GEOLOGICAL SURVEY OF CANADA

DEPARTMENT OF ENERGY, MINES AND RESOURCES

BULLETIN 220

THE STRATIGRAPHY AND MINERALOGY OF THE SOKOMAN FORMATION IN THE KNOB LAKE AREA, QUEBEC AND NEWFOUNDLAND

I. S. Zajac

Price $5.00

Ottawa
Canada
1974

THE STRATIGRAPHY AND MINERALOGY OF THE SOKOMAN FORMATION IN THE KNOB LAKE AREA, QUEBEC AND NEWFOUNDLAND

Technical Editor
R. G. BLACKADAR

Critical Reader
G. A. GROSS

Layout
LEONA R. MAHONEY
MARTHA J. MCLEAN

Artwork by GSC CARTOGRAPHIC UNIT

GEOLOGICAL SURVEY
OF CANADA

BULLETIN 220

THE STRATIGRAPHY AND MINERALOGY OF THE SOKOMAN FORMATION IN THE KNOB LAKE AREA, QUEBEC AND NEWFOUNDLAND

I. S. Zajac

DEPARTMENT OF
ENERGY, MINES AND RESOURCES
CANADA

Available by mail from Information Canada, Ottawa,

from the Geological Survey of Canada
601 Booth St., Ottawa

and

Information Canada bookshops in

HALIFAX
1683 Barrington Street

MONTREAL
640 St. Catherine Street West

OTTAWA
171 Slater Street

TORONTO
221 Yonge Street

WINNIPEG
393 Portage Avenue

VANCOUVER
800 Granville Street

or through your bookseller

Price: $5.00 Catalogue No. M42-220

Price subject to change without notice

Information Canada
Ottawa, 1974

PREFACE

An objective of the studies carried out by the Geological Survey of Canada is to estimate the potential abundances and probable distribution of the mineral and fuel resources available to Canada and to accomplish this, an understanding of the geological characteristics favourable to the formation of a given type of deposit is necessary.

The iron deposits of Labrador-Quebec have been known since the reconnaissance work carried out by A.P. Low in the 1890's for the Geological Survey of Canada. Exploitation of the deposits began in 1954 but the mining industry has continued geological exploration in order to extend the known reserves.

The Labrador geosyncline, a belt of Proterozoic rocks that extends 700 miles southeast from the west shore of Ungava Bay to within 200 miles of the St. Lawrence River, contains amongst the sediments in its western part one of the most extensive iron-formations in the world. The name Sokoman Formation has been applied to this and in this report the author presents new information on the stratigraphy and mineralogy of this unit within a 500-square-mile area centred about Knob Lake.

The results of this study give a better understanding of the depositional environment, the origin of the significant minerals and of Precambrian iron-formations in general.

Y.O. Fortier,
Director,
Geological Survey of Canada.

Ottawa, July 17, 1972.

CONTENTS

ILLUSTRATIONS

ABSTRACT

The Labrador Trough or geosyncline, a belt of Proterozoic rocks, contains among the sediments in its western part one of the most extensive iron-formations known.

After discussing the general geology of the Knob Lake area a detailed description of the 10 identifiable members of the Sokoman Formation (iron-formation) is given. The mineralogy of the primary (pre-alteration) minerals of the iron-formation is discussed and chemical analysis presented. The origin of the formation is discussed at length and the characteristics of the various members are used to establish a sequence of depositional events.

RÉSUMÉ

Le fossé ou géosynclinal du Labrador, qui est une ceinture de roches protérozoïques, comprend parmi les sédiments de sa partie occidentale une des plus grandes formations ferrifères connues.

Après avoir étudié la géologie générale de la région de Knob Lake, l'auteur fait une description détaillée de 10 termes identifiables de la formation de Sokoman (formation ferrifère). La minéralogie des minéraux primaires (avant altération) de la formation ferrifère est aussi étudiée et l'auteur en présente une analyse chimique. Il étudie en détail l'origine de la formation et il utilise les caractéristiques des différents termes pour établir une série des phénomènes qui ont présidé à la déposition.

