iron-formations are either typical continental shelf or eugeosynclinal assemblages and the interlayered iron beds must have come from a separate distinctive source, probably volcanic. Volcanism can also provide an adequate source of silica to account for the chert that is so abundant a part of these beds. The divergent chemical properties of iron and silica must require a special source for either the iron or the silica to account for their concentration in the same formation. Evidence of deeply weathered land areas that could have supplied the iron and silica for the iron-formation has not been observed, although there are several reasons why it may not have been preserved.

A volcanic source for iron and silica for the Gunflint Formation is indicated by Goodwin (1956) who states that "a direct and sympathetic relationship between volcanism and iron-formation is indicated by the presence of pyroclastics and lava along certain horizons in the iron-bearing rocks, and in a larger and more significant sense, the cyclical coordination of volcanism and sedimentation." He cited direct evidence for the contribution of iron from a volcanic source in the case of volcano Santorin in the Aegean Sea where periodic outbursts between 1650 and the present are listed by Behrend (1936). Periodic outbursts of ash and tuff alternated with relatively long periods of rest characterized by sustained discharge of ferrous carbonate-bearing gases and solutions, and observers report that submarine discharges of ferrous carbonate solutions quickly oxidize to red ferric iron precipitate upon approaching surface waters. Hjelmqvist (1951) gave further evidence of direct contribution of iron from the active volcano Fossa di Vulcano, one of the Eolian Islands, where pyrite and marcasite are deposited in vesicles and pores of lava and replace obsidian, and limonite occurs farther from the fumaroles. In one place submarine emanations heat the Mediterranean to a temperature of 97°C.

A fascinating example of direct contribution of iron to the sea of Okhotsk from thermal springs at the volcano Ebeko was given by Zelenov (1958) and part of his description as translated stated

> The sources of the Iur'eva River lie in two deep fissures which cut ancient lavas about 400-500 m below the crest of the volcano. Two cold acid springs and three groups of hot springs with water temperature of 46°C, pH 1.4, Fe'' content of 150 mg/l, Fe''' content of 90 mg/l and a total discharge of more than 40 1/sec contribute to the stream of the right tributary fissure, which heads in a small snow field at an elevation of 1019 m. A series of hot springs with an equally high discharge, contributing to the left tributary stream is located in a gully in a very steep slope of Zelenaia Mountain, whose top is crowned by a steaming fumarole. The uppermost spring is a tub-like basin which is kept filled with boiling water flowing down a steep wall of bedrock through fissures coated with bright yellow sulfur. The pH of this water is less than 1.2; the content of Fe'', 350 mg/l; and of Fe''', 8 mg/l. The old lavas here are porphyritic andesites and in the channels of the streams they are leached bone-white. This is due to the fact that their minerals have been completely decomposed and the groundmass opalized. Where the two streams join, the pH of water is 1.5, the content of Fe'' is 50 mg/l and of Fe''' 150 mg/l, and the discharge is measurable in hundreds of liters per second. Approximately 500 m lower, a flow of recently erupted basalitic andesite lava 80-100 m wide enters the canyon. At the end of this flow there is another group of hot (over 90°C) acid springs with a fairly large total discharge, which have their source in the leached rock. The pH of these waters is not less than 1.2, the content of Fe'' about 800 mg/l and of Fe''', 8 mg/l.
>
> Such are the sources of the acid Iur'eva River, a greenish clear mountain stream which rushes tempestuously into the Sea of Okhotsk. On cold days clouds of white vapor rise above the river throughout its three-kilometer course and the valley is filled with suffocating, ill-smelling gas.
>
> Downstream, ferrous iron changes rapidly into ferric, and the total content of iron in the stream slightly increases. In a sample taken below the confluence of all thermal waters, the pH is 1.3, the content of Fe'' is 150 mg/l and the Fe''', 120 mg/l. In a sample taken 700-800 m lower, below the junction with a small brook, the pH is 1.4, the content of Fe'' is 20 mg/l and of Fe''', 200 mg/l. In a sample taken one kilometer lower, just above a series of picturesque waterfalls, the pH is 1.6, the content of Fe'' is 10 mg/l and of Fe''', 320 mg/l. At the mouth of the Iur'eva River, one sample had the pH 1.4, the content of Fe'' was 5 mg/l and of Fe''', 320 mg/l; a second sample taken in the same place after a continuous two-day rain had the pH 1.5 and the Fe''' content of 220 mg/l.
>
> Considering the total volume of water in the Iur'eva River, whose discharge at the mouth is 1.8 m^3 sec., the amount of iron carried in solution is impressive and reached 400-500 g/sec or about 35-50 tons per day.

Zelenov described the yellow-coloured off-shore patch of sea water where the ferric iron salts are coagulated as the river waters mix with the sea, and showed that the volume of water in the patch represents the exact amount of sea water required to change the pH of the river water from 3.2 to that of normal sea water, about 7.6. In a later paper (1959) he showed that the content of alumina in these waters is more than twice that of iron but the iron and silica are separated from the alumina and deposited near shore whereas the alumina remains in suspension and is carried farther out to sea before being deposited. Limonite deposits, 10 to 12 metres thick, are in the process of being laid down in one of the crater lakes fed by cold acid springs that are remarkably pure with insignificant amounts of Mn, V, and Ni. There are a number of streams where a similar phenomenon takes place and it can readily be appreciated that if these streams contributed only 50 tons of iron per day a billion tons of iron could be supplied to a sedimentary basin in 50 or 60 thousand years. If several volcanoes contributed similar amounts of iron to thermal springs along a volcanic arc, sufficient iron to form an iron range could be supplied in a much shorter time. Estimates using this rate of deposition indicate that about three streams per mile along a volcanic arc would be necessary to supply sufficient iron to produce the Labrador geosyncline iron-formations, if they were to be deposited in 50 thousand years.

It may be highly significant in this connection that the basaltic rocks in the Labrador geosyncline, according to Baragar's study (1960), are characterized by extreme iron enrichment and low potassium content. The ratio of FeO/Fe_2O_3 for the series is unusually high and differentiation of basaltic sills took place in situ from water-deficient magma. Alternative end products of gabbro pegmatite and albite-gabbro pegmatite probably resulted from slight differences in volatile contents. Under such circumstances large amounts of iron could have been supplied in emanations from these sills during their differentiation.

Contribution of iron from a volcanic source on such a scale is conceivable, but volcanic emanations highly charged with iron are reported so infrequently from the long belts of active volcanism in the world today that it is still doubtful if this is the main source. Our knowledge of the nature and extent of volcanism in Precambrian time is still insufficient for a realistic appraisal of this possible source of iron.

Influence of the Atmosphere

Many believe that the answer to the problem of an adequate source of iron is to be found in the fact that less oxygen was present in the Precambrian atmosphere and that ferrous iron consequently accumulated in the seas in a manner analogous to the same way that calcium and magnesium are concentrated today. In atmosphere with a higher proportion of CO_2 and less oxygen the pH of the sea may have been lower and if so much ferrous iron would be held in solution. Then as plant life, even primitive plant life, developed oxygen was gradually released to the point where the accumulated ferrous iron was oxidized and precipitated over a relatively short period to produce the widely distributed iron-formations. Indeed a gradual change in ratio of CO_2 to oxygen in the atmosphere may have taken place throughout geological time.

It is doubtful, however, if this change had much influence on the origin of iron-formations. Some of the earliest known iron-formations have jasper beds with abundant hematite dispersed in the chert interlayered with iron carbonate and silicate facies, and these have all the characteristics of primary beds. Iron is a very delicate indicator of oxidizing conditions and even though there may have been less oxygen present there certainly was enough to oxidize very large quantities of iron in these early formed beds.

Sources of Silica and Origin of Chert

As already emphasized, the origin of iron in iron-formations cannot be considered without also considering the origin of the chert beds so universal in all iron-formations intimately associated with volcanic rocks. Earlier studies on the geochemistry of silica in sedimentary environments and the results of recent experimental work on silica have been reviewed by Krauskopf (1959), Bien, *and others* (1959). The main properties and habits of silica are summarized below.

Solubility of Silica in Natural Waters

As much as 100 to 140 ppm amorphous silica dissolves in either fresh or marine water at ordinary temperatures. In true solution it is in the form of monosilicic acid, H_4SiO_4, and not in the form of a colloidal dispersion. Crystalline forms of silica, such as crystobalite, chalcedony, and quartz, have much lower solubilities, that of quartz being the range of 6 to 14 ppm at ordinary temperatures. Opal however may dissolve in about the same proportions as amorphous silica. The solubility of amorphous silica increases directly with increasing temperature and at 0°C is in the range of 50 to 80 ppm, at 25°C in the range of 100 to 140 ppm, and at 100°C between 360 and 420 ppm. When solutions saturated with silica are cooled, or evaporated, or when the pH is reduced from 9, the excess silica forms colloidal particles that remain in suspension unless flocculation is induced by some means. Silica behaves in this respect like many other saturated solutions of fairly insoluble compounds and an equilibrium is set up between monomeric acid and silica in the form of a solid, gel, or colloidal particles. An unusual characteristic of silica solutions is the slowness of both the dissolution and the polymerization reactions; supersaturated solutions of silica can exist for weeks or months before the excess silica coagulates.

According to recent work the solubility of silica is practically independent of pH for values below 9 but for values above this increases abruptly to nearly 5,000 ppm when the pH is 11. Of particular significance is the fact that the rate of polymerization of silica coming out of a solution increases decidedly when the solutions are changed from weakly alkaline to weakly acid and by an increase in temperature. Also the greater the initial degree of supersaturation of silica the faster does the concentration of dissolved silica approach the equilibrium value.

In the experimental work the solubility of silica in sea water was not appreciably different from that in freshwater, but the rate at which solubility equilibrium was attained was greater. Cations capable of forming very insoluble silicates generally reduce the solubility of amorphous silica. Okamoto, *et al.* (1957) showed that even small amounts of Al^{+3} in a pH range of 8 to 9 reduce the concentration of silica appreciably. The addition of aluminum sulphate to supply 1 mg/l Al reduced the concentration of colloidal silica from 45 ppm to about 5 ppm between pH values of 4 and 5 but did not have a marked effect at other pH values. Concentrations of 20 mg/l Al had less effect on molecularly dispersed silica reducing its concentration from 35 to 15 ppm between pH values of 5 and 10 and greater concentrations of Al reduced the silica in solution even more. Where the ratio of silica to aluminum is high, 92 mg/l SiO_2 to 1 mg/l Al, very little colloidal silica is precipitated. Other solids such as suspended calcite, ferric hydroxide, kaolinite, or montmorillonite added to silica solutions at equilibrium affected the solubility of silica very little. The addition of ferric oxide, however, causes rapid coagulation of excess silica in suspension according to Krauskopf (1959).

Diatoms have the ability to reduce the concentration of silica in sea water from over 100 ppm to less than 1 ppm, and their influence in nature is evident in the formation of large deposits of diatomaceous oozes and cherts. Algae in hot spring waters, according to White (1956), apparently have only minor effects in causing polymerization and precipitation of silica.

Most river water and groundwater contain less than 35 ppm SiO_2 and connate waters usually between 20 and 60 ppm. Most of this is in the form of monomeric silicic acid. Data on the nature of silica in natural waters that indicate whether the silica is in true solution or is colloidal are very limited. Only low concentrations of silica are found in sea water: 0.1 to 4 ppm in surface water and 5 to 10 ppm in deep water, according to Krauskopf (1959). As the solubility of silica in sea water is very like that in freshwater, according to experimental evidence, there must be particular causes for the reduced silica content of the sea. As much of the silica is carried in solution in streams, the idea of electrolytes in sea water causing coagulation and deposition of colloidal silica is not an adequate reason unless other factors affect the equilibrium of the silica in solution.

Bien, *and others* (1959) have studied the removal of soluble silica from freshwater in the Mississippi River delta and present considerable data to show "that removal of all or most of the dissolved silica in river water occurs when this water mixes with sea water. Biological uptake can account for an appreciable amount of this removal during the final mixing of the two water masses, but probably not initially. Silica is also precipitated inorganically by coprecipitation and adsorption in which suspended solids in the river water and electrolytes in the sea water are required."

Summary of Existing Ideas

One idea regarding the origin of chert is that it occurs as masses replacing other sediments either during diagenesis or during some stage that post-dates lithification of the beds. The chert is believed to be derived either from mineral constituents in the beds or from silica precipitated and dispersed with the clastic or carbonate beds. Another concept is that the silica is extracted from the water by organisms such as silica secreting sponges, radiolara (protozoa), or diatoms (planktonic plants). In the absence of other clastic or chemical sediments, the organic remains accumulate to form siliceous oozes which consolidate to chert during diagenetic processes with destruction of many of the spicule or radiolaran forms during lithification. Many workers feel that these processes do not give an adequate explanation for thick sections of banded chert, which are commonly associated with volcanic rocks, and they conclude that the silica is supplied directly from volcanic emanations or dissolved from hot lavas, tuffs, and ignimbrites by sea water and then is precipitated as a gelatinous sediment from solutions saturated with silica during active volcanism. Where deposition of clastic or volcanic fragmental material is inhibited, silica gels may accumulate to form banded chert beds. Still others have concluded that chert forms by redeposition of silica in the zone of weathering. Well-documented and convincing examples of chert formed by any one of these processes are known and evidently they are all valid concepts and may apply to certain specific chert occurrences. Others conclude that silica in thick successions of banded cherts may be derived in part from the chemical weathering of a landmass and in part by silica supplied from volcanic emanations or dissolved from hot lavas or pyroclastics. According to experimental results, sea water is undersaturated with respect to silica and could conceivably dissolve silica from bottom sediments. This could then be precipitated in more acidic waters near the mouths of rivers or where the pH was decreased by the addition of volcanic emanations. Correns (1950) and Krauskopf (1959) suggested that since calcite solubility increases appreciably when the pH decreases from 8.5 or 9 and the solubility of silica decreases rapidly below this pH range, that a small decrease of pH in sea water could lead to simultaneous dissolution of calcite and precipitation of silica.

Silica Derived by Weathering of Rocks

It can be appreciated that, given sufficient time, streams carry enough silica to form bedded chert deposits, if clastic sedimentation and chemical precipitation of other constituents were excluded. Suitable circumstances for this would seem to be intermittent rainfall in a warm climate with dry intervals when the pH of groundwaters would rise and silica would be taken into solution. With the flushing of the soil during periods of heavy rain this silica would be removed in solution. No doubt with the influx of water and lowering of the pH some silica would come out of solution but be carried as dispersed colloids and precipitated rapidly in the presence of electrolytes on reaching the sea, especially where ferric oxide or other solids and organisms were present. Where laterization takes place on a landmass, silica is readily leached and is precipitated on reaching the sea. Possible conditions for a near-shore environment where clastic deposition is negligible were outlined in the discussion of the chemistry and deposition of iron. An example of jell-like masses of pure to impure silica forming in recent miogeosynclinal sediments off the Gulf

Coast in the Mississippi River delta are cited by Bissell (1959) and Russell (1936). Gobs, blebs, and irregular masses of siliceous material are forming in shallow depressions in brackish and marine waters on the delta. These probably represent the nodules, thin layers, and lenses of chert in limestones and argillaceous beds in many miogeosynclinal sediments. Thick accumulations of recently deposited silica, comparable in extent to thick-banded chert formations found in many rock successions, have not been observed in contemporary sedimentary basins.

Silica Derived from Volcanic Sources

Many who have studied the origin of thick formations of banded cherts, including Goldstein (1959) who examined the Palaeozoic cherts and novaculites of the Ouachito Mountains and Bissell (1959) who examined the chert formations in the upper Palaeozoic of the Cordilleran area, have concluded that the silica was derived mainly from volcanic sources. An abundance of silica can be taken into solution: by submarine leaching and weathering of large masses of volcanic materials; by solution from submarine outpouring of lavas and from tuff and ignimbrites deposited in the sea; and directly from volcanic emanations, hot springs, and thermal water expelled from near-surface magmas. The increased temperatures of sea water caused by submarine extrusions will greatly increase the amount of silica that can be dissolved. Sea water may readily dissolve silica from the finely divided siliceous fragments in tuff or ignimbrite masses and so become saturated. Large amounts of silica may then readily precipitate where the temperature is lowered, where the pH of the water decreases near shore or by flocculation of suspended colloidal silica in the presence of electrolytes or solid matter such as ferric oxide, alumina or other solid particles. Thermal waters emerging from bore-holes in the Wairokei area of New Zealand contain between 200 and 700 ppm of metasilicic acid (H_2SiO_3) and the great amount of silica discharged into the sea during the active period of a volcano can be appreciated. Travertine deposits, including the 500-foot deposit near Huancavelica mine in Peru mentioned by White (1955) that has been precipitated from hot springs, are good evidence for volcanic thermal waters supplying large amounts of silica. The direct and intimate relationship of most laminated chert formations with volcanic rocks, especially tuffs and rhyolitic extrusive rocks in a geosynclinal basin, in many parts of the world, is further evidence of a volcanic origin for the silica.

There is abundant evidence found in the thick-banded chert masses to indicate that the silica was precipitated as a gel which later hardened and crystallized to form chert. Syneresis or shrinkage cracks, septarian structures, lenticular cracks or desiccation cracks that have been distorted during compaction all indicate that the silica was deposited as a gel and later dehydrated and compacted. Many slump structures, folds and slips confined to single beds or to a restricted group of beds indicate penecontemporaneous deformation of very soft and incompetent beds. Structures identical to those visible in chert formations are found in cherty iron-formations.

Many of the recent thick rhythmic banded and laminated chert formations have zones that contain from 5 to 25 per cent iron in the form of hematite, magnetite, or siderite in iron-rich laminae. These are lithologically similar to the Algoma-type iron-formations so commonly associated with early Precambrian volcanic rocks and there can be no doubt that they formed in a similar way. Large zones of chert or ferruginous chert with less than 15 per cent iron are commonly associated with many early Precambrian iron-formations so that the occurrences of bedded cherts in younger rocks seem to differ only in having a smaller proportion of iron. Many of the younger formations have large zones of red jasper or green chert that are richly coloured but contain only small amounts of iron. Ferruginous zones are present in the Franciscan (Jurassic?) cherts in California, in the Palaeozoic ferruginous cherts of Notre Dame Bay, Newfoundland, in the cherts associated with tuffs and slate in the Sicker Group (Jurassic) on Vancouver Island, and in many other chert formations associated with volcanic rocks.