CHAPTER I

INTRODUCTION

SCOPE AND PURPOSE

The belt of Proterozoic rocks known as the Labrador Trough contains, among the sediments in its western part, one of the most extensive iron-formations in the world. This study deals with the stratigraphy and mineralogy of this iron-formation within an area of approximately 500 square miles centred about the town of Schefferville and the adjacent Knob Lake (Fig. 1).

The purpose of the study is to present new information on the stratigraphy and mineralogy of the iron-formation in the Knob Lake area with the objective of contributing to a better understanding of its depositional environment, the origin of its minerals, and the origin of Precambrian iron-formations in general.

HISTORY AND GEOLOGICAL INVESTIGATIONS

Jacques Cartier, who was the first to explore the Gulf of St. Lawrence in 1534, was not very impressed with the country on its northern shores. He described it as a land "composed of stones and horrible rugged rocks," and remarked that very likely it was "the land God gave to Cain" (Williams, 1963). Settlers, traders and explorers of the next two and a half centuries were equally disenchanted with the "New Land," and the vast areas of Labrador remained unknown. Erland Erlandson, a trader of the Hudson's Bay Company, was the first European to cross the centre of Labrador in 1834 (Davies, 1963). He returned to the central highland two years later and established a trading post, Fort Nascopie (15 miles southeast of Knob Lake) which survived in this nearly barren and isolated wilderness for 31 years.

The earliest report of iron ore in Labrador came from the Jesuit missionary, Pierre Babel, who discovered it on one of his trips to the interior of the peninsula in the years 1866 to 1870. The first geological exploration of Labrador was made by the legendary A. P. Low (1896) of the Geological Survey of Canada. He obtained a broad outline of the rocks now known as the Labrador Trough, recorded the occurrence of iron-formation, and pointed out its potential as a source of iron ore. Despite Low's reports, Labrador attracted little attention at the time and was soon forgotten once again.

Direct-shipping iron ore was discovered just west of Knob Lake in 1929 by J. E. Gill, and in 1936 the Labrador Mining and Exploration Company began active exploration of the Labrador Trough.[1] The activity subsided at the beginning of World War II but resumed in 1942. In the years that followed,

[1] Exploration and development of iron ore areas in the Labrador Trough is described in detail by Gustafson and Moss (1953), and by the Iron Ore Company of Canada Staff (1955).

Original manuscript submitted: May 1972
Final version approved for publication: March 1973
Author's address: 2550 Saulsbury Street
Lakewood, Colo. 80215