Chert Formed by Organisms

Examples of diatomaceous and radiolarian oozes forming on deep sea floors by the accumulation of silica-secreting organisms are well known. These deposits, which are readily recognized, are considered to be a separate type of siliceous sediment; they were not formed in the same way as the thick laminated chert formations that have so many features in common with the major types of siliceous iron-formation. Many of them contain numerous well-preserved sponge spicules or other recognizable organic forms enclosed in finely crystalline chert. The question arises as to whether these silica-collecting organisms are present because they thrive in a silica-rich environment, or whether they were responsible for collecting the silica to form the chert bed. Many who have studied the origin of these chert beds in detail believe that the former is so and that the silica-rich environment was created by inorganic processes.

Environmental Changes During World History

The environment at the surface of the earth may be quite different today from what it was during much of the past. Some of the factors that may have caused major changes are mentioned below without attempt to elaborate on their effect on the origin of iron-formations.

As has been pointed out, certain types of iron-formation are more abundant with rocks of some ages than with others but none is restricted to one geological period. Algoma-type formations are numerous in early Precambrian rocks but occurrences are known in rocks of nearly all ages. Superior-type beds are most common in the Proterozoic, and Clinton-type in the Palaeozoic, but these types are interbedded in some areas (O'Rourke, 1961; Blondel, 1955; Edwards, 1958), indicating fluctuation between the two types of sedimentation and a close relationship in their mode of origin. A gradual change through geological time in the depositional environment for these chemical sediments has been suggested. Thus the lithological differences between Superior, Clinton, and Minette types may reflect progressive changes in the nature of the sedimentation probably caused by evolving tectonic patterns on the continents or in geosynclinal belts.

A gradual increase in the amount of free oxygen in the atmosphere may have taken place with the passing of time due to its release by photosynthesis as plant life became increasingly abundant. An increase in the abundance of atmospheric oxygen would in turn affect the solubility of iron and its whole geochemical cycle. It is suggested that volcanic extrusions and surface rocks were more basic and richer in iron during early periods of the earth's history when there was less differentiation of salic and mafic materials in the earth's crust than at present. Differentiation of near-surface magmas may have produced iron-rich emanations that were emitted in large quantities by most of the active volcanoes. Furthermore, a decrease in frequency and amount of volcanic activity would be expected if the earth cooled from a molten mass. Orogenic disturbances may have been less frequent in early geological periods than later and therefore weathering, erosion, and depositional cycles continued uninterrupted for longer periods between orogenies. The fundamental causes of glaciation are not known but the frequency and intensity of cold periods in the earth's climate could have a marked effect on the chemistry of rock-weathering and the accumulation of sedimentary iron. Accumulative effects of radiation originating within the earth, or from the sun or outer space are not well enough known to predict their effects in major geological processes.

Probably one of the most significant factors relating to the origin of iron-formations has been the gradual increase in the amount of organic life on the earth and the effects this has had in geochemical processes.

Depositional Environments for Iron-formations

Cherty Iron-formations

It is important to distinguish between the two major types of cherty iron-formation, Algoma and Superior (*see* p. 91), as they appear to differ considerably in their manner of occurrence and origin.

Algoma Type

The persistent occurrence of Algoma-type iron-formations and ferruginous chert with volcanic rocks, greywacke, and slate in eugeosynclinal rock assemblages in most geological periods indicates a genetic relationship between these rocks and volcanism. Both the iron and silica contributed to these formations appear to be the products of volcanic activity.

In some regions, as Goodwin (1962) has demonstrated in the Michipicoten area, particular facies of the iron-formation vary with the type of associated rock, which in turn varies according to proximity of the rocks to the source of volcanism. He showed that in the Michipicoten area:

> Banded iron formations contain two mutually gradational facies associated respectively with sedimentary and volcanic rocks. Iron-formation of sedimentary association, composed of alternating layers of chert, siliceous magnetite, and jasper occur in lensy, discontinuous bands. Iron-formations of volcanic association, on the other hand, typically comprises banded chert, pyrite, and carbonate members in descending stratigraphic succession; this type of iron-formation generally overlies felsic volcanic rocks and underlies intermediate to mafic volcanic rocks.
>
> The relative stratigraphic positions of siderite below and banded chert above in iron-formations of volcanic association appears to represent the chemical response of large quantities of siliceous bicarbonate solutions to sudden entry into the marine environment; under these conditions escape of CO_2 favored immediate precipitation of siderite, whereas the more soluble silica remained longer in solution. The present stratigraphic arrangement in the iron formation may indicate, in addition, minor internal diagenetic and/or post-consolidation migration of the iron and silica constituents.

From this concept one would expect to find oxide facies present in predominately sedimentary rock groups where there has been sufficient time for ferrous carbonate to oxidize as it became mixed with oxygenated waters during migration from the source area. The consistent association of oxide facies with sediments would support this concept. Only limited work has been done in regard to the spatial relationship of the different facies of iron-formation with respect to the source and kind of volcanism. Observations by the writer in a number of areas in northern Ontario indicate that the relationship described by Goodwin at Michipicoten is present in many other iron-formation and volcanic rock belts, including the Kaministikwia iron range, the Cunningham-Garnet township range, the Kukatush range, and others.

Iron-formations in a number of areas occur between rhyolitic to dacitic flows or pyroclastic rocks below and andesitic to basic flows above. The occurrence of this rock sequence is frequent enough to suggest that iron salts are differentiated and released in abundance during certain periods in the history of volcanism in an area. If iron is enriched in rhyolitic and dacitic magma formed by differentiation in the magma reservoir that supplied the volcanoes, iron-rich exhalations would be expected to follow or coincide with the explosive release of rhyolitic extrusions.

Volcanic rocks of the andesite-rhyolite suite in the Michipicoten area, according to Goodwin (1962), are enriched somewhat in siderite constituents (Fe, Mn, CO_2) and depleted in CaO, alkalies, and minor SiO_2 in comparison to a typical andesite-rhyolite suite found in the Cascade province of California. Direct contributions of iron and silica can be expected from magma that follows this trend of differentiation. The kind of facies of iron-formation that is precipitated will be controlled mainly by the Eh and pH conditions of the water in the basin.

The dacite-rhyolite rocks below the iron-formation at Michipicoten are highly carbonatized to depths of 100 to 500 feet. Goodwin found that rocks are moderately altered to depths of 200 feet and mildly altered to depths of several thousand feet. The alteration consisted of additions of iron and manganese carbonates and removal of silica, lime, and alkalies. This alteration is independent of volcanic rock types and structural features, and is attributed to the action of hot springs or fumaroles that affected the flat-lying volcanic beds during the development of the overlying iron-formations. Estimated quantity of silica leached is about the same as that precipitated in the overlying iron-formations.

The relationship of the various facies of iron-formation to types of volcanic rocks and the position of the iron-formation in the succession of volcanic and associated sedimentary rocks need to be studied in many other areas. Nonetheless, evidence is accumulating from this type of study that indicates a volcanic source for the iron and silica in this type of iron-formation.

Derivation of iron and silica from deep weathering of a landmass is considered unlikely for Algoma-type iron-formation. The associated sedimentary rocks are greywackes, fine-grained slates, and other clastic rocks derived from the mechanical disintegration of volcanic rocks and are typical eugeosynclinal assemblages. Sediments such as orthoquartzites, limestones, dolomites, alumina-rich mudstones, and beds rich in those well-rounded heavy detrital minerals typically derived from mature weathered areas are generally absent. The thin layers of sedimentary rock interbedded with chert and iron-formation have a normal amount of iron, except for those that show evidence of post-depositional alteration as noted above, and are similar to other sediments in the rock sequence that lie some distance from the iron-formation. If an abundance of iron was being supplied to the sea from a land source, a gradual increase would be expected in the amount of iron oxide and carbonate in the sedimentary beds as the iron-formation is approached. As only clastic beds containing a normal amount of iron distributed in the common rock-forming silicate minerals are interlayered with iron-formation and chert, a more direct and possibly intermittent source of iron is suggested. Volcanic emanations could provide such a direct source with rapid precipitation of the iron as the thermal waters mix with the sea water. Iron derived from weathering of a landmass is likely to be precipitated in the shallow-water parts of a basin near shore. The Algoma-type iron-formation is characteristically present in eugeosynclinal volcanic-sedimentary rock assemblages, and not with continental shelf or deltaic rocks.

Primary sedimentary features indicative of deposition of gelatinous silica, mentioned in the previous section, are consistently present in this type of iron-formation. Penecontemporaneous deformation structures including crenulation, slumpage, folding, sliding, and brecciation, confined to single beds or to a limited group of beds are very common and are like those noted in many banded chert and fine-grained sediments. Rather unstable conditions in the depositional basin are suggested by these features. Gentle warping and upheaval of the basin floor, a phenomenon that is recorded around active volcanoes, would also trigger slumpage and deformation of beds that were in a semi-solid state. More violent upheavals or strong current action could cause brecciation of beds that were rigid enough to fracture.

The iron and chert are believed to have been precipitated mainly by inorganic processes, as evidence of significant organic activity is not found in these beds. Probably only very primitive organisms existed during the early Precambrian when many formations of this type were formed. Algoma-type formations in the Precambrian are similar in major respects to those found in the lower Palaeozoic and younger rocks. If organisms played an active part in the deposition of these rocks the evidence for their role is obscure. Thin beds of graphitic slate, containing variable amounts of iron-sulphide and less commonly economically significant amounts of other metallic elements, are present with early Precambrian iron-formations. The carbon in these beds was probably derived from primitive organisms but as yet its origin has not been demonstrated. Transitions from cherty iron-formation to black cherty graphitic slate with pyrite occur along strike in many formations. The black slates are believed to be closely related in origin to the iron-formations but to have formed in different parts of the sedimentary basin. They no doubt formed some distance from the source of volcanism and probably in deeper water where only very fine grained clastic material was contributed and carbon was preserved in waters low in oxygen or where primitive organisms were destroyed by toxic volcanic emanations.

Textures in Algoma-type iron-formations differ from those found in other cherty and siliceous iron-formations in that granules or oölites are absent or at least very uncommon. They are rarely found in Precambrian or Palaeozoic beds but some structures in Jurassic beds may be either granules or globules segregated during dehydration and consolidation of the chert. Several explanations have been offered to account for their general absence. One suggestion is that they have been obliterated during metamorphism and recrystallization of the rock. Evidence of these forms is preserved in highly metamorphosed beds of the Superior type and if ever present should be preserved in relatively unmetamorphosed cherty Algoma-type beds. Another suggestion is that the formation of granules and oölites is due to or at least facilitated in some way by organisms and as organisms or organic processes did not play an important part in the formation of this type of iron-formation, granular textures are absent. Most of the other types of iron-formation with oölitic textures were formed in shallow water where wave and tidal action may have caused agitation of the water. Organisms would likely be more abundant also in shallow near-shore waters. Physical-chemical factors affecting the rate of precipitation of iron and silica and agitation of the water are thought to be dominant factors in controlling the formation of these structures.

The consistent occurrence of Algoma-type iron-formations with volcanic rocks in many parts of the geological column leaves little doubt that they are related genetically to volcanism even though many aspects of their origin are not well known.

Superior Type

Superior-type iron-formations are widely distributed in late Precambrian rocks. They are associated with continental shelf and miogeosynclinal sediments composed of well-sorted detrital and chemical constituents derived from a maturely weathered landmass. Considerable volcanic material is interbedded with the sediments in most occurrences. They form long continuous formations extending for hundreds of miles, whereas Algoma-type formations are mainly lenticular and discontinuous in distribution. Granular to oölitic textures are characteristic of the Superior type and most have features that are distinctive of shallow-water deposits. It was the belief of many geologists in the past that this type of iron-formation occurred only in Precambrian rocks and formed in conditions that were unique in that time. More recent work by Blondel (1955) and O'Rourke (1961) indicates that, although they are most abundant in the Precambrian, they also occur in Palaeozoic rocks and possibly with some Mesozoic.

Most of the sediments associated with this type of iron-formation are typical of the assemblages formed under stable shelf conditions in an epineritic environment. Broad coastal belts of shallow water are visualized where water depth probably does not exceed 200 feet, and both clastic and chemical sedimentary constituents are contributed from a land area of relatively low relief. Coarser clastic sediments deposited near shore have crossbedding, ripple-marks, and other primary features typical of a shallow-water environment where wave and current action are prominent. Coarse clastic material grading seaward to clastic limestone, argillaceous limestone, and calcareous muds is to be expected. Farther from shore, where chemically precipitated sediments predominate over clastic sediments, thin chert layers and nodules and cherty limestone are interbedded with limestone and fine clastic sediments. Organic life is commonly abundant in the neritic zone and fossil shells and debris may be abundant in the limestones. Black carbon shales are usually prevalent in the transitional zone between neritic and deeper water environments.

These iron-formations thus were deposited under very uniform stable conditions on ancient continental shelves that extended for hundreds of miles. In the Labrador geosyncline the iron-formation exists as a continuous stratigraphic unit more than 600 miles long before being disrupted by erosion or structural features. Apparently a single iron-formation unit was once present around the entire shoreline of the Ungava craton for a distance of more than 2,000 miles. It is evident that the factors that led to the deposition of iron-formation existed on a very broad scale and were a major part of the tectonic framework of the time.

In most areas deposition of iron-formation took place following accumulation of considerable thicknesses of shelf type epineritic sediments, although in some places the Gunflint iron-formation was deposited directly on granitic basement rocks. In many belts the iron-formation overlies thin members of black carbon-rich slates, has the highest content of iron in the lower part of the formation, and the cherty upper iron-formation is overlain by thick carbon-rich shale or slate. These are in turn transitional upward to thin-banded, fine-grained clastic rocks interbedded with dolomite, coarse clastic rocks or tuff and extrusions. Deposition of the iron-formation took place where clastic sedimentation had practically ceased and in most belts coincided with a period of active volcanism. Lava flows and tuff are interbedded with the iron-formation in the Labrador geosyncline, in the Gunflint range, and in most other areas.

The tectonic setting in the Labrador geosyncline is thought to be representative of that present in most belts where this type of iron-formation was deposited. The continental shelves or miogeosynclinal belts were bounded by a landmass on one side and by a volcanic belt on the seaward side. In places where the miogeosyncline was narrow, volcanic rocks are interbedded with shelf sediments, but in other places broad zones of shelf sediment exist with very little volcanic material.

Thick rhythmic banded cherty iron-formation apparently was not formed in the succession of shelf-type sediments in the Labrador geosyncline before the initiation of volcanic activity in the off-shore belt. Banded cherts and chert breccias that contain negligible amounts of iron, such as the Fleming member in Labrador that lies between a dolomite member and the Wishart quartzite, may be composed mainly of silica derived from a weathered landmass. They are present as relatively thin lenticular members deposited in local basins or downwarps in the shelf floor. Their occurrence is analogous in many respects to the thin chert laminae and nodules present in many dolomites that are not associated with iron-formations.

On the other hand, the persistent thick cherty iron-formations are hundreds of feet thick, and apparently represent a different phenomenon. The silica in these beds is believed to be derived mainly from volcanic sources in the manner described previously and precipitated rapidly as it is carried into the shallow-water epineritic environment. It may of course be argued that this silica accumulated in the sea as it was released during weathering of the landmass and precipitated after clastic sedimentation had practically ceased, but the amount of water required to hold this amount of silica in solution while other constituents from the landmass were deposited would be vastly greater than the amount present in the shallow basin. These siliceous rocks are present along the entire length of the shallow-water belt and it must be concluded that the water in the whole basin was saturated with silica that must have been supplied in great abundance from a direct source.

Derivation of iron and silica from a deeply weathered landmass of low relief in a manner such as Sakamoto (1950) described is an interesting possibility. Seasonal alteration in the pH of the drainage waters is visualized as a result of alternate wet and dry periods in a hot climate. Groundwaters during the dry season would become alkaline and dissolve much silica, but with the influx of water in the rainy season the silica would be flushed from the deeply weathered zones and a decrease in pH would lead to solution of appreciable amounts of iron. If this process were to lead to the precipitation of hundreds of feet of iron-formation without any clastic material, very stable and uniform conditions would have to prevail over the land area and adjacent sea shelf for a very considerable period of time. It must be assumed also that the other chemical constituents that would be dissolved in quantity with the iron and silica must have been dispersed in the sea and removed from the shelf zone. Evidence for large deposits of residual aluminous material that would be expected to remain on the land surface or be deposited in later sediments seems to be lacking. Certainly the shale and slate directly overlying the iron-formation are not abnormally high in alumina. If the iron and silica were derived by a weathering process as simple as this it is rather surprising that formations of this type are not more common in the geological column and even forming today.

Some believe that orogenies have been more frequent in recent geological times so that sufficiently long quiet periods conducive to the development of cherty iron-formation in the manner suggested may have existed only during late Precambrian and occasionally during Palaeozoic time. This hypothesis seems unsatisfactory when there is direct evidence of volcanic activity and orogenic disturbance coincidental in time with the iron-formation deposition.

It is suggested therefore that the great abundance of silica in these sediments was derived from volcanic sources. Minor amounts of silica may also have been contributed from the land area by weathering. Less conclusive evidence is available concerning the source of the iron. Considering the recently published descriptions by Zelenov (1959) of the action of thermal springs in the Kurile volcanic area which leach silica, iron, and other constituents, and selectively deposit iron and silica near shore in the sea, there is little doubt that the iron also could be derived from a volcanic source.