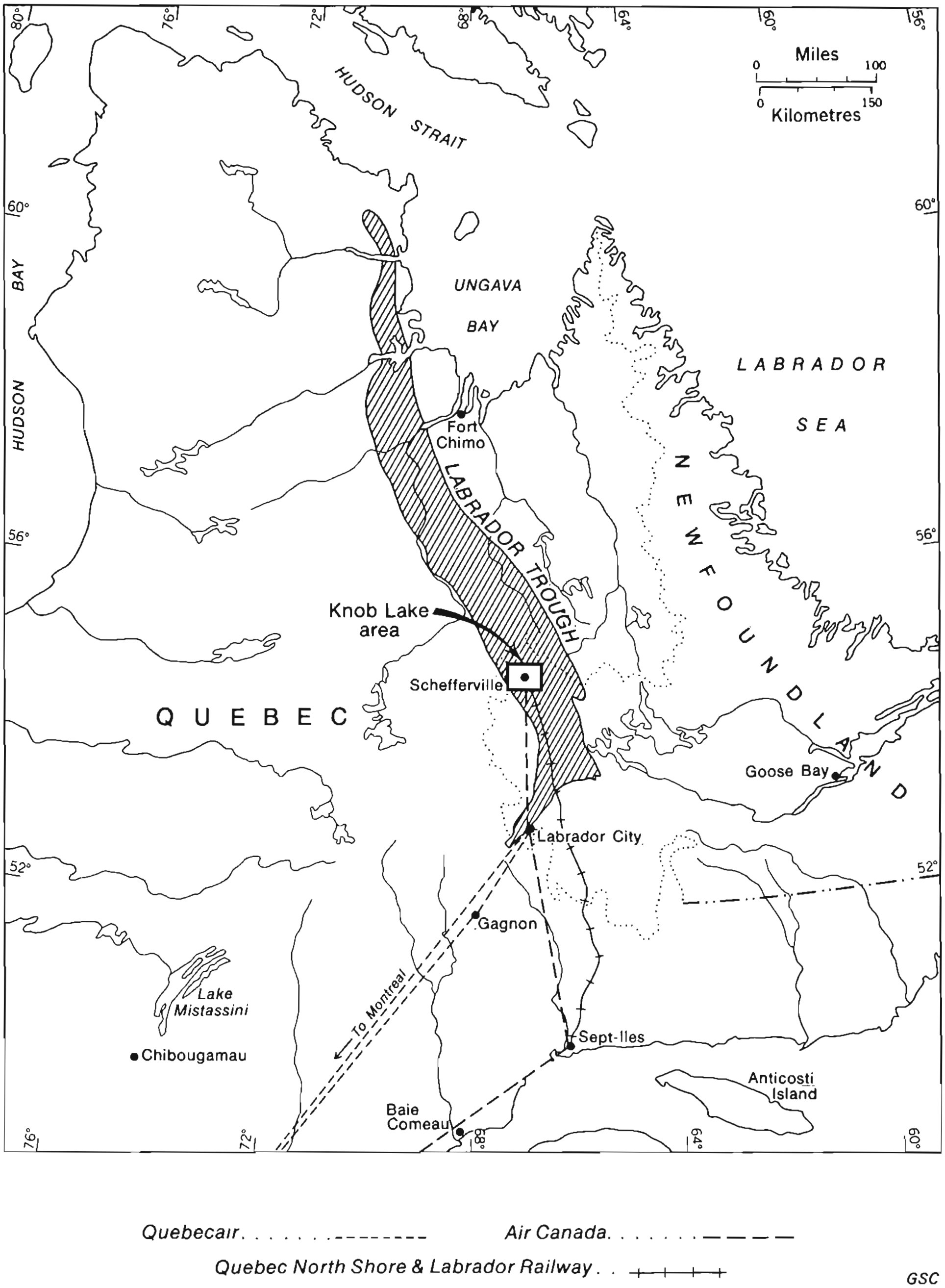


Figure 1. Location and accessibility of Knob Lake area.

investigations by the same company and by the Hollinger North Shore Exploration Company resulted in the discovery of 12 iron ore deposits. In 1949 it was estimated that 400 million tons of direct-shipping ore were needed to justify the 200 to 300 million dollars required to mine and ship the ore from this remote area. By 1950 this goal was exceeded by nearly 20 million tons.

In 1949 the Hollinger North Shore Exploration Co. Ltd. and the Labrador Mining and Exploration Co., in association with five American steel companies, formed the Iron Ore Company of Canada Ltd. to develop the proven ore reserves. In less than four years, port facilities in Sept Isles were constructed; 357 miles of railway to Knob Lake were completed; a new townsite (Schefferville) was built; a hydroelectric plant was established; and several orebodies were prepared for production. The first ore was delivered to Sept Isles in 1954, and since then the mines in the Knob Lake area have produced from 7 to 13 million tons annually.

The accomplishments of the company geologists are no less imposing. By 1952 nearly 15,000 square miles were mapped - over one-fifth of which was on a scale of 1 inch to 1,000 feet or less - and major stratigraphic and structural elements delineated. Geological investigations by company staff continued and much of what is known about the geology of the Labrador Trough today, particularly about the geology of the deposits of direct-shipping iron ore, which were the main targets of the investigations, can be attributed to their efforts.

The Geological Survey of Canada and the Quebec Department of Natural Resources have also made important contributions to Labrador geology. The work of the geologists of the Geological Survey of Canada is summarized by Frarey and Duffell (1964). The recent reports by Baragar (1967) and Gross (1968) are valuable additions. Publications of the Quebec Department of Natural Resources deal mainly with the northern part of the Labrador Trough.