Large quantities of iron could be contributed from a land area with a minimum of other constituents if it were previously concentrated in laterite deposits. A slight warping of the land probably coinciding with initial volcanic disturbances offshore would cause flooding of coastal areas. Swamps might develop where organisms could thrive and large quantities of organic acids and compounds might form from their residues in the stagnant water. The acidic waters from such a strongly reducing environment could dissolve large quantities of iron from the underlying iron-rich soils and transfer it to the neritic zone for deposition. It is interesting to note that where the Gunflint iron-formation rests directly on granitic rocks algae-like forms are found in the lower part of the iron-formation and rest directly on the granite. The granite appears fresh and unaltered and leaves no suggestion of being previously weathered. If the iron was previously concentrated in lateritic deposits and transported to the sea one would expect to find appreciable clay and alumina mixed with it but this is not so.

The role played by organisms in the precipitation of the iron and silica is not well known. Carbon in the associated slates in the Labrador geosyncline, according to evidence given by Harrison (1953), is of organic origin but whether organisms played a distinctive part in the formation of the iron sediments is a speculative point. A variety of blue-green algae and fungi have been identified in the non-ferruginous cherts of the Gunflint iron-formation of Tyler and Barghoorn (1954) but their significance in the genesis of these rocks is not known. The characteristic granular and oölitic textures in these formations are believed to be caused by inorganic processes but organisms could have had some part in their development.

Various facies of the Superior-type iron-formations reflect variation in the pH and Eh of the water in the depositional environment. The oxide and silicate-carbonate facies are most common and sulphide facies are usually absent or relatively uncommon. The sediments were deposited in the form of silica gels presumably from colloidal suspensions; syneresis, desiccation, and contraction cracks are evidence of considerable compaction and contraction of these gels during dehydration and diagenesis.

Clinton-type Iron-formation

The Clinton iron-formations differ from the Superior type in their lithological characteristics but not to a great extent in their environment of deposition. In the Clinton beds the iron content is considerably higher and silica content is considerably lower. Most of the silica is present in the iron-silicate mineral chamosite or in fine quartz grains in contrast to the extensive amount of chert in layers and in the matrix of the Superior-type iron-formation. The alumina content of the Clinton beds is five or ten times greater than that of the Superior type, and the phosphorus content, ranging from 0.5 to 1.5 per cent, is high. Both types are characterized by their granular and oölitic textures, and fossil shells and organic debris are common constituents of the Palaeozoic Clinton beds. It may be highly significant in the study of the origin of these formations that chamosite is the common iron-silicate mineral in Clinton-type and younger formations but greenalite and minnesotaite with very little chamosite are the principal primary silicate minerals in the Superior iron-formations. Both types are associated with shelf sediments and the Clinton type in particular shows primary features such as crossbedding, ripple-marks, or even raindrop impressions, indicative of shallow-water deposition. The Clinton beds are in general much thinner than the Superior beds but may be distributed along strike in one stratigraphic group of rocks for hundreds of miles and must therefore be considered as a regional rather than a local phenomenon.

Many of these beds are lenticular and pinch and swell along a single layer. For this reason beds thick enough to be mined economically are found in only certain parts of the Appalachian area. A smaller-scale feature that is well demonstrated in the Wabana beds is the occurrence of thicker lenticular shoots of iron-formation that appear to fill channels scoured by currents. These local runs or shoots may parallel the old shorelines or be irregular in pattern depending on the course of currents or tidal waves.

Sediments associated with the Clinton iron-formations are normal clastic rocks and calcareous beds that do not have an unusual iron content. Sandstones with shaly partings and black shales high in carbon are typically associated with this type of iron-formation. The hematite-siderite-chamosite beds interlayered with the clastic beds are well-defined discrete units of fairly consistent composition. The thicker iron-formation layers are usually massive and have distinct well-defined joint patterns with jointing perpendicular to the beds.

The general consensus of opinion in the past has been that the iron and other constituents in these beds have been derived, to quote Hayes (1915), "by long continued weathering of earlier crystalline and sedimentary rocks, the solution of their iron content by mineral and vegetable acids and subsequent transportation of the iron salts by streams into the sea." The conditions under which the Wabana beds of Newfoundland were formed, as determined by Hayes, are particularly interesting:

> A much longer period is represented by the Wabana ore series than would be required merely for the continuous deposition of such a thickness of rocks. About 200 feet below the Dominion bed, the rocks were, when soft, exposed above water, for raindrop impressions

> are here preserved pointing to an emergence shortly after their deposition. A change in fauna at the top of zone 2 occurs coincidently with a disconformity marked by an erosion surface and a deepening of the sea followed. The Scotia and Upper beds also give evidence of many oscillations in the relative level of land and sea. Professor Gilbert van Ingen has suggested that successive marginal tilting probably occurred contemporaneously with the deposition of the rocks of Bell island, causing the shore-line to advance and raising the underlying measures above sea-level. These soft sediments, with their large iron content, would then be available to further enrich the sea with iron solutions.
>
> The chemistry of the processes involved in the formation of the ore must of necessity be very complex. I have concluded from petrological and chemical investigations that chamosite, with perhaps other similar green iron silicates, are the most generally distributed primary iron-bearing constituents of both the ore and their accompanying rocks. The hematite is intimately associated with the chamosite and is by far the most abundant iron-bearing mineral of the ore....
>
> It has been shown that the ore of zones 2, 4, and 5, was formed as off-shore deposits on the bottom of a sea in which marine life was abundant, some organisms, such as worms, actually burrowing into the oölitic ferruginous sediments. The question regarding the mode of precipitation of the iron on the bottom of an open sea, still remains unanswered. It is possible that during the deposition of the hematite-chamosite ore, the sea was restricted by some means to a basin-like form, but was still sufficiently shallow to allow the migration of a fauna found in similar rocks in Europe. During the deposition of the material in which the oölitic pyrite is found in zone 3, a sufficient depth must have been attained for open currents to pass, carrying marine plankton, as is shown by the graptolites which occur in the oölitic pyrite and accompanying shales.

The higher alumina, calcium, and magnesium content is more in keeping with the composition to be expected if the constituents were leached from a deeply weathered landmass. From this viewpoint the iron-formation is distinctive as a high iron, low silica chemical sediment. The leaching of iron and other constituents with a minimum of silica can be expected if organic acids and compounds were abundant in a reducing instead of an oxidizing environment. Primitive algae and plant life on the land areas may have been very influential in causing solution of abundant iron. If this idea explains the source and manner of separation and contribution of the iron it has also to explain the abrupt and erratic appearance of iron-formation in an otherwiso normal succession of continental shelf or deltaic sediments.

Borchert (1959) has questioned the whole concept of deriving the iron from a landmass on the grounds that ferric iron being highly insoluble could not be removed in solution until after Devonian time when plant life became abundant, and derived organic acids and reducing environments resulting from their carbon decay could reduce the iron to the ferrous state. He believed that the iron in these beds was leached from sediments in the ocean in an off-shore belt of deeper water where the decay of abundant organisms produced a strong reducing environment, with low pH and high CO_2 content. Once in solution he believed the iron to have been transported by currents to areas of deposition where oxidation and pH conditions were different. Apparently any iron dissolved by such a mechanism in the oceans today is precipitated before being carried far and is disseminated with other sediments. If this process takes place on any appreciable scale it should be going on in parts of the ocean today but direct evidence of concentration of iron by this process on the scale found in these iron-formations has not been reported.

Direct association of volcanic rocks and evidence of volcanic activity coinciding with the deposition of this type of iron-formation have not been reported. Direct contribution of iron in solution to ocean currents could readily account for the erratic appearance of these beds in an otherwise normal or common type of sedimentary succession. If currents with an abnormally high iron content come into a near-shore environment slight changes in pH and Eh would cause precipitation of the iron as the current water mingled in a neritic zone. The actual volcanism and thermal springs contributing to these currents might indeed be some considerable distance from the shallow water zones where the iron was precipitated.

Evidently in the Wabana formations the shelf environment was fairly acidic, with a pH sufficiently low to exclude the deposition of limestone or dolomite. In other areas these rocks are directly associated with the iron-formation. At certain intervals, rather special conditions must have prevailed, such as a decrease in pH, for the iron-formations to be precipitated without a large amount of calcium and magnesium. A constant influx of acidic river waters with a high organic acid content could maintain lower pH conditions in the shallow near-shore area and thus prevent the deposition of much limestone. The persistent association of these beds with black shales in a shallow-water environment points rather directly to major organic factors in the solution and deposition of these ferruginous beds.

Primary deposition of chamosite, hematite, and siderite is indicated for much of the material in these beds with rather delicate and subtle fluctuation in the chemical environment to cause alternate deposition of ferric and ferrous iron. In general, Hayes (1915) showed chamosite deposition in oölites with concentric rings of hematite followed by carbonate deposition principally as a matrix around the granules and oölites. The primary nature of the hematite-chamosite oölites is indicated by delicate algae borings that cut through the concentric shells or spheres of the oölites. Extensive alteration and replacement of various constituents by siderite took place during diagenesis of these rocks. The complex diagenetic effects in these formations are still not well known and generalizations on this type of formation at this stage might be misleading.

In general, a terrestrial source of iron is favoured because of the association of alumina, calcium, and magnesia in quantities that might be expected from weathering in an acidic environment. Concentration of small quantities of TiO_2, V_2O_5, and MnO in this type of formation in contrast to their virtual absence in the cherty iron-formations may be due to the increased amount of organic carbon and the activity of organisms in the deposition of these beds. The oölitic pyrite seam above one iron bed at Wabana is believed to have been formed by sulphur-reducing organisms present in these sediments during deposition and diagenesis.

Minette-type Iron-formation

Iron-formations of this type are very similar to the Clinton type in texture, composition, and depositional environment. Beds rich in hematite are however not so abundant or so typical, and in general the Minette beds have a higher silica, ferrous iron, alumina, calcium, and magnesium content. Correspondingly, the total iron content is 10 to 15 per cent less in the Minette beds. Small amounts of TiO_2, Cr_2O_3, V_2O_3, and As, as indicated in analyses of some beds in Yorkshire, seem to be a significant feature of both types. The Minette iron-formations appear to cover smaller areas and may be more local in their distribution than the Clinton beds. The two types are comparable in thickness, ranging from a few inches to 15 feet with occasional shoots or channel fillings that have maximum thicknesses in tens of feet.

These marine chemical sediments are of limited areal extent and were apparently deposited in shallow bays and restricted arms of the sea. Local basins were apparently connected to the open ocean, as the presence of a normal fauna indicates, and the abundance of granules and oölites shows that the water was agitated frequently by wave or current action. Of the Yorkshire Lias Hemingway (1951) said that "The essential difference between the neritic shelf-cycle of the British Jurassic sequence and the shelf cycle of the Yorkshire Lias is manifestly the culmination to limestone on the one hand and to ironstone on the other. The many Jurassic limestones, though lithologically variable, even with subsidiary clays and marls, are typically the deposits of a widespread shallow sea. The ironstones, more limited in their areal extent, but also with thin shales are the deposits of small, shallow and restricted arms of the sea . . ." Repetition of depositional conditions are recorded by Dunham (1959) who said: "Three cycles of sedimentation may be recognized, in which the sequence is black shale-grey shale-sandy shale or sandstone-ironstone. The ideal cycle began with near-euxinic conditions, and continued with the deposition of mud in an infraneritic environment which gave place to sandier sedimentation before culminating in an ironstone phase." Describing these cycles in detail Hemingway (1951) concluded that

> . . . the topographically subdued land masses which bordered the Liassic sea were subjected to deep chemical weathering of a warm humid climate, probably associated with a dense plant cover. This produced on transport only fine grained clastics and chemical derivatives. Three times during Liassic times gentle tectonic pulses feebly uplifted the adjacent land areas and the sensitive neritic environment located in part of N.E. Yorkshire responded sedimentalogically in its three cycles. At the beginning of each tectonic phase clastic sediments were deposited in an infraneritic and even bathyal environment, which was later succeeded by an epineritic phase. Finally with the cessation of mechanical erosion, as peneplanation developed, only chemical derivatives were transported and deposited in an epineritic but restricted environment.

The occurrence of these iron-formations within a well-established sedimentary cycle coordinated with the cycle of tectonic events in the area, as defined by means of broader regional evidence, leaves little doubt about the sedimentary origin of the iron and its source from a weathered landmass. Borchert (1959) questioned the general source of iron and depositional process outlined here because of the occurrence of iron-formation below the North German Plain which appears to indicate that deposition took place in "wide regions of open sea" rather than in marine basins and troughs. His general ideas appear to have only limited application to the study of iron-formations in Yorkshire where a strong reducing shallow-water environment with restriction of clastic deposition has been demonstrated. The general environment of deposition as visualized by Dunham (1959) shows that:

> The presence of abundant decaying vegetable or animal matter is held to be fundamental to the formation of both the clay ironstones of the Coal Measures and the siderite-chamosite-oölites of the Jurassic, to provide the required conditions of low Eh, low pH and available CO_2. Oxide, silicate-carbonate, carbonate and sulphide facies are recognized in the Cleveland ores and associated rocks, and the primary chamosite oöliths are held to have been formed in the region between the carbonate and oxide zones rather than near the sulphide facies. It is noted that Al and Si in solution are necessary, in addition to Fe^{++}, for the formation of the silicate ores.

Three distinguishable phases of sedimentation were recognized by Taylor (1951) in the Northampton sand ironstones, which he describes as follows:

> a. The carbonate phase:— This phase was one in which iron was precipitated mainly as siderite and commonly in company with calcite. The deposits are dominantly non-oölitic though there was some formation of calcareous oöliths during this phase of sedimentation. Chamosite is a minor constituent and only sporadically present in the oölitic form. The chief rock types formed were siderite mudstones and sideritic siltstones, sandstones, and limestones. . . .
>
> b. The mixed phase:— The mixed phase is so called because it comprises oöliths, dominantly alumino-silicate but with some hydrated ferric oxide, set in a groundmass dominantly carbonate. It provides the major part of the workable ore in the Northampton Sand and the typical specimens of 'oölitic ironstone'. The commonest rock is a sideritic chamosite oölite, but types with chamosite or calcite in the groundmass, and with some limonite, siderite or kaolinite in the oöliths, also occur. These rocks are typically developed in the main Oölitic-Ironstone Group.
>
> c. The alumina-silicate phase:— This phase is characterized by the complete absence of the carbonates as primary minerals. Chamosite and kaolinite are the dominant constituents, and the rock types include chamositic and kaolinitic oölites and sandstones. They constitute the Lower and Upper Chamosite-Kaolinite Groups.

Some idea of fluctuation in the sea-floor environment can be deduced from Taylor's descriptions of the oöliths. He noted that,

> . . . the commonest type of oölith is built of chamosite in minute flakes arranged tangentially, often with a central core of amorphous chamosite. Sometimes alternating zones of a lighter and darker green colour are visible. These may represent zones of chamosite of slightly different composition, but more probably the lighter zones result from an ultra-microscopic association of kaolinite with the chamosite. Frequently oöliths are built round a broken fragment of an earlier chamosite oölith. . . .
>
> In some oöliths, alternating zones of siderite and chamosite are present, suggesting successive changes in the physico-chemical conditions — presumably in the concentration of the CO_3 ions — during the life of an oölith. A similar change in other cases after the formation of the chamosite oölith has led to marginal replacement by siderite or calcite.
>
> Frequently chamosite oöliths have a core of limonite. . . . Commonly a rapid alternation of zones of chamosite and limonite occurs and it is questionable whether the limonite results from direct precipitation or from oxidation of successive skins of chamosite. In either case an alternation of oxidizing and reducing conditions on the sea-floor is indicated.

It is also noted that some kaolinite oöliths show different degrees of flattening. These were obviously soft when deformed and did not develop through agitation on the sea bottom as many other oöliths appear to have done, but were possibly formed in place in a colloidal medium.

Non-oölitic and Clastic Iron-formations

Various occurrences are included in each of these types and very little generalization can be made regarding their origin. Some of the non-oölitic type are siderite bands deposited below coal beds or with high carbon sediments, some are sideritic or limonitic and formed in bogs or swampy depressions and later were buried, some are the product of volcanic exhalation of iron and silica on a small scale, and other conformable bands appear to be post-diagenetic replacements of sediments or volcanic rocks. Clastic beds range from heavy mineral assemblages concentrated in beach sands, bars, or alluvial fans to greywacke rocks that contain appreciable quantities of magnetite.

Metamorphism of Iron-formations

Regional metamorphism of most iron-formations has been isochemical with reconstitution of the elements into mineral assemblages appropriate to the metamorphic environment. As there are not less than four main types of iron-formation any of which may develop into four main facies (depending on the depositional environment), as well as great differences in the metamorphic environment, obviously a very large number of mineral assemblages are to be expected in metamorphosed iron-formations. Few of these mineral suites have been adequately described and, in general, studies on equilibrium conditions have been very limited. A few particular aspects of this topic that are important in appraising iron-formations as ore are discussed here, other metamorphic features are considered in the descriptions of particular occurrences of iron-formation.

Much of the study of metamorphic rocks has been directed towards gaining an understanding of mineralogical equilibrium under varied physical-chemical conditions and to defining the limits or conditions under which various minerals can form and coexist as stable rock constituents. The metamorphic changes affecting grain size and texture, which are of vital importance in determining the fineness of grinding required to separate minerals for concentration, have received relatively little study.

James (1955) outlined some of the features of iron-formation in different metamorphic zones in northern Michigan and illustrated textural changes and their implications in the formation of beneficiating ores and of soft ores derived through natural leaching of silica. The progressive increase in grain size found in zones of higher grade metamorphism, which he recognized, is even more prevalent in iron ranges in the Precambrian Shield of Canada. Metamorphic features in Algoma- and Superior-type iron-formations in particular are of major importance in Canada.

Textural features of specimens that consist of magnetite, hematite, and chert or quartz, and which show evidence of originally having granules and oölites of about the same size are shown in Plates III to V. These are representative of certain iron oxide zones or facies in Superior-type iron-formation or in metamorphic derivatives of it that are similar in mineral composition. Variations in texture with increasing metamorphic rank are illustrated in this series of pictures.