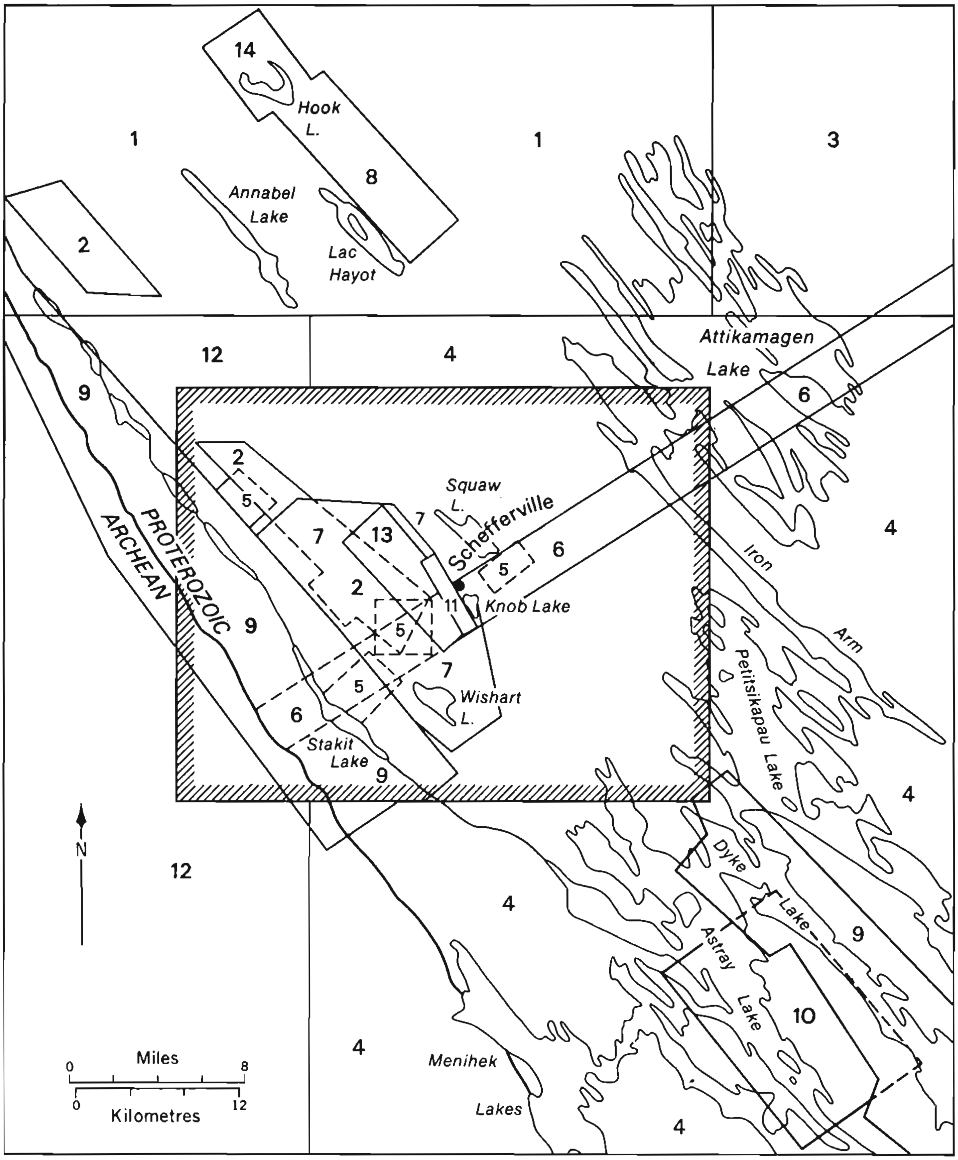
All maps and reports which deal with specific parts of the Knob Lake area and vicinity and to which reference is made in this report are indicated on Figure 2.

The writer's work in the Knob Lake area extends over a period of five and a half years. Nearly five of these years were in the employ of the Iron Ore Company of Canada. The field seasons of 1965 and 1966 were under the auspices of the Geological Survey of Canada and were devoted almost entirely to unraveling the stratigraphy of the iron-formation in the area.

ACKNOWLEDGMENTS

The writer is grateful to the Geological Survey of Canada and to the Iron Ore Company of Canada, Ltd., for making the present study of the iron-formation possible. The Company provided free access to the mine areas, made its geological records available for study and extended many courtesies in the field. The Survey fully supported the field work done during the summers of 1965 and 1966, and supplied most of the chemical analyses, polished sections and thin sections used in this study.

The writer owes special thanks to Dr. G.A. Gross of the Geological Survey of Canada for his support of the project, encouragement and numerous helpful suggestions.