The sediments associated with the Gunflint iron-formation (Pl. III A) indicate that this range probably contains the least metamorphosed Superior-type iron-formation in Canada. The extremely fine grained character of the chert is of special interest, as well as the distribution of the iron oxide minerals as dusty disseminations in concentric rings in the oölites. A specimen selected from cherty oxide facies (Pl. III B) consisting of magnetite and quartz has about the same size of oölites or granules as the Gunflint specimen and is therefore thought to have had about the same original texture. The very fine grained quartz forming the cherty matrix of this iron bed is distinctly recrystallized to give a mosaic texture, and the iron oxide minerals show some recrystallization into discrete grains. Metamorphism was within the lower greenschist facies and the associated rocks contain chlorite, sericite, and, where the composition is appropriate, some albite plagioclase. The iron-formation illustrated in Plate IV A is from an area where metamorphism reached the epidote-amphibolite grade. The oölites and granules in Plate IV B are still distinct but the quartz is recrystallized and the iron oxide minerals have larger grains with distinct boundaries. Metamorphism in this specimen had reached the epidote-amphibolite grade.

Plate V A illustrates iron-formation where rocks show upper epidote-amphibolite grade of metamorphism. Note the coarser grain size of the iron oxide minerals and the indistinct outlines of granules or oölites, which results from the iron having been collected in fewer but large magnetite and hematite grains. The specimen of Plate V B is from iron-formations metamorphosed to the upper epidote-amphibolite grade. Note the well-defined mosaic of quartz grains, larger iron oxide grains, and almost complete disappearance of the original oölitic texture. This specimen was selected from the centre of a fairly massive bed and relict granules are still discernible. The bulk of the iron oxide grains in this material can be separated from the quartz by grinding the material to −35 to +48 mesh screen size.

Plate VIII A in this series illustrates a specimen with about the same grade of metamorphism as the previous specimen but with a conspicuous amount of recrystallization. Some of the iron-formations nearby are even coarser grained. None has perfectly uniform textures because of some variation in grain size between beds, but the specimens are representative of the bulk of the material of economic interest.

The mineralogical and textural changes are less complex in the above-illustrated oxide facies than for other facies of iron-formation. In the chert – iron-silicate – carbonate, or the magnetite – iron-silicate – carbonate – chert facies, the mineralogy is more complex but there too the average grain size increases as the grade of metamorphism becomes higher.

The genesis of the iron-silicate minerals in certain facies poses several difficult problems and is not fully understood. Regardless of origin, however, greenalite is the predominant iron-silicate in the relatively fresh and unmetamorphosed Gunflint iron-formation; minnesotaite is predominant in the greenschist metamorphic facies at Knob Lake; minnesotaite and grunerite are found in the lower part of the epidote-amphibolite facies southwest of Ungava Bay; grunerite with some hypersthene is predominant in the epidote-amphibolite metamorphic facies west of Ungava Bay and in the Wabush Lake and Mount Wright areas. A number of other iron-rich silicate minerals are found in these areas, but the above-named form the bulk of the material. Field observation consistently shows that these iron-silicates occur in individual beds with magnetite and chert or quartz and not with any appreciable amount of hematite. The iron-silicates listed contain ferrous iron for the most part and would be incompatible in a chemical system where the partial pressure or chemical potential of oxygen was suitable for the formation of ferric oxide or hematite (Abelson, *et al.*, 1959). Small amounts of silicate minerals may be present in the oxide facies in the epidote-amphibolite or amphibolite metamorphic zones. These are usually anthophyllite, actinolite, tremolite, diopside, and hedenbergite, and may represent silication of material that was formerly present as dolomite or ankerite.

The composition of the carbonate minerals is known to vary considerably, siderite, ankerite, and ferruginous dolomite being the most common. The presence of siderite has not been confirmed in areas where metamorphism is in the amphibolite facies or higher; indeed in many beds where the grade of metamorphism is much lower siderite breaks down with the development of magnetite. This latter reaction is well illustrated in contact-aureoles adjacent to diabase dykes or sills cutting iron-formation in the Michipicoten area and in the Gunflint iron-formation where iron from siderite is converted to magnetite. Less commonly iron from siderite appears to contribute to the formation of iron-silicate minerals (Tyler, 1949).

It is an accepted principle in the study of banded or laminated metamorphosed iron-formations that each band or lamina behaves as a separate chemical system and that stable minerals, formed under a given set of physical-chemical conditions, are appropriately developed to account for the composition of each band (Yoder, 1957; Abelson, *et al.*, 1959). Delicate interbanding of uniform layers consisting of grunerite, hypersthene, magnetite, and quartz, or layers of grunerite, a mixture of carbonates, and quartz is very common, and can be followed along strike for the length of the outcrops. It is of further interest to note that the proportions of magnetite and hematite appear to be constant within one band but vary from band to band or lamina to lamina in the iron oxide facies. James and Howland (1955) suggested that original differences in oxidation state are retained to high grades of metamorphism. This principle appears to be valid for many of the recrystallized iron oxide facies that have not been sheared, but numerous examples have been examined where platy specular hematite has formed in sheared iron-formation and plates of hematite cut across coarse granular aggregates of magnetite and hematite. It is difficult to determine whether there has been any change in the oxidation state of this material but coarse specular hematite was obviously the last iron oxide to recrystallize. These layers follow primary beds or laminae and there appears to be very little migration of elements outside these laminae during metamorphism.

Many iron-formations are not uniform in composition and mineralogy because of original changes in sedimentary facies within the formation; most of these compositional variations were preserved during metamorphism. Regardless of the rank of metamorphism, exploration for parts of an iron-formation suitable for beneficiation should be guided by detailed stratigraphic studies. Particular attention should be given to the detailed stratigraphic succession and distribution of the various facies or types of iron-formation in a regional belt or member, and sampling and beneficiation tests should be performed on individual facies that are well defined by stratigraphic and structural work.

Metamorphism is equally important in Algoma-type iron-formations. The increase in grain size with higher grade of metamorphism is less conspicuous, however, than in the Superior type. Most of the material examined from zones of amphibolite grade of metamorphism is considerably finer grained than Superior-type iron-formations of the same composition and metamorphic grade. Most of the Algoma-type iron-formations in the Precambrian are in the upper part of the greenschist metamorphic facies or higher, and the nature and characteristics of these beds when they were first consolidated as sedimentary rocks is uncertain.

Variations in grain size and texture of magnetite-hematite-chert iron-formation from different grades of regional metamorphism in northern Ontario are shown in Plates VI and VII. Plate VI A shows iron-formation from the lower greenschist metamorphic facies or muscovite-chlorite subfacies. The extremely fine grained chert and rather diffuse and dusty appearance of the iron-oxide minerals are characteristic. Plate VI B is from the upper greenschist metamorphic facies (biotite-muscovite subfacies) and the coarser grain size of both the chert and iron-oxide material can be seen. Magnetite cherty iron-formation from the upper greenschist metamorphic facies, which provides ore after concentration and grinding of a large proportion to −325 mesh screen size is shown in Plate VII A. Textural features of magnetite iron-formation from northwestern Ontario, which is associated with garnet-biotite-hornblende schist, are illustrated in Plate VII B. The coarse grain size is conspicuous.

The difference in grain size between Algoma- and Superior-type iron-formations of similar mineralogical composition and metamorphic grade is not understood. It is possible, because of difference in sedimentation or in the subsequent history prior to metamorphism, that the Superior beds contained more water trapped in the cryptocrystalline chert. During metamorphism the water vapour would permit greater mobility of the silica and iron ions and help to promote the growth of larger mineral grains.

The amalgamation of the iron oxide into coarser grains during recrystallization is well illustrated by changes in jaspilites during metamorphism. The deep red colour of jasper is caused by very finely disseminated dusty hematite in chert but jasper beds are not present where there has been much recrystallization. This loss in colour with recrystallization is found in both the north and south ends of the Labrador geosyncline where jasper-bearing formations pass into zones of higher grade metamorphism. The last traces of jasper in these transition zones are in isolated cherty granules that are not as coarsely crystalline as the surrounding material.

Significance of Metamorphism on Beneficiation of Ore

Both the texture and mineralogy of iron-formations are affected during metamorphism and the characteristic metamorphic textures are of particular interest and significance in the crushing and grinding stages of the preparation of an iron ore concentrate. The manner in which mineral grains break apart during crushing to give clean discrete particles of one mineral with a minimum of mixed mineral fragments depends largely on the extent to which recrystallization had taken place in the natural crude ore. Concentrates can be produced from the finer grained cherty iron-formations that have undergone very little recrystallization but much finer grinding is required. Observation of the grade of regional metamorphism may indicate whether grinding problems should be anticipated.

The mineralogy of the crude ore determines the method of concentration to be used. Iron present as magnetite can be easily concentrated by magnetic separation from either fine- or coarse-grained crude ore. If hematite or mixtures of coarse-grained magnetite and hematite occur, a concentrate may be obtained by the use of gravity methods, such as Humphrey spirals or jigs. Where these mixtures are fine grained, as they are in the less metamorphosed iron-formations, they not only require fine grinding to liberate the iron oxide but may need other techniques, such as flotation or reduction roasting followed by magnetic separation, to prepare concentrates. Other methods such as high intensity magnetic separation or high tension may be suitable for concentrating some of the finer grained ores.

Chapter VI

IRON ORES IN CANADA

The general distribution of major types of iron deposits is shown on Map 1187A. For the purposes of this report Canada may be conveniently divided into six major geological regions: the Appalachian region, east of the St. Lawrence River; the St. Lawrence and Hudson Bay Lowlands; the Canadian Shield, extending from the St. Lawrence River, the Great Lakes, Lake Winnipeg, and Great Slave Lake to the Arctic Ocean; the Central Plains, between the western border of the Canadian Shield and the eastern border of the Cordillera and from the 49th parallel of latitude to the Arctic Ocean; the Cordillera forming a northwest-trending mountainous belt along the west side of the continent; and the Arctic Archipelago extending north from the mainland to the polar regions.

Appalachian Region

The Appalachian region is composed of a northeasterly trending belt of complexly folded sedimentary and volcanic rocks that are metamorphosed in places and disrupted by igneous intrusions. These rocks range in age from Precambrian to Mesozoic but consist mainly of Palaeozoic geosynclinal assemblages. Clinton-type bedded deposits of hematite-siderite iron-formation are extensively developed in dark shale and quartzite of Ordovician age on Bell Island and under a large part of Conception Bay, Newfoundland. Although this occurrence, which has been mined for over half a century, is the largest of its type in Canada, other similar but much smaller deposits occur in Cape Breton and westward into Nova Scotia Peninsula. They were deposited in a similar geological environment to that of the beds at Bell Island but occur in both Ordovician and Silurian rocks. Farther southwest this type of iron-formation is of Devonian age and magnetite is the predominant mineral where the rocks are highly metamorphosed.

Of the iron-formations elsewhere in the Appalachian region, the Austin Brook Formation in northern New Brunswick is probably the best known. Although similar in most respects to facies of Algoma-type iron-formation in the Precambrian Shield, the Austin Brook occurs in Ordovician rocks. A somewhat similar but thinner cherty, magnetite-hematite iron-formation interbedded with fine-grained green Ordovician slate and schist occurs on the southeast shore of the Burin Peninsula in Newfoundland. Hematite and chert iron-formations rich in manganese are present in Silurian rocks in the Woodstock area and ferruginous clastic formations occur in the Shickshock-Gaspé belt of rocks. Residual deposits of hematite and goethite occur in carbonate zones in the Londonderry area of Nova Scotia.

Vein deposits composed of magnetite, specular hematite, goethite, or siderite are widely distributed in the Appalachian region. Magnetite and titaniferous magnetite deposits are associated with intermediate and basic igneous rocks near Stephenville in southwestern Newfoundland.

Precambrian Shield

Most of the iron deposits in the Precambrian Shield can be related to broad-scale stratigraphic and tectonic features. Deposits in iron-formations predominate but other types are common; Superior-type and Algoma-type occur in all the facies outlined in Chapter V.

Superior-type iron-formations are most extensively developed in the Labrador geosyncline. a belt of younger Precambrian rocks that extends for more than 700 miles from west of Ungava Bay southeastward through Schefferville to the Wabush Lake area then southwest for 200 miles to the Matonipi Lake area. Iron-formation is distributed mainly along the western half of the central part of this belt and is not highly metamorphosed. Because of the cherty character of the iron beds they were especially amenable to leaching of the silica and concentration of the iron by natural processes to give pockets and masses of direct-shipping Lake Superior-Knob Lake type hematite-goethite iron ore. More than 45 such deposits are present in a 10 by 50 mile area that runs northwest through the Schefferville-Knob Lake and Sunny Lake area of Quebec and Labrador, and iron occurrences of this type are present throughout the central geosyncline belt as far north as Finger Lake near Ungava Bay. The northern part of the geosyncline in the Ungava Bay area is more highly metamorphosed, and zones within the iron-formation are medium to coarse grained and form metataconite iron deposits suitable for beneficiation.

The iron-formation and Proterozoic metasedimentary rocks in the southern part of the geosyncline, from Wabush Lake area through Mount Wright and southwestward to Matonipi Lake, pass into the Grenville orogenic belt and are highly metamorphosed, folded, and faulted. Very large deposits of metataconite iron-formation in this southwestern part of the geosyncline are coarse grained and readily amenable to concentration. Three large deposits have been developed for production of ore concentrate; the area contains the largest reserves of low-grade ore on the continent.

Superior-type iron-formations also occur west of Ungava Bay in the Cape Smith-Wakeham Bay belt of Proterozoic rocks and along the east side of Hudson Bay, on the Hopewell, Nastapoka, and Belcher Island chains. They are also present near Sutton Lake south of Hudson Bay in northern Ontario in Precambrian rocks exposed through a window in Palaeozoic rocks. Superior-type iron-formations form the Mistassini-Albanel Lakes iron ranges in Central Quebec where they are exposed along strike for about 35 miles and contain large zones of relatively unmetamorphosed iron-formation that are potential taconite type of ore. The Gunflint iron range, which extends along the west shore of Lake Superior and southwest across the International Border to link up with the Mesabi iron range in Minnesota, forms still another prominent Proterozoic range. This iron range, except where cut by basic sills and dykes, shows little evidence of metamorphism and apparently contains no deposits of direct-shipping Lake Superior-Knob Lake type of ore of commercial size. Zones in the Gunflint range have been investigated as potential sources of taconite ore.

Most of the Algoma-type iron-formations occur in belts of volcanic and sedimentary rocks in other parts of the Precambrian Shield west of the Grenville belt. Several hundred bands of iron-formation have been mapped in belts of volcanic rocks in the central and western parts of the Precambrian Shield and many have proven to be suitable to provide high-grade ore concentrate. Most of these bands, although more or less metamorphosed, are fine grained and constitute taconite types of deposits. Siderite-pyrite iron-formations of this group have been mined for many years in the Michipicoten ranges.

Iron-formations occur in at least three areas in highly metamorphosed rocks in the southwestern part of the Grenville belt. Other iron-formations and associated high grade deposits occur on Baffin Island. These are highly metamorphosed and their primary type or affinity is not known.

The large hematite-goethite deposits within carbonate, leached slate, and tuffaceous rocks in the Steep Rock iron range are considered to be a residual type of occurrence and are a unique type in the Precambrian Shield. The hematite-goethite ore in siderite iron-formation of the old Helen mine in the Michipicoten ranges is classified as a Lake Superior-Knob Lake type of ore.

Various types of iron deposits occur within the Grenville Province of the Canadian Shield, which lies between the Georgian Bay-Mistassini Lake line and the St. Lawrence River and extends northeast from Georgian Bay to the Labrador coast. With the exception of the iron-formations of the south part of the Labrador geosyncline, which occur in the Grenville Province northeast of Mushalagan Lake, the largest iron deposits in the Grenville Province are associated with gabbro and anorthosite intrusions. The ilmenite deposits in anorthosite near Allard Lake and Baie St-Paul are well known and many titaniferous magnetite deposits are distributed throughout the Grenville Province. A large number of titaniferous magnetite deposits occur in gabbro and gabbroic anorthosite in the Lac St-Jean and Chibougamau areas and in the southwest part of the Grenville near the Ottawa River. Numerous small magnetite deposits are distributed in southeastern Ontario and in the adjoining parts of Quebec. They vary in type from titaniferous magnetite in gabbro to contact metasomatic, replacement, and vein types that are relatively free of titanium and apatite. Contact metasomatic deposits associated with intermediate to acid intrusions, such as the Bristol and Marmora deposits, are important ore producers.

Considerable magnetite associated with niobium, apatite, titanium, and rare earth elements is present in a number of alkaline rock complexes distributed in a zone running through Chapleau and Kapuskasing in northern Ontario. A large number and variety of magnetite, specular, and red hematite veins, mostly too small to be of economic interest, are distributed throughout the Shield.

St. Lawrence and Hudson Bay Lowlands

Magnetite deposits in the St. Lawrence Lowlands area actually occur within the deeply buried Precambrian rocks; their position has been verified by drilling through the Palaeozoic cover rocks. A few very thin lenses of Clinton-type iron-formation have been intersected by drill-holes in Silurian rocks in southern Ontario. Irregular-shaped replacement bodies of siderite occur in limestone along the Mattagami River in the Hudson Bay Lowlands.

Central Plains

Very few iron deposits are known in the Central Plains. Algoma-type iron-formation has however been intersected in the Precambrian basement rocks by drilling through Palaeozoic formations in the Choiceland area of Saskatchewan and near Neepawa in Manitoba. Beds of oölitic siderite iron-formation are present in Cretaceous rocks over a large area near Hines Creek in the Peace River area of Alberta. A few thin ferruginous beds have been reported in the Mackenzie River valley, and thin lenses and nodules of siderite occur in Cretaceous rocks in many parts in the Central Plains.