1. Baragar (1963,67)
2. Dufresne (1952)
3. Frarey (1952)
4. Frarey (1961)
5. Gross (1951)
6. Harrison (1952)
7. Howell (1954)
8. Kirkland (1950)
9. Perrault (1955)
10. Sauve (1953)
11. Seguin (1961)
12. Stevenson (1962)
13. Schwellnus (1957)
14. Neal (1949)

GSC

Figure 2. Geological mapping and investigations in the Knob Lake area (shaded) and vicinity.

The laboratory study was done at the University of Michigan. The writer is indebted for guidance and help in various phases of the work to Professors E.W. Heinrich, D.R. Peacor and W. Bigelow, and particularly to Professors W.C. Kelly, L.I. Briggs and F.S. Turneaure who were burdened with most of the writer's problems. Their interest and excellent advice are gratefully acknowledged.

The dissertation was completed in 1970-1971 when the writer was employed by the Hanna Mining Company and Mine Finders, Incorporated. Their generous support is very much appreciated. Messrs. D. Mahling and D. Dion gave able assistance with the field work.

Financial assistance for the study at the University of Michigan was received from the National Research Council of Canada. The analyses of greenalite and crocidolite were paid for by the Rackham Graduate Students Fund of the University of Michigan.

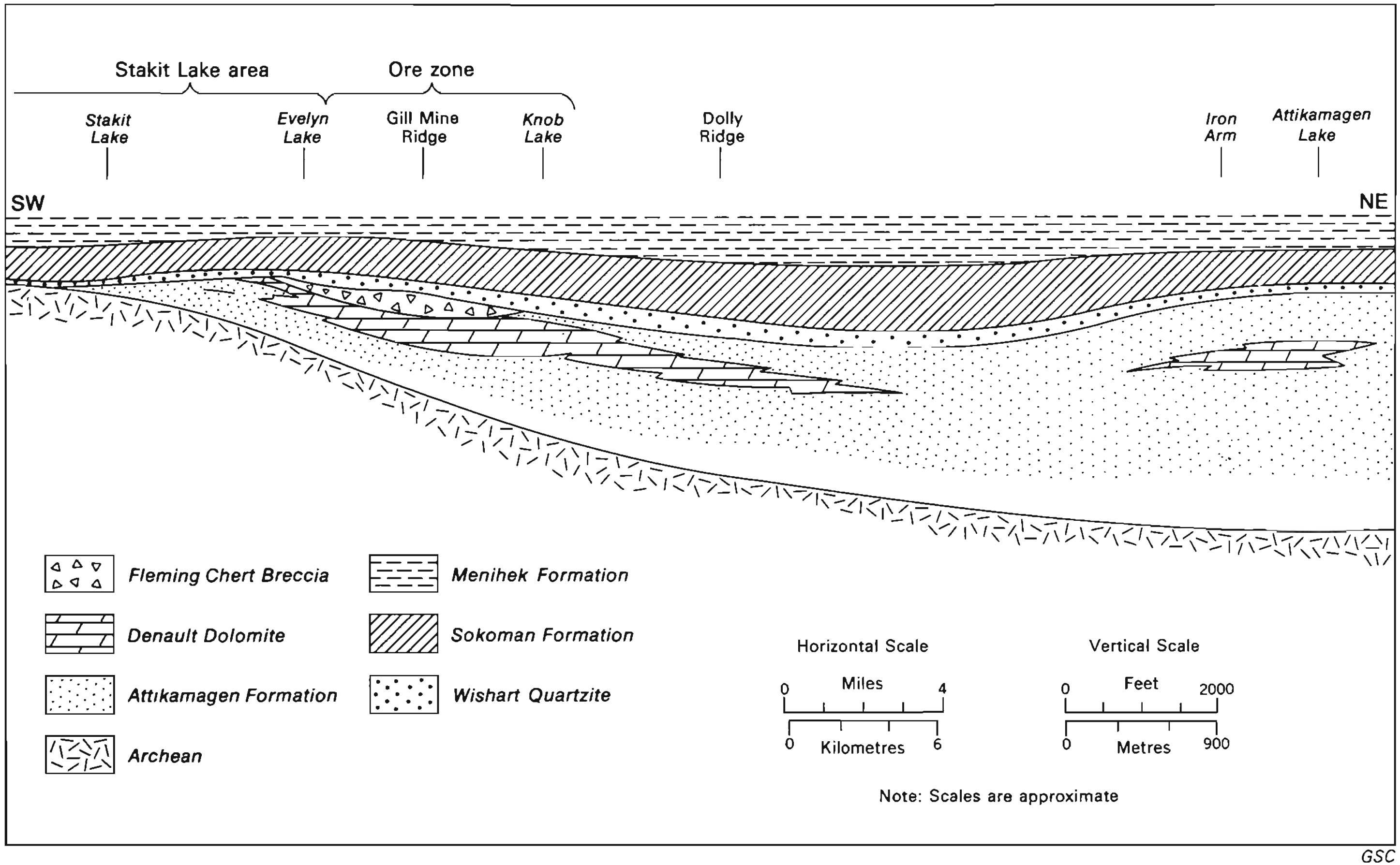


Figure 3. Cross-section through the central part of the Knob Lake area before deformation.

CHAPTER II

GENERAL GEOLOGY

The Labrador Trough, a belt of Proterozoic rocks, extends northwest for nearly 700 miles across the centre of the Labrador Peninsula. It is approximately 60 miles wide near Knob Lake but narrows to the north and south.

The eastern half of the Labrador Trough consists mainly of basic volcanic and intrusive rocks, and the western half, almost entirely of sediments. In the west, the practically unmetamorphosed sediments rest unconformably on the Archean basement. In the east, the rocks become progressively more metamorphosed and merge almost imperceptibly with older schists and gneisses.

STRATIGRAPHY

The Knob Lake area lies in the western, dominantly sedimentary and least metamorphosed part of the Labrador Trough. With the exception of the Archean rocks that appear in the southwestern corner of the map, the area is underlain wholly by the sediments and minor volcanic and intrusive rocks which comprise the Knob Lake Group.