Cordillera

Many different types of iron deposits are found in the Cordillera. Contact metasomatic deposits near intermediate to acidic intrusive masses occur along the west coast and on the coastal islands and are the best known type in the region. A replacement mass of magnetite and specular hematite in volcanic rocks forms an important part of the mineralized zone of the copper deposit at the Craigmont property near Merritt in south-central British Columbia. A number of magnetite veins are distributed throughout this area, and towards the southeast, in the Kootenay area, veins of specular hematite are common. Magnetite veins and replacement masses are associated with copper-bearing zones near Whitehorse in Yukon Territory. Titaniferous magnetite deposits occur in ultrabasic rocks on Lodestone Mountain west of Princeton in southern British Columbia. Chert and jasper magnetite-hematite iron-formations lithologically like some Algoma-type iron-formation are present in the Snake and Bonnet Plume River area in Yukon Territory. Beds more than 200 feet thick in the lower Palaeozoic Rapitan Group, which have been traced for several miles, may prove to be an important source of beneficiating ore. Ferruginous chert and iron-formation occur in Mesozoic volcanic and sedimentary rocks on southern Vancouver Island and Saltspring Island. Thin clastic beds and lenses of titaniferous magnetite and siderite form an interesting type of iron-formation in the Belly River sandstone near Blairmore in the Foothills of Alberta. Bog iron deposits are widely distributed and have been reported in Taseko Valley, 65 miles northwest of Lillooet, along Summit Creek 38 miles east of Copper City on Skeena River, south of Nicola Lake, and at Alta Lake north of Squamish. Pyrite deposits are also widely distributed, probably one of the largest being the Ecstall River deposit 35 miles west of Kitimat. The pyrite and pyrrhotite tailings accumulated at Kimberley from the treatment of zinc and lead ores and iron sulphides of the Sullivan orebody constitute a reserve of iron ore in the Cordilleran region.

Arctic Archipelago

Highly metamorphosed iron-formations have been investigated along the southwest coast of Baffin Island, west of Clyde near the Barnes ice-cap, and in the northeast part of the island. A very large deposit of massive magnetite and hematite of exceptionally high grade and quality occurs about 50 miles south of Milne Inlet, the southern part of Pond Inlet, in northern Baffin Island.

Other iron deposits in the Arctic Archipelago are not well known due to limited exploration of this region. Magnetite occurrences have been reported around Admiralty Inlet on the north end of Baffin Island, one was found at the head of Strand Fiord on Axel Heiberg Island, and others are known along the north coast of Ellesmere Island. A thin bed of oölitic hematite iron-formation occurs in lower Palaeozoic rocks on Victoria Island.

Chapter VII

STATISTICAL SUMMARY OF THE CANADIAN IRON ORE INDUSTRY AND WORLD IRON ORE MARKETS

by T. H. Janes[1]

Introduction

Iron ore mining in Canada has had a long history marked by extended periods of little or no progress. Between 1886 and 1924, about 5.9 million long tons[2] of iron ore was produced, valued at about $16.8 million; from 1924 to 1938 no shipments were recorded. The modern era, dating from the resumption of iron ore mining operations in 1939 in northwestern Ontario, has been characterized, however, by widespread rapid growth and diversification. From 1939 to the end of 1962, shipments totalled 187,491,894 tons, valued at $1,615,933,731. Production in 1962 came from 22 mines (excluding those of by-product producers) operated by 16 companies in Newfoundland, Quebec, Ontario, and British Columbia.

Production

The decline in iron ore shipments experienced in 1960 and 1961 (Table XI) was sharply reversed in 1962 when shipments reached an all-time high of 24,909,565 long tons — up 37 per cent from 1961. Output of all producing provinces was higher.

In 1962 three mines in British Columbia and one in Labrador (Newfoundland) commenced production. A company started development of a property near Kirkland Lake, Ontario, for production from a one-million-ton-a-year pellet plant in 1964, and another continued development of a six-million-ton-a-year project in Labrador for production in 1965. Canadian iron ore productive capacity increased during 1962 from 26 million to nearly 30 million tons a year and by 1965 is expected to approach 45 million tons. The designed capacity of Canadian iron ore mines and mills operating at the end of 1962 was about 38 million tons a year. Two large mills were in the tune-up stage at the end of 1962, which accounts for the considerable difference between productive and designed capacity.

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[1] Mineral Resources Division, Department of Mines and Technical Surveys.

[2] The long or gross ton of 2,240 pounds is used throughout unless otherwise noted.

Table XI

Canadian Iron Ore Shipments, by Province, 1939 to 1962

(thousands of long tons)

Year	Newfoundland[1]	Quebec	Ontario	British Columbia	Other	Total
1939	1,404.1	—	110.4	—	—	110.4
1940	1,408.9	—	370.2	—	—	370.2
1941	1,182.2	—	460.7	—	—	460.7
1942	865.9	0.2	486.7	—	—	486.9
1943	889.4	—	444.9	—	127.7[2]	572.6
1944	584.6	—	494.0	—	—	494.0
1945	988.8	—	1,013.8	—	—	1,013.8
1946	1,235.2	—	1,383.5	—	—	1,383.5
1947	1,280.2	—	1,713.7	—	—	1,713.7
1948	1,717.8	—	1,194.0	—	—	1,194.01
1949	1,480.3	—	1,796.2	4.9	—	3,281.4
1950	1,044.2	—	2,174.8	—	—	3,219.0
1951	1,540.2	—	2,537.4	101.4	—	4,179.0
1952	1,476.7	—	2,426.3	804.0	—	4,707.0
1953	2,398.6	—	2,528.7	885.0	—	5,812.3
1954	3,355.8	580.8	2,158.0	478.3	—	6,572.9
1955	6,434.7	3,663.5	3,894.8	545.6	—	14,538.6
1956	7,556.8	7,104.0	4,962.7	330.3	—	19,953.8
1957	7,298.9	7,922.3	4,345.6	319.1	—	19,885.9
1958	4,813.2	5,411.0	3,254.4	562.6	—	14,041.3
1959	5,451.6	10,281.4	5,373.3	758.3	—	21,864.6
1960	6,795.9	6,658.9	4,754.6	1,027.6	4.8[3]	19,241.8
1961	6,795.8	5,035.7	5,154.2	1,192.0	—	18,177.7
1962[4]	6,914.8	10,343.3	5,939.6	1,711.9	—	24,909.6

Source: Dominion Bureau of Statistics

[1] Shipments included in Canada total in 1949, when Newfoundland entered Confederation, and subsequent years

[2] New Brunswick

[3] Alberta, test shipment

-nil

[4] preliminary

The Industry and Its Prospects

Table XII lists the companies in Canada that shipped iron ore from properties operated solely for the production of iron ore, the three companies that produced by-product iron ore, and the one company that produced ilmenite — the production of which is not included in iron ore shipments. Of the companies listed, three produced direct-shipping ore; one, direct-shipping ore and concentrate; seven, magnetite concentrate; three, hematite concentrate, mainly a spiral concentrate; two pelletized concentrate; and one produced sinter from siderite. In addition, three companies produced iron oxide pellets, calcine or sinter obtained as a by-product from the roasting of pyrite and pyrrhotite concentrate.

In 1962 the grade of iron ore shipments ranged from a low of 50.45 per cent iron for sinter from Algoma Ore Properties Division of The Algoma Steel Corporation, Limited to a high of 68.0 per cent iron in the pellets produced as a by-product

Table XII

Canadian Iron Ore Producers and Grade of Shipments, 1960 to 1962

Company and Property Location	Material Mined or Treated (percentage natural grade)	Process and Product	Shipments					
			('000 long tons)			(average analyses, per cent)		
			1960	1961	1962	1960	1961	1962
NEWFOUNDLAND								
Wabana Mines Division Dominion Steel and Coal Corporation Limited, Bell Island	Hematite-chamosite from underground and open pit mines (49.8 Fe)	Sink-float process Heavy media concentrate	2,808	2,292	1,275	50.01 Fe 13.13 SiO_2 0.88 P	50.38 Fe 12.93 SiO_2 0.90 P 0.10 Mn	51.01 Fe 12.64 SiO_2 0.90 P 0.10 Mn
Iron Ore Company of Canada, near Labrador City, Labrador	Specular hematite from open pit mine (37.5 Fe)	Spiral concentration Specular hematite concentrate	—	—	740	. . .	. . .	63.69 Fe 4.04 SiO_2 0.005 P 0.128 Mn
Iron Ore Company of Canada, near Schefferville, Que.	Hematite-goethite from open pit mines (53.42 Fe)	Direct shipping ore	3,941	4,448	5,115	53.37 Fe 6.16 SiO_2 0.089 P 1.037 Mn	54.588 Fe 6.09 SiO_2 0.088 P 0.956 Mn	55.805 Fe 6.27 SiO_2 0.081 P 0.719 Mn
QUEBEC								
Hilton Mines, Ltd., near Bristol, Que., 40 miles NW of Ottawa	Magnetite from open pit mine (approx. 17 Fe)	Iron oxide pellets	747	800	780	65.00 Fe 1.75 SiO_2 0.004 P 0.08 Mn	66.02 Fe 2.02 SiO_2 0.0048 P 0.08 Mn	66.28 Fe 2.34 SiO_2 0.005 P 0.08 Mn
Iron Ore Company of Canada, near Schefferville	Hematite-goethite from open pit mines (53.08 Fe)	Direct shipping ore	5,890	2,996	4,682	52.03 Fe 7.16 SiO_2 0.059 P 1.448 Mn	51.694 Fe 7.15 SiO_2 0.067 P 1.563 Mn	53.590 Fe 7.34 SiO_2 0.058 P 1.386 Mn
Quebec Cartier Mining Company, Gagnon	Specular hematite from open pit	Spiral concentration Specular hematite concentrate	—	1,240	4,620	. . .	64.41 Fe 4.87 SiO_2 0.021 P 0.06 Mn	64.50 Fe 5.00 SiO_2 0.021 P 0.04 Mn

ONTARIO								
Algoma Ore Properties Division, The Algoma Steel Corporation, Limited, near Wawa	Siderite from open pit and underground mines (34.3 Fe)	Ore beneficiated by sink float and sintered	1,437	1,644	1,567	50.41 Fe 10.83 SiO_2 0.18 P 2.81 Mn	50.36 Fe 11.19 SiO_2 0.015 P 2.84 Mn	50.45 Fe 11.49 SiO_2 0.011 P 2.81 Mn
Caland Ore Company Limited, E. arm of Steep Rock Lake N. of Atikokan	Hematite and goethite from open pit mines (52.59 Fe)	Direct shipping ore	755	1,009	2,003	53.29 Fe 6.44 SiO_2 0.029 P 0.25 Mn	52.46 Fe 7.85 SiO_2 0.027 P 0.33 Mn	53.74 Fe 6.10 SiO_2 0.023 P 0.23 Mn
Canadian Charleson, Limited, S. of Steep Rock Lake, near Atikokan	Hematite-bearing gravels (12 Fe)	Jig and spiral concentrate	112	18	119	55.32 Fe 6.02 SiO_2 0.030 P 0.64 Mn	55.76 Fe 6.16 SiO_2 0.031 P 0.62 Mn	55.4 Fe 5.8 SiO_2 0.030 P 0.55 Mn
Lowphos Ore, Limited, Sudbury area, 20 miles N. of Capreol	Magnetite from open pit mine (31.37 Fe)	Magnetite concentrate	519	578	401	59.52 Fe 7.61 SiO_2 0.13 P 0.50 Mn	59.81 Fe 7.18 SiO_2 0.015 P 0.43 Mn	60.04 Fe 7.35 SiO_2 0.017 P 0.037 Mn
Marmoraton Mining Company, Ltd., near Marmora	Magnetite from open pit mine (35-37 Fe)	Iron oxide pellets	282	529	408	64.3 Fe 4.75 SiO_2 0.006 P 0.087 Mn	65.1 Fe 3.8 SiO_2 0.007 P 0.083 Mn	64.4 Fe 3.95 SiO_2 0.008 P 0.074 Mn
Steep Rock Iron Mines Limited, Steep Rock Lake, N. of Atikokan	Hematite-goethite from open pit and underground mines (50.93 Fe)	Direct shipping ores and gravity concentrate	1,586	1,214	962	53.56 Fe 7.26 SiO_2 0.30 P 0.24 Mn	54.206 Fe 6.94 SiO_2 0.30 P 0.24 Mn	54.188 Fe 6.78 SiO_2 0.03 P 0.22 Mn

Table XII—Continued

Canadian Iron Ore Producers and Grade of Shipments, 1960 to 1962

Company and Property Location	Material Mined or Treated (percentage natural grade)	Process and Product	Shipments					
			('000 long tons)			(average analyses, per cent)		
			1960	1961	1962	1960	1961	1962
BRITISH COLUMBIA								
Brynnor Mines Limited near Kennedy Lake, Vancouver Island	Magnetite from open pit mine (+50 Fe)	Magnetite concentrate	—	—	410	. . .	. . .	61.4 Fe 5.0 SiO_2 0.05 P — Mn
Empire Development Company, Limited, Benson R., 25 miles S.W. of Port McNeill, Vancouver Island	Magnetite from open pit mine (48.4 Fe)	Magnetite concentrate	414	265	22	58.0 Fe 6.38 SiO_2 0.025 P	57.84 Fe 6.81 SiO_2 0.039 P	56.49 Fe 4.27 SiO_2 0.025 P — Mn
Jedway Iron Ore Limited, Moresby Island, Queen Charlotte Is.	Magnetite from open pit mine (51-52 Fe)	Magnetite concentrate	—	—	48	. . .	. . .	58.97 Fe 0.57 SiO_2 0.02 P 0.17 Mn
Nimpkish Iron Mines Ltd., 26 miles W. of Beaver Cove, Vancouver Island	Magnetite from open pit mine (41.6 Fe)	Magnetite concentrate	253	378	324	59.3 Fe 5.22 SiO_2 0.013 P 0.016 Mn	59.1 Fe 5.10 SiO_2 0.014 P	58.6 Fe 5.40 SiO_2 0.013 P 0.15 Mn
Texada Mines Ltd., Texada Island	Magnetite from open pit mine (41.20 Fe)	Magnetite concentrate	375	442	538	61.39 Fe 3.93 SiO_2 0.014 P 0.13 Mn	61.42 Fe 3.93 SiO_2 0.11 P 0.13 Mn	61.81 Fe 3.92 SiO_2 0.011 P 0.13 Mn
Zeballos Iron Mines Limited, near Zeballos, Vancouver Island	Magnetite from open pit mine (48 Fe)[e]	Magnetite concentrate	—	—	226	. . .	. .	62.6 Fe 5.45 SiO_2 0.01 P

BY-PRODUCT IRON ORE The Consolidated Mining and Smelting Company of Canada Limited, Kimberley, B.C	Pyrrhotite flotation concentrates are roasted for acid production. Calcine is pelletized and sintered (65.4 Fe)	Iron oxide pellets are further processed into pig iron at the plant	—	41	43	. . .	65.0 Fe 3.7 SiO_2 0.029 P	65.0 Fe 3.7 SiO_2 0.131 S 0.029 P
Noranda Mines, Limited plant at Cutler, Ontario, purchased by Canadian Industries Limited, Nov. 1962, operated by Cutler Acid Limited	Pyrrhotite and pyrite flotation concentrates treated	Iron oxide calcine and sinter	106	57	59	67.68 Fe 2.0 SiO_2 0.25 P 0.5-0.9 Mn	64.66 Fe 2.0 SiO_2 0.025 P 0.02 Mn	64.66 Fe 2.0 SiO_2 0.1 S 0.025 P 0.7-2.0 CaO
The International Nickel Company of Canada, Limited, mines and plant in Sudbury area, Ont.	Pyrrhotite flotation concentrates treated	Iron oxide pellets	192	231	257	68.0 Fe 1.5 SiO_2 0.03 P 0.05 Mn	68.0 Fe 1.5 SiO_2 0.03 P 0.05 Mn	68.0 Fe 1.5 SiO_2 0.3 P 0.5 Mn
BY-PRODUCT IRON Quebec Iron and Titanium Corporation produces ilmenite from Allard Lake, Quebec, used to produce pig iron and titania slag at company's electric smelter, Sorel, Quebec.	Ilmenite hematite from open pit mine (40 Fe, 35 TiO_2)	TiO_2 slag and various grades of desulphurized iron 'remelt iron' from smelter	864	1,032	750	. . .	. . .	. . .

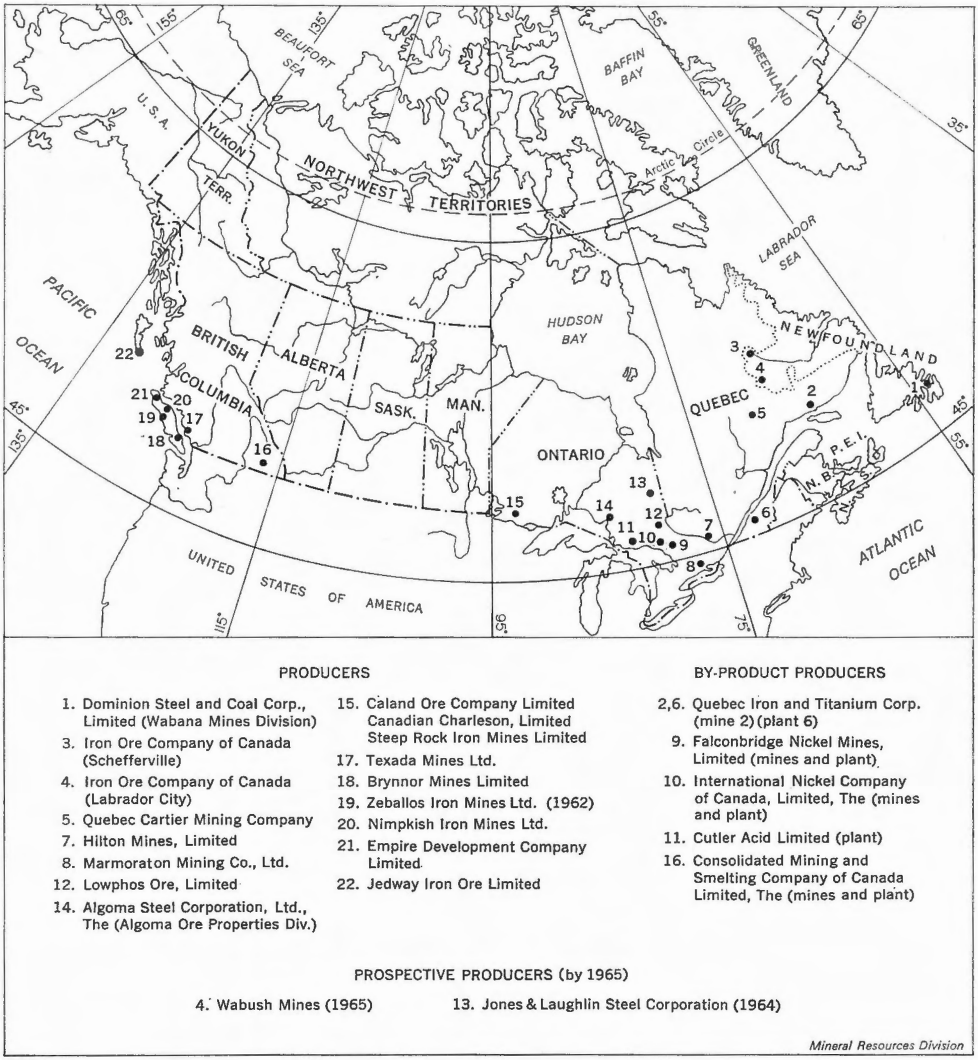
Source: Company annual reports, personal communications and other. — nil . . . not applicable e estimate by The International Nickel Company of Canada, Limited (*see* Table XII). Each year, producers of medium-grade iron ores carrying from 50 to 60 per cent iron are having increasing difficulties in marketing their output, as steel companies are demanding higher grade ores or agglomerates carrying up to 65 per cent iron.

The growth of the Canadian iron ore industry will be almost entirely dependent on the export market, mainly in the United States and to a lesser degree in Britain, western Europe, and Japan. Iron ore requirements of integrated steel producers in Canada are still being met largely by imports from the United States, but by 1966-67 about 80 per cent of these requirements should be supplied by Canadian producers. There is no longer a shortage of iron ore in the world such as was envisaged in the early 1950's, because many countries are rapidly developing very large reserves of high-grade, direct-shipping ore or beneficiating-type deposits in Africa, South America, Asia, and Australia. Canada will meet stiffer competition in major iron ore markets in the years ahead. Competition in the United States will be from domestic ores, South American, and African ores; in Britain and western Europe it will come mainly from Sweden, Africa, and South America; and in Japan, mainly from Asia, South America, and Australia.

The continuing trend of steelmakers in demanding improved ores is being satisfied either by developing orebodies that grade over 60 per cent iron or by beneficiating lower grade ores to produce a high-grade iron ore concentrate or agglomerate. Several countries in Africa, Asia, and South America have very large reserves of high-grade, direct-shipping ore. Canada possesses large reserves of iron ore but all of it must be classed as medium- or low-grade, direct-shipping or as concentrating grade. Therefore, to sell in the major market areas Canadian producers must concentrate, or concentrate and agglomerate their ores. With greater competition and an increasingly captive international iron ore market, large producers in Canada have intensified their ore research programs towards producing a beneficiated product for marketing whether it be from medium-grade, direct-shipping ore, or lower grade (34-38% iron) concentrating ore.

The ownership of captive ore sources by the North American iron and steel industry has always been a major factor in marketing North American ores; about 90 per cent of North American iron ore supply is produced by integrated steel and merchant ore companies. This practice has not been so pronounced in Britain and Europe, as traditionally, steel producers there have been purchasers of iron ore on the open market and have contracted supplies for relatively short periods. This pattern of iron ore supply has begun to change, and in recent years European producers have begun to participate in the financing of iron ore developments in Africa and, to a much lesser extent, in Canada. Imports of iron ore to Britain and western Europe are now approaching 50 per cent captive supply. Although there is a large potential market for Canadian iron ore in Europe it will be strongly competitive. The trend towards the use of iron ore pellets in blast furnaces, already pronounced in North America, is only beginning in Europe as sources of pellets are becoming available. Steel-producing countries in Europe will probably follow the pattern set in North America but less rapidly because of social and economic problems connected with iron ore mining in France, Britain, and West Germany, and because of the large capacity of integrated steel companies in producing sinter from screened, high-grade, direct-shipping ores. To market pellets in Britain and western Europe, Canadian producers will have to compete in price on an iron-unit basis plus perhaps a low premium for structure.

Japan has traditionally purchased nearly all its iron ore requirements in competitive world markets and obtained its supply from many countries, usually



GSC

FIGURE 4. Iron ore producers in Canada, 1962.

on a short-term basis. The pattern of using pellets for blast furnace burden that has been set in North America and which is just beginning in Europe is also gaining momentum in Japan. That country is beginning to participate in the financing of iron ore developments by lending funds for their development in other countries.

The Mineral Resources Division has periodically made forecasts of Canadian iron ore production projected on the basis of company intentions and probable iron ore developments in Canada and abroad. It appears that Canadian production by 1965-66 will be about 35 million tons a year, and by 1970 will range between 42 and 45 million tons a year. These estimates are dependent to a certain extent on developments that may occur in other countries such as increased integration from iron ore to finished steel products by European steelmakers and on the possible development of very large reserves of low-grade taconite ores in Minnesota and Michigan. Sufficient exploration has taken place to indicate that Canadian resources of iron ore are tremendous, amounting in the aggregate to more than 15 billion tons. The large tonnage deposits are those of the 30-38 per cent iron concentrating type, located mostly in the southern end of the Quebec-Labrador Trough and to the southwest of it in Quebec. Other large concentrating-grade orebodies exist in several other areas. It is meaningless to place an arithmetic figure on total reserves as insufficient detailed diamond drilling has been carried out for adequate assessment. There is, however, an absolute abundance of iron ore, more than sufficient to supply the domestic and export markets for generations to come.

Trade

A large part of the iron ore consumed by integrated iron and steel producers in Canada has been imported (*see* Table XIII) with nearly all of it coming from the head of the Great Lakes region in the United States. In 1962, of the 7.9 million tons consumed 4.6 million tons was imported, with over 4.4 million tons coming from the United States.

No iron ore was mined in Canada from 1924 to 1938, and it was not until 1945 that the domestic output exceeded a million tons a year. With nearly 80 per cent of pig iron capacity being concentrated in Ontario, it was necessary for the three large producers there to assure themselves of a continuing supply of iron ore. The Steel Company of Canada, Limited (STELCO) at Hamilton participated in the financing of many mines on the Mesabi Range in Minnesota and has obtained a major portion of its iron ore requirements from those mines. The Algoma Steel Corporation, Limited (Algoma) at Sault Ste. Marie, which has produced sinter from its own mines for many years, consumes some of the output in its blast furnaces, the remainder being marketed by an iron ore merchant company in the United States which in turn provides Algoma with American ore. Dominion Foundries and Steel, Limited (DOFASCO) at Hamilton brought its first blast furnace into operation in 1951 and since then has imported nearly all its requirements. STELCO and DOFASCO are participating in the financing of Wabush Mines, a six-million-ton-a-year pellet operation with mine in Labrador and pelletizing plant at Pointe Noire, Quebec, and by 1965 will begin to take their share of the mine's output. STELCO

Table XIII

Canadian Iron Ore Imports, by Source, 1950 to 1962

(thousands of long tons)

Country	1950	1951	1952	1953	1954	1955	1956	1957	1958	1959	1960	1961	1962
United States	2,657	3,295	3,667	3,579	2,621	3,973	4,362	3,778	2,985	2,404	4,342	3,959	4,449
Brazil	76	110	143	114	79	60	133	265	62	95	156	173	155
Liberia	—	8.2	1	27	10	19	31	—	—	.5	—	—	—
Other	9.2	7.9	.01	—	—	.01	.01	10	.2	.07	15	.4	—
Total	2,742	3,421	3,810	3,721	2,710	4,052	4,526	4,053	3,047	2,500	4,133	4,132	4,604

Source: Dominion Bureau of Statistics, *Trade of Canada* — nil

Table XIV

Canadian Iron Ore Exports, by Destination, 1950 to 1962

(thousands of long tons)

Country	1950	1951	1952	1953	1954	1955	1956	1957	1958	1959	1960	1961	1962[2]
United States	1,815	1,951	1,796	1,844	2,726	9,984	13,737	12,613	8,596	13,395	10,080	9,381	16,944
Britain	127	693	629	1,076	930	1,342	2,506	3,047	2,002	2,822	3,350	2,315	1,669
Japan	—	101	709	856	483	485	305	336	493	654	1,040	1,159	1,545
West Germany	48	135	301	528	693	1,036	1,088	1,097	812	736	939	726	568
Netherlands[1]	—	—	—	—	—	161	438	546	465	823	912	822	546
Belgium	—	—	—	—	—	—	—	146	27	82	164	348	262
Italy	—	—	—	—	—	—	20	109	—	41	74	104	89
France	—	—	—	—	—	—	—	79	—	—	—	12	23
Norway	—	—	—	—	—	.3	.09	—	—	—	—	—	—
Total	1,990	2,880	3,435	4,304	4,832	13,008	18,094	17,973	12,391	18,552	16,571	14,868	21,646

Source: Dominion Bureau of Statistics, *Trade of Canada*

[1] Most of which is for transshipment to West Germany [2] preliminary — nil

will continue to receive about 500,000 tons of pellets a year from Hilton Mines, Ltd. in Quebec which it owns jointly with Jones & Laughlin Steel Corporation of Pittsburgh. There is no indication that Algoma plans to change its sources of iron ore supply. Dominion Steel and Coal Corporation, Limited (DOSCO), with integrated steel facilities at Sydney, Nova Scotia, obtains nearly all its iron ore requirements from domestic sources.

With the changing pattern of iron ore supply, the proportion of domestic ore to imported ore consumed by Canadian integrated steel producers will rise so that towards the end of this decade about 80 per cent of the ore consumed is expected to be of domestic origin and 20 per cent imported. It is possible that, because of geographical considerations, a certain proportion of the iron ore requirements of Ontario's blast furnaces will continue to be met by imports from the head of the Great Lakes region in the United States.

From 1950 to 1962 iron ore exports rose from just under 2 million tons a year to over 21.6 million tons (*see* Table XIV). The United States, which has always been the largest market for Canadian iron ore, in 1962 took about 79 per cent of total exports, followed by Britain with 7.8 per cent, and Japan with 7 per cent. The United States will continue to be Canada's largest customer because of proximity and participation by United States steel and merchant ore companies in Canadian iron ore mines. Exports to Britain and western Europe have declined since 1957 due mainly to increased availability of good-grade iron ore from a number of countries in Asia and South America, and from Sweden. Canadian producers will be hard pressed in the next few years to maintain exports to Europe at even the 1962 rate but, as steel producers there adapt their blast furnace operations to the American practice of prepared feeds — such as pellets and close sizing of coke and limestone — the market for high-grade Canadian pellets should improve substantially. Up to 1963, pellets were not available for sale in Europe.

The market for Canadian iron ore in Japan in the next few years will probably not vary much from the 1.5 million tons shipped in 1962. Most ore contracts with that country are on a short-term basis with prices being negotiated each year. Japan draws her iron ore supply from a large number of sources; her market is extremely competitive and will become even more so as production is either expanded or commenced from iron ore deposits in western Australia and the west coast of South America.

Iron Ore Consumption, Pig Iron, and Raw Steel Production

The four fully integrated steel companies — STELCO, Algoma, DOFASCO, and DOSCO — form the nucleus of the Canadian steel industry (listed in magnitude of steel ingot capacity). Three other companies produce pig iron for sale but have no steelmaking facilities; one of these companies uses small amounts of iron ore while the other two obtain pig iron as a by-product of their main operations. The four integrated producers account for nearly all the Canadian iron ore consumption; small amounts are used by electric-furnace steel producers, who use scrap as the main raw material, and by ferroalloy producers.

Table XV

Iron Ore Consumption, Pig Iron and Steel Ingot and Castings Production in Canada, 1950 to 1962

Year	Iron Ore Consumption[1] (long tons)	Pig Iron Production[2] (net tons)	Steel Ingot and Castings Production (net tons)
1950	3,971,734	2,317,121	3,383,575
1951	4,719,787	2,552,893	3,568,720
1952	5,082,597	2,681,585	3,703,111
1953	5,229,834	3,012,268	4,116,068
1954	3,812,366	2,211,029	3,195,030
1955	5,583,041	3,215,367	4,534,672
1956	6,385,508	3,568,203	5,301,200
1957	5,965,805	3,718,350	5,068,149
1958	4,697,347	3,059,579	4,359,466
1959	5,812,982	4,182,775	5,901,487
1960	6,814,269	4,278,425	5,789,570
1961	7,441,795	4,925,395	6,466,324
1962[3]	7,868,626	5,288,589	7,173,475

Source: Dominion Bureau of Statistics, *Primary Iron and Steel*

[1] Indicated consumption (i.e. shipments plus imports minus exports): no account is taken of changes in consumers' stocks

[2] Commencing in 1959 'remelt' iron, produced in the electric smelting of ilmenite at Sorel, Quebec, is included

[3] preliminary

Table XVI

World Production of Iron Ore, Iron Ore Concentrates and Iron Ore Agglomerates, by Country, 1950, 1955 and 1960 to 1962

(thousands of long tons)

Country	Av. Iron Content[1] %	1950	1955	1960	1961	1962
U.S.S.R.	55	..	70,800	104,343	116,137	126,072
United States	49	98,046	102,999	88,784	71,329	72,324
France	35	29,509	49,525	65,907	65,525	65,568
Canada	55	3,219	14,539	19,242	18,199	24,228
China	50	1,968	8,600	54,100	59,053	49,200[2]
Sweden	60	2,055	17,080	20,975	22,771	21,672
Venezuela	60	187	8,306	19,182	14,425	13,044
Britain	30	12,935	16,175	17,088	16,512	15,276
West Germany	30	10,710	15,436	18,571	18,568	11,244
Subtotal		158,629	303,460	408,192	387,766	398,624
Other Countries		84,470	62,240	104,137	109,844	115,000[2]
World Total		243,099	365,700	512,329	497,610	513,624[2]

Source: U.S. Bureau of Mines, *Minerals Yearbook* (various years) and, for 1962, UN *Monthly Bulletin of Statistics, March 1963*

[1] U.S. Bureau of Mines, *Minerals Yearbook, 1961* (estimate)

.. not available

[2] estimate

Table XVII

World Trade in Iron Ore, Iron Ore Concentrates and Agglomerates, Selected Years, 1950 to 1961

(thousands of long tons)

Country of Origin	Year	Production	Approx. Iron Content %	Total Exports	Destination of Exports by, Main Countries							
					United States	West Germany	Belgium Luxembourg	Britain	Japan	Netherlands	Saar	Canada
France	1950	29,509	33	11,024	1	137	6,762	360	—	..	3,596	—
	1955	49,517	33	20,751	—	350	12,350	640	—	139	7,262	—
	1960	65,907	35	26,726	—	9,625	16,563	531	—	6	b	—
	1961	65,525	..	24,845	—	9,373	15,025	450	—	—	b	—
Canada	1950	3,219	55	1,989	1,814	47	—	128	—	—	—	—
	1955	14,539	55	13,006	9,984	1,036	—	1,343	485	160	—	—
	1960	19,242	55	16,942	10,433	1,598	164	3,360	1,041	272	b	—
	1961	18,199	..	14,782	9,685	1,437	—	2,334	1,097	146	b	—
Venezuela	1950	198	64	—	—	—	—	—	—	—	—	—
	1955	8,306	64	8,305	7,823	314	—	151	c	—	—	—
	1960	19,182	60	19,271	14,842	1,716	33	1,679	—	—	b	15
	1961	14,425	..	14,644	10,480	1,573	—	1,694	—	—	b	—
Sweden	1950	13,396	61	12,739	2,013	3,816	1,449	3,420	—	..	—	9
	1955	17,080	60	15,407	1,204	6,315	2,432	4,068	—	248	—	—
	1960	20,957	60	19,406	94	8,215	3,618	4,824	—	592	b	—
	1961	22,771	..	18,786	78	8,292	4,109	4,670	—	657	b	—
U.S.S.R.	1950	42,210	..	3,169e	—	—	—	—	—	—	—	—
	1955	70,800	55	8,661e	—	—	—	—	—	—	—	—
	1960	104,343	55	14,942	—	—	—	—	—	—	b	—
	1961	116,137	..	16,174	—	—	—	—	—	—	b	—
Chile	1950	2,929	60	2,555	2,555	—	—	—	—	—	—	—
	1955	1,693	64	1,218	1,056	110	—	—	—	—	—	10
	1960	4,615	60	5,109	3,750	636	11	—	303	4	b	21
	1961	5,142	..	5,776	2,604	499	27	—	2,403	36	b	—
Brazil	1950	1,956	68	876	675	22	7	18	—	..	—	104
	1955	4,086	68	2,525	1,091	391	—	544	9	50	—	30
	1960	9,197	70	5,079	1,407	1,361	—	667	366	165	b	136
	1961	9,193	..	5,894	888	2,468	10	569	393	155	b	119
Malaya	1950	499	45	521	—	—	—	—	521	—	—	—
	1955	1,466	45	1,592	—	—	—	—	1,505	—	—	—
	1960	5,641	55	5,500	—	9	—	—	5,374	—	b	—
	1961	6,734	..	6,545	—	—	—	—	6,545	—	b	—

Peru	1950	—	—	—	—	—	—	—	—	—	—	—
	1955	1,703	62	1,711	1,711	—	—	—	—	—	—	—
	1960	6,880	60	5,111	2,732	1,577	—	14	652	—	b	—
	1961	5,567	..	5,334	1,211	1,526	—	139	2,348	—	b	—
United States	1950	98,045	49	2,551	—	—	—	c	—	—	—	2,551
	1955	103,003	49	4,517	—	—	—	—	285	—	—	4,232
	1960	88,784	49	5,236	—	—	—	—	839	—	b	4,391
	1961	71,329	..	4,953	—	130	—	—	884	—	b	3,925
Liberia	1950	—	—	—	—	—	—	—	—	—	—	—
	1955	1,870	68	1,716	857	286	—	208	—	294	—	60
	1960	3,003	66	2,916	782	1,049	—	542	—	401	b	31
	1961	3,200	..	2,826	719	937	—	418	—	514	b	—
India(a)	1950	2,958	53	55	—	—	—	—	55	—	—	—
	1955	4,653	53	1,287	—	148	24	—	903	—	—	—
	1960	10,514	61	3,415	—	94	—	—	1,966	c	b	—
	1961	12,061	..	1,987	—	—	—	—	1,666	—	b	—
All others	1950	50,081		21,020	1,032	864	45	4,179	770	..	42	—
	1955	80,005		16,502	796	4,689	204	5,760	1,678	622	22	9
	1960	116,389		21,775	254	6,647	44	5,823	4,384	685	b	15
	1961	147,327		21,535	185	5,972	1,068	4,498	5,038	849	b	—
World Total	1950	250,000		40,591	8,090	4,886	8,263	8,105	1,346	..	3,638	2,664
	1955	365,721		97,198	24,522	13,639	15,010	12,714	4,925	1,513	7,284	4,341
	1960	512,329		151,428	34,264	32,727	20,433	17,440	14,967	2,125	b	4,609
	1961	497,610		144,081	25,850	32,207	20,239	14,972	20,374	2,357	b	4,044

Source: United States Bureau of Mines: *Minerals Year Book*, 1952, 1957 and 1961. American Iron and Steel Institute, *Annual Statistical Report, 1961*; all 1961 export figures are from The Cleveland Cliffs Iron Ore Company, *Cliffs Iron Ore Analyses, 1963*.

(a) Does not include Goa.

(b) Included in West Germany

(c) Under 500 tons

— nil

.. not available

World Production of Iron Ore

The nine countries listed in Table XVI accounted for about 78 per cent of the world's total iron ore output in 1962. Since 1950 production from Canada, Sweden, and Venezuela has risen dramatically; production from France and the U.S.S.R. has risen at a much slower rate; that from the United States, West Germany, and Britain has declined from the 1955 rate of output. The trends indicated in Table XVI will probably continue, output from West Germany and Britain will likely decline at a faster rate, and that from France may also decline from the present high level. In each of these countries the grade of ore mined ranges from 28 to 35 per cent iron and presents metallurgical problems in blast furnace production of pig iron because the ores must be blended and coke consumption rates are high. The tendency to import high-grade ores to replace domestic ores of lower grade is increasing in the three countries.

In addition to the countries listed, India, Brazil, Peru, and Chile are substantial producers of iron ore. Several emerging countries of Africa, particularly Liberia, Mauritania, and possibly the Republics of Guinea and Gabon, will also become substantial iron ore producers. The total output from these west African countries by the late 1960's will probably be from 20 to 25 million tons a year, most of which will be destined for markets in Britain and western Europe. Australia will probably supply Japan with 3 to 5 million tons of iron ore a year by the end of the decade.

World Trade in Iron Ore

In 1961, of the total world production of 497.6 million tons, 144.1 million tons was classified as exports, with France, Sweden, the U.S.S.R., Canada, and Venezuela, in that order, being the main exporting nations. The main importing nations were West Germany with 32.2 million tons, the United States with 25.9 million tons, followed in order by Japan (20.4), Belgium-Luxembourg (20.2), and Britain (15.0).

It will be noted that the output from France has an average iron content of about 35 per cent, that from the United States about 49 per cent, whereas that from most other countries ranges from 55 per cent to as high as 68 for Brazil and Liberia. Within the next decade, production from countries with shipments of high-grade ores or concentrates will probably increase, whereas production of direct-shipping ores of low- or medium-grade will decline.

World Production of Pig Iron and Raw Steel

Iron ore consumption in any country is based on pig-iron production and the grade of iron ore that is used in the blast furnaces. For example, the production of 289.4 million net tons of pig iron in 1961 (Table XVIII) might in a broad sense be balanced against world iron ore production of 497.6 million long tons, which would give an approximate iron content of ores consumed of 58 per cent. In other words, about 1.71 tons of iron ore were required per ton of pig iron produced. In France, where the average grade of ore used in blast furnaces is in the order of 35 per cent iron, about 2.85 long tons of iron ore would be required, whereas a country such as Sweden which uses only high-grade ores and agglomerates, grading better than 60 per cent iron, would require about 1.68 tons of iron ore. The average consumption of iron ore per ton of pig iron produced in any country can be estimated in a similar manner.

Table XVIII

World Production of Pig Iron[1], by Country, Selected Years 1950 to 1962

(thousands of net tons)

Country	1950	1955	1960	1961	1962[3]
United States	66,400	79,264	68,566	66,565	67,595
U.S.S.R.	21,500	36,376	51,540	56,100	60,053
China	1,120	4,000	30,300	24,300[3]	31,967
West Germany[2]	12,306	21,284	28,372	28,021	26,732
Japan	2,534	5,982	13,604	18,061	19,808
France	8,565	12,216	15,835	16,367	15,381
Britain	10,789	13,966	17,655	16,540	15,334
Belgium and Luxembourg	6,827	9,342	11,396	11,330	11,416
Czechoslovakia	2,262	3,307	5,176	5,480	5,710
Poland	1,640	3,439	5,030	5,258	5,456
Canada	2,458	3,334	4,175	4,701	5,159
Italy	632	1,912	3,113	3,528	3,951
Other Countries	9,349	17,245	30,238	33,099	34,607
Total	146,382	211,667	285,000	289,350	303,169

Source: American Iron and Steel Institute, *Annual Statistical Reports, 1959, 1960 and 1962*

[1] Includes ferroalloys made in blast furnaces

[2] Includes the Saar

[3] estimate

Table XIX

World Production of Steel Ingots and Castings, by Country, Selected Years, 1950 to 1962

(thousands of net tons)

Country	1950	1955	1960	1961	1962[2]
United States	96,836	117,036	99,282	98,014	98,328
U.S.S.R.	30,400	50,265	71,971	77,990	84,106
West Germany[1]	15,453	26,992	37,590	36,881	35,898
Japan	5,343	10,370	24,403	31,160	30,366
Britain	18,248	22,313	27,222	24,736	22,950
France	9,536	13,872	19,069	19,375	19,006
Belgium and Luxembourg	6,878	10,060	12,416	12,252	12,504
Italy	2,562	5,945	9,071	10,059	10,409
China	600	3,145	20,337	13,200	8,800
Czechoslovakia	3,197	5,000	7,460	7,761	8,420
Poland	2,767	4,868	7,364	7,973	8,410
Canada	3,343	4,500	5,790	6,466	7,173
Other Countries	12,666	23,110	39,607	44,195	47,884
Total	207,829	297,476	381,582	390,062	394,254

Source: American Iron and Steel Institute, *Annual Statistical Reports, 1959, 1960 and 1962*

[1] Includes the Saar

[2] estimate

Table XX

Imports[1] of Iron Ore, Iron Ore Concentrates and Agglomerates, by Major Importing Countries, Selected Years, 1950 to 1961

(thousands of long tons)

Importing Country	1950	1955	1960	1961
West Germany	4,793	14,099	33,122	32,157
United States	8,216	23,443	34,585	25,807
Japan	1,438	5,373	14,688	20,694
Belgium—Luxembourg	8,124	14,731	20,337	20,215
Britain	8,392	12,859	17,968	14,966
Czechoslovakia	..	..	7,097	7,844
Poland	235	4,337	7,204	7,548
Canada	2,742	4,052	4,515	4,132
Italy	181	823	2,580	3,255
Netherlands	799	1,043	2,284	2,219
France	165	544	1,482	1,672
Total	35,085	81,304	145,862	140,509

Source: *Statistical Summary of the Mineral Industry (1956-1961)* Mineral Resources Division, Overseas Geological Surveys, Britain

[1] Includes manganiferous iron ores but does not include burnt pyrites

.. not available

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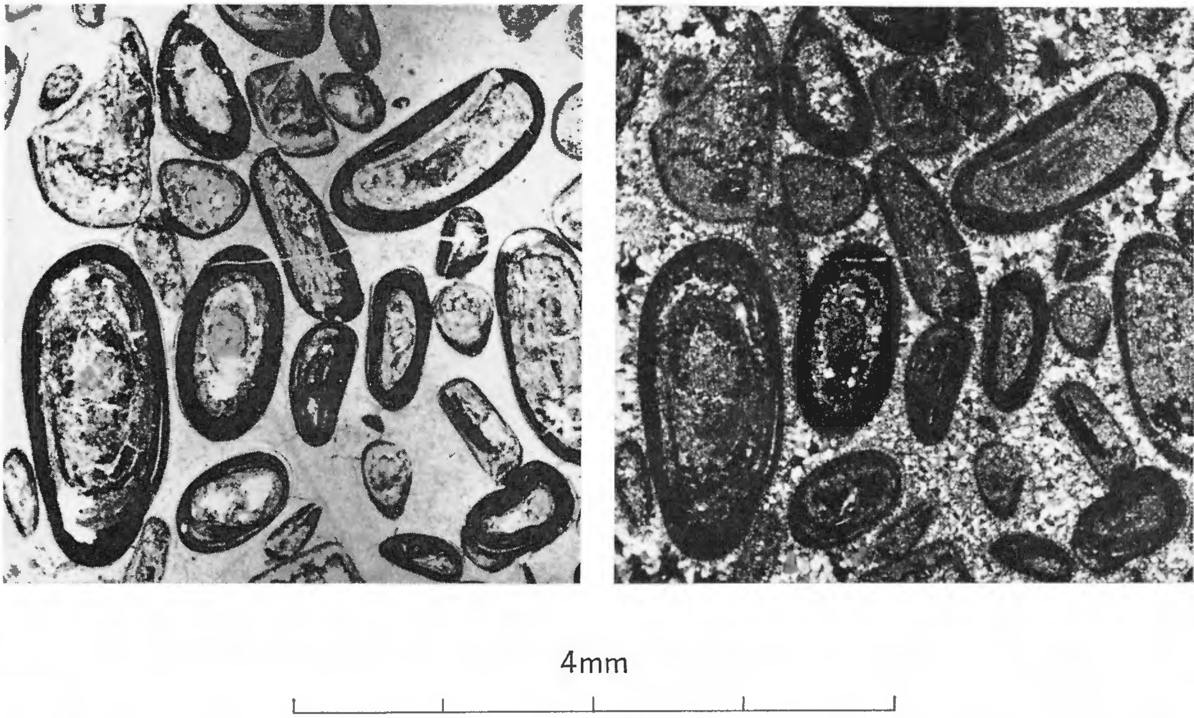
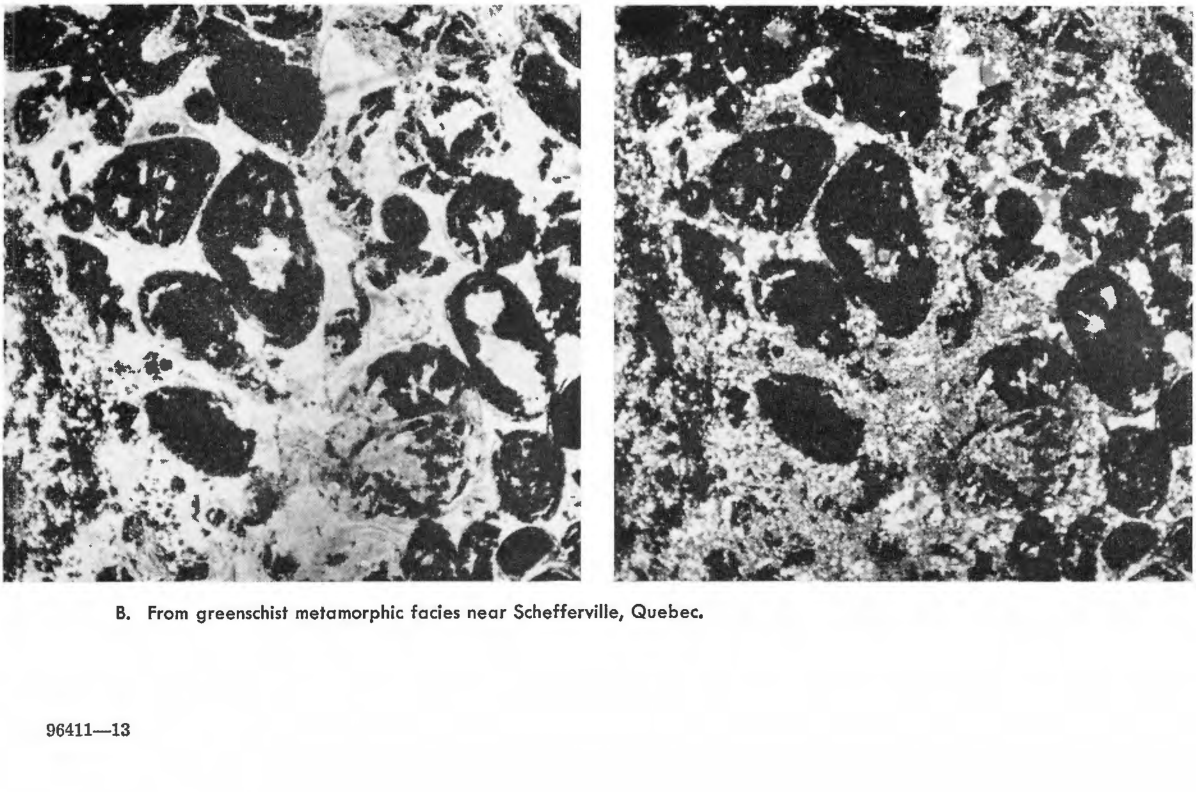
PLATE III. TEXTURES OF SUPERIOR-TYPE MAGNETITE-HEMATITE-CHERT IRON-FORMATION.
(Photomicrographs with transmitted light on left and with crossed nicols on right.)

A. From the Gunflint Range near Port Arthur, Ontario, with very little metamorphism.

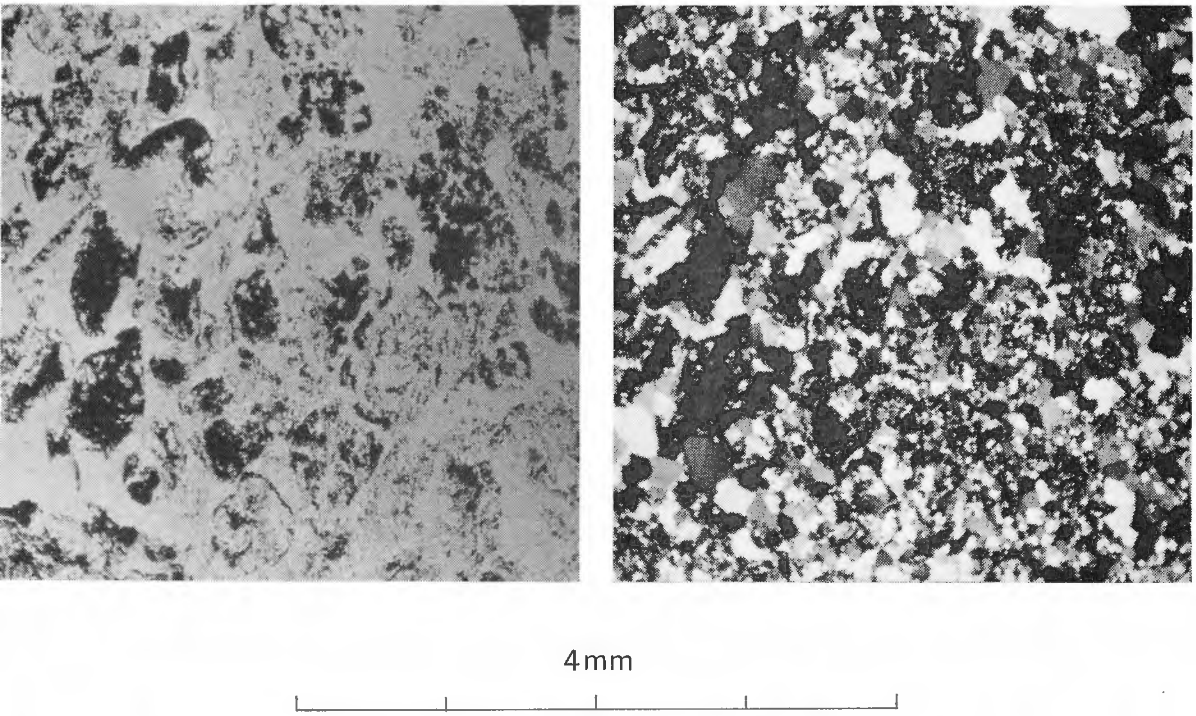


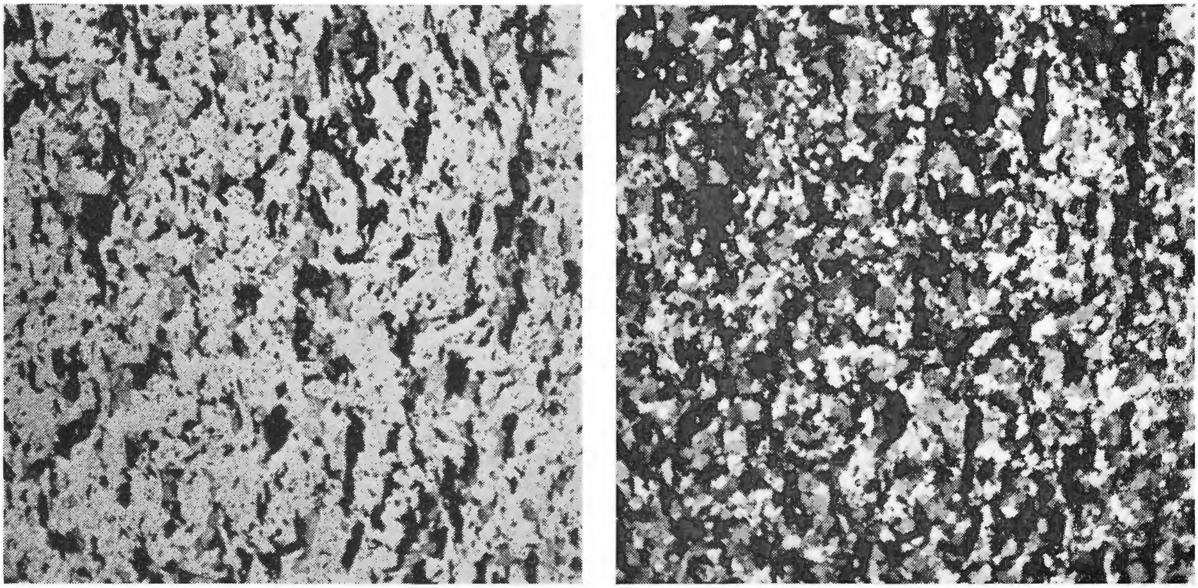


B. From greenschist metamorphic facies near Schefferville, Quebec.

PLATE IV. TEXTURES OF SUPERIOR-TYPE MAGNETITE-HEMATITE-QUARTZ IRON-FORMATION.
(Photomicrographs with transmitted light on left and with crossed nicols on right.)

A. Lower epidote-amphibolite metamorphic facies from southwest of Ungava Bay, Quebec.



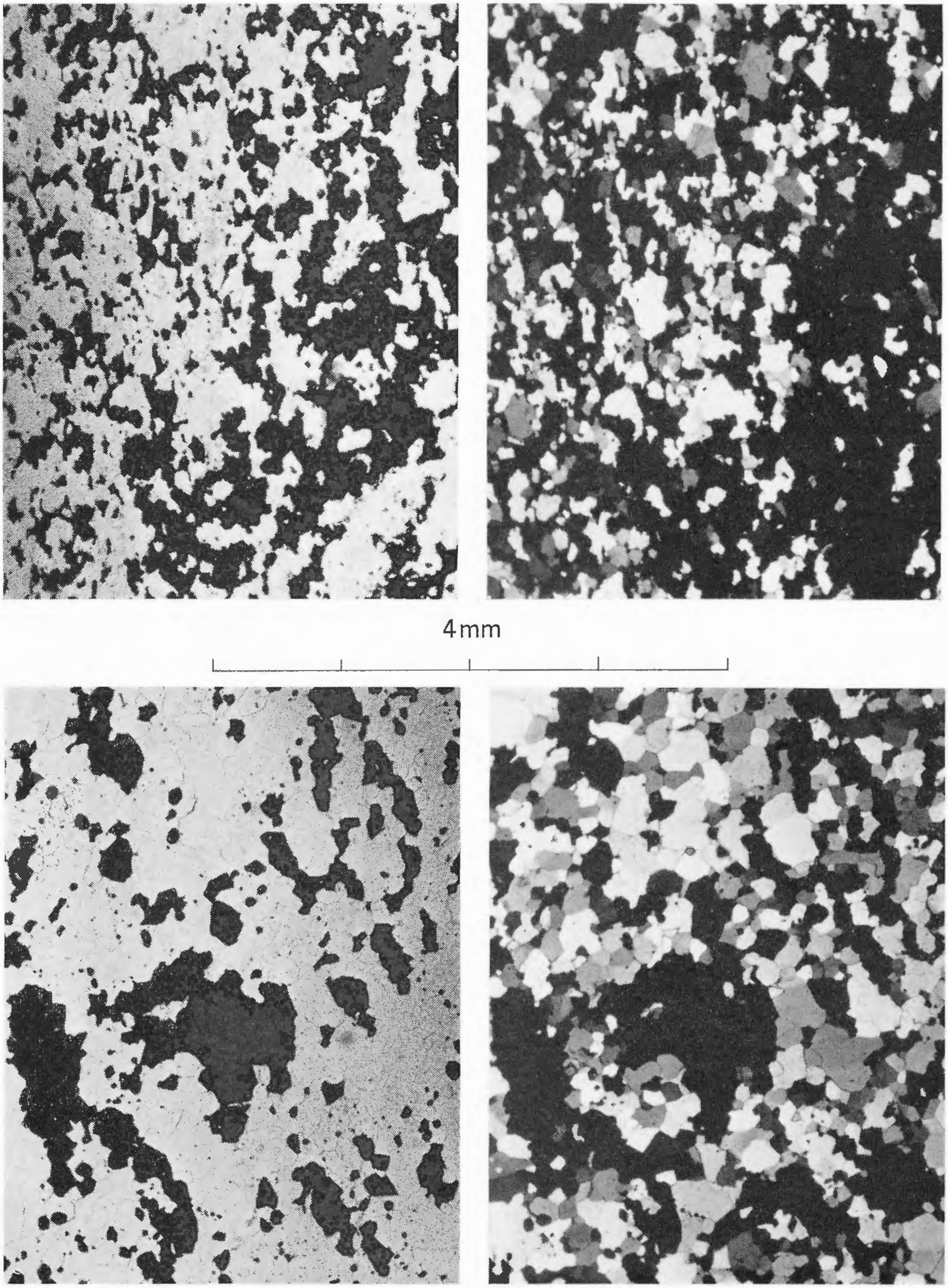


B. Epidote-amphibolite metamorphic facies from west of Ungava Bay, Quebec.

PLATE V. TEXTURES OF SUPERIOR-TYPE MAGNETITE-HEMATITE-QUARTZ IRON-FORMATION.

(Photomicrographs with transmitted light on left and with crossed nicols on right.)

A. Upper epidote-amphibolite metamorphic facies from Wabush Lake area, Labrador, Newfoundland.

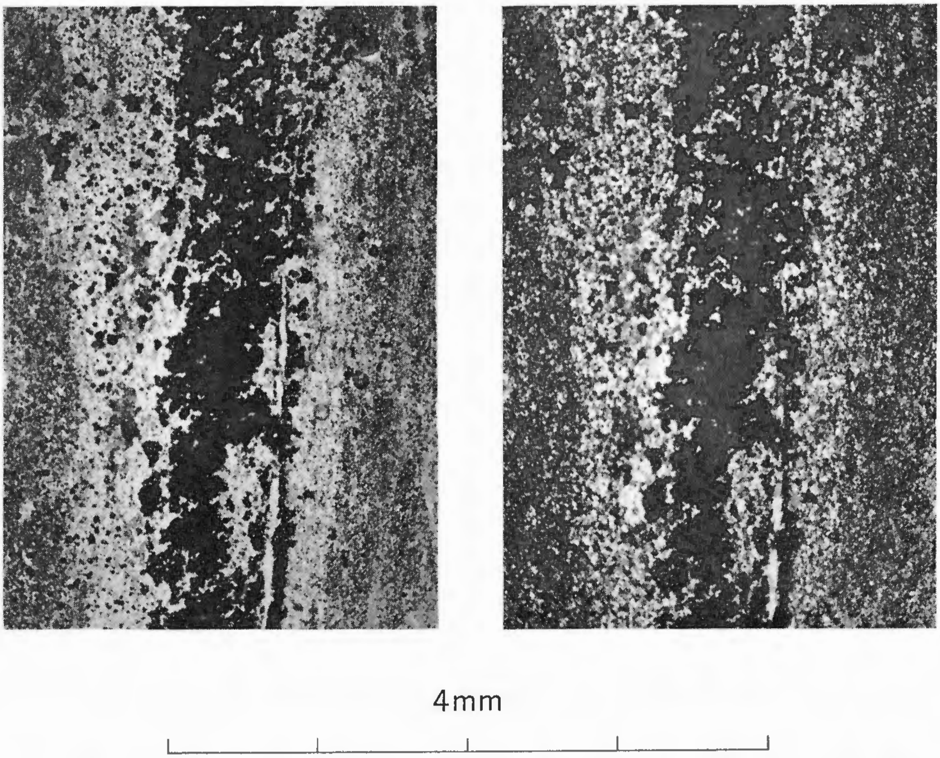


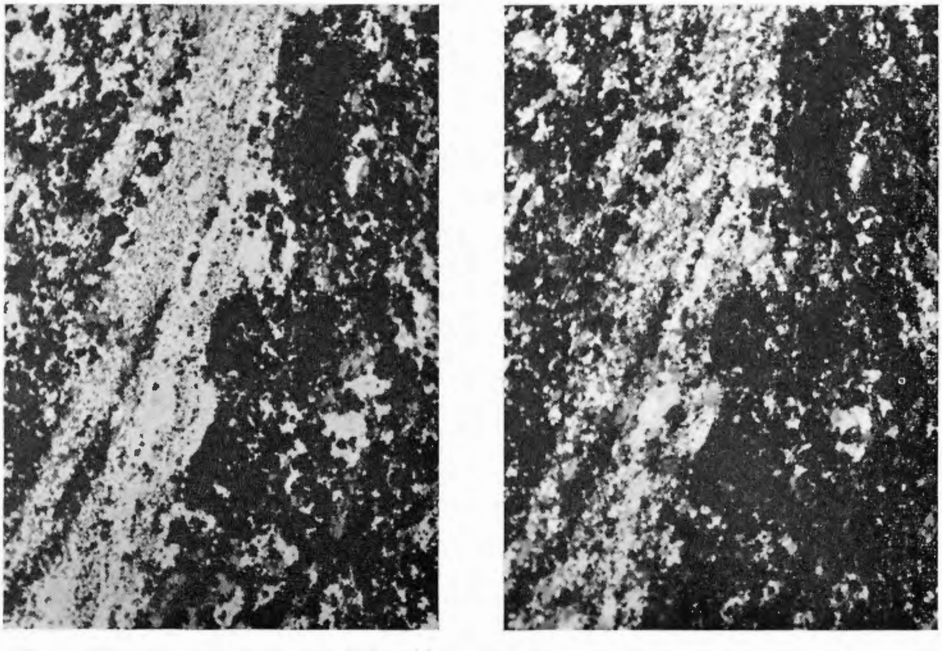
B. Upper epidote-amphibolite metamorphic facies from Mount Wright area, Quebec.

PLATE VI. TEXTURES OF ALGOMA-TYPE MAGNETITE-HEMATITE-CHERT IRON-FORMATION.

(Photomicrographs with transmitted light on left and with crossed nicols on right.)

A. Lower greenschist metamorphic facies from Timagami Lake area, Ontario.

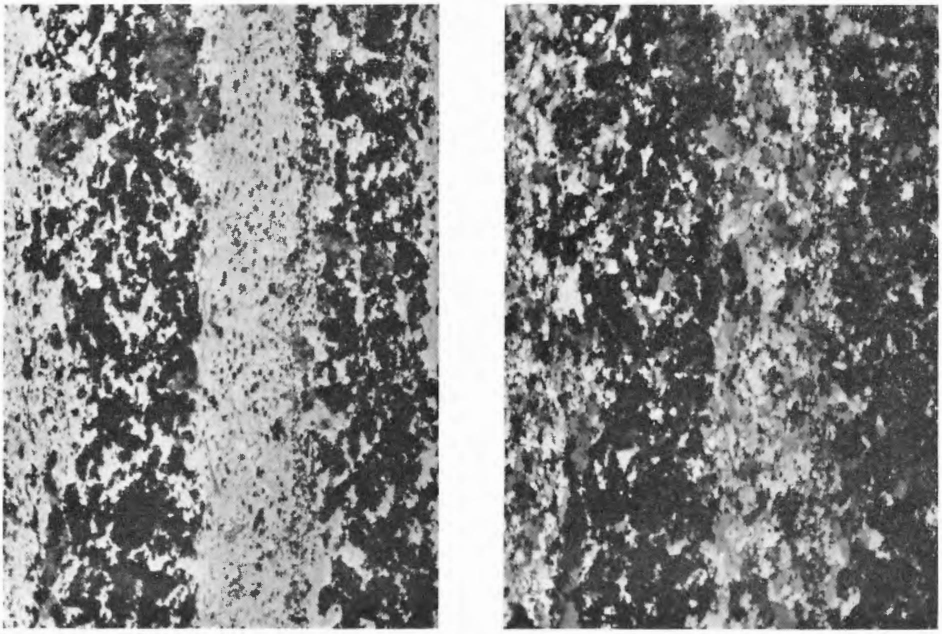


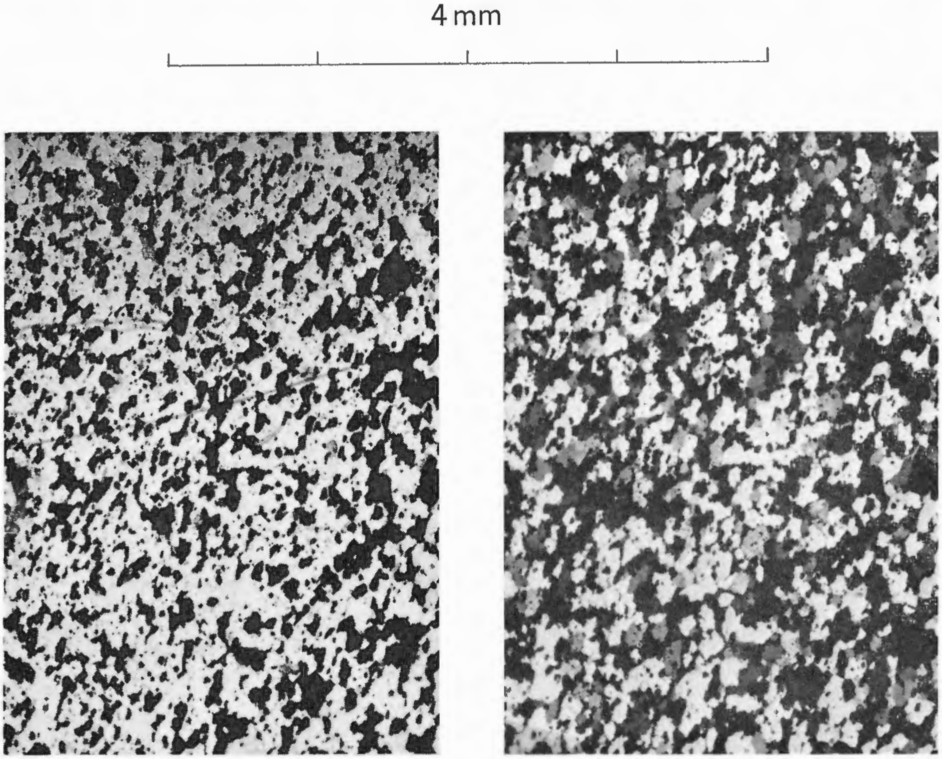


B. Greenschist metamorphic facies from Kirkland Lake area, Ontario.

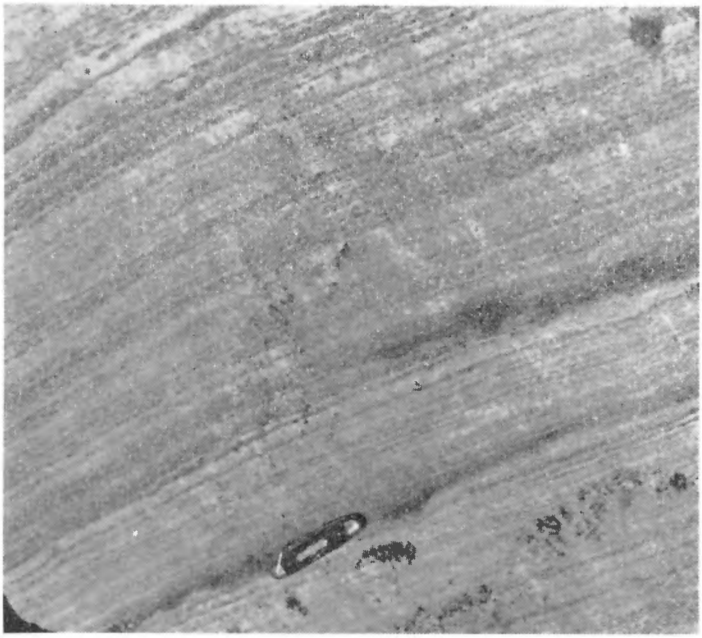
PLATE VII. TEXTURES OF ALGOMA-TYPE MAGNETITE-HEMATITE QUARTZ IRON-FORMATION.
(Photomicrographs with transmitted light on left and with crossed nicols on right.)

A. Upper greenschist metamorphic facies from Capreol area, Ontario.





B. Associated with garnet-biotite-hornblende schist from Zealand township, Ontario.



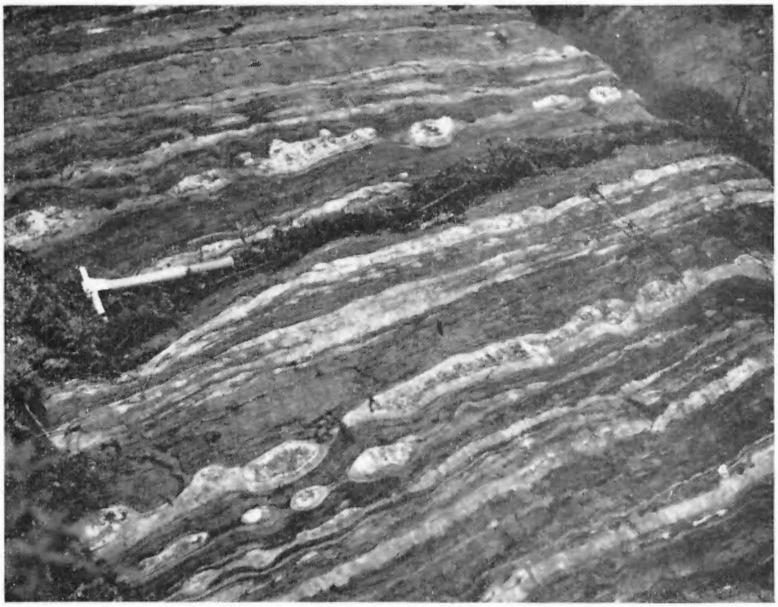
A. Highly metamorphosed hematite-quartz facies, Lac Jeannine mine, Quebec.

Koda 99

PLATE VIII. SUPERIOR-TYPE IRON-FORMATIONS

B. Highly metamorphosed hematite-magnetite-quartz facies, near Wabush Lake, Labrador, Newfoundland.

Gross, 6-1-60



A. Chert-carbonate facies of Gunflint iron-formation, east of Port Arthur, Ontario.

Gross, 4-1-62

PLATE IX. SUPERIOR-TYPE IRON-FORMATIONS

B. Iron-silicate-magnetite facies with typical lenticular bands of chert (light grey), near Lake Albanel, Quebec.

Gross, 16-7-62

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