Archean

The Archean rocks exposed below the unconformity along the western margin of the Labrador Trough are distinctly banded quartz-feldspar-biotite gneisses, and small granitoid bodies which range in composition from granite to quartz diorite (Stevenson, 1962; Baragar, 1963; Frarey, 1961). Potassium-argon dating of the rocks shows them to be 2365 to 2505 million years old (Lowden, 1960, 1961).

Attikamagen Formation

The Attikamagen Formation is the oldest known Proterozoic formation in the area. Its lower contacts are not exposed, however, and it is not known whether it lies directly on the Archean basement or is separated from the basement by the quartzites, arkoses and greywackes of the Seward Formation which is present both to the north (Baragar, 1967) and to the south (Frarey and Duffell, 1964) of the Knob Lake area.

The Attikamagen Formation is at least 100 feet thick in the Stakit Lake area but thins out westward so that the overlying Wishart Quartzite rests directly on the Archean (Fig. 3). The Attikamagen Formation becomes thicker to the east and may be over 2,000 feet thick in the central part of the Knob Lake area.

Within the map-area, the Attikamagen Formation consists mainly of thin-bedded to laminated, fine-grained, commonly cherty, silty or dolomitic, greenish grey shales and slates. Black carbonaceous shales and slates are common immediately north and east of Knob Lake, and reddish grey shales, east and southeast of lac de la Squaw. Beds and lenses of chert, siltstone, quartzite and dolomite are present also, but collectively constitute only a

TABLE OF FORMATIONS
KNOB LAKE AREA

ERA	SUPER GROUP	GROUP	FORMATION	APPROXIMATE THICKNESS (in feet)	DOMINANT LITHOLOGY
MESOZOIC			Redmond		Rubble, clay
UNCONFORMITY					
PROTEROZOIC					Diabase
	INTRUSIVE CONTACT				
	KANIAPISKAU	KNOB LAKE	Menihek	1000+	Shale, slate
			Sokoman	300-700+	Iron-formation
			Nimish	0-300	Greenstones
			Local (?) Unconformity		
			Wishart	60 - 200	Quartzite
			Local Unconformity		
			Fleming	0-400	Chert breccia
			Denault	0-600	Dolomite
			Attikamagen	0-1000+	Shale, slate
UNCONFORMITY					
ARCHEAN					Gneisses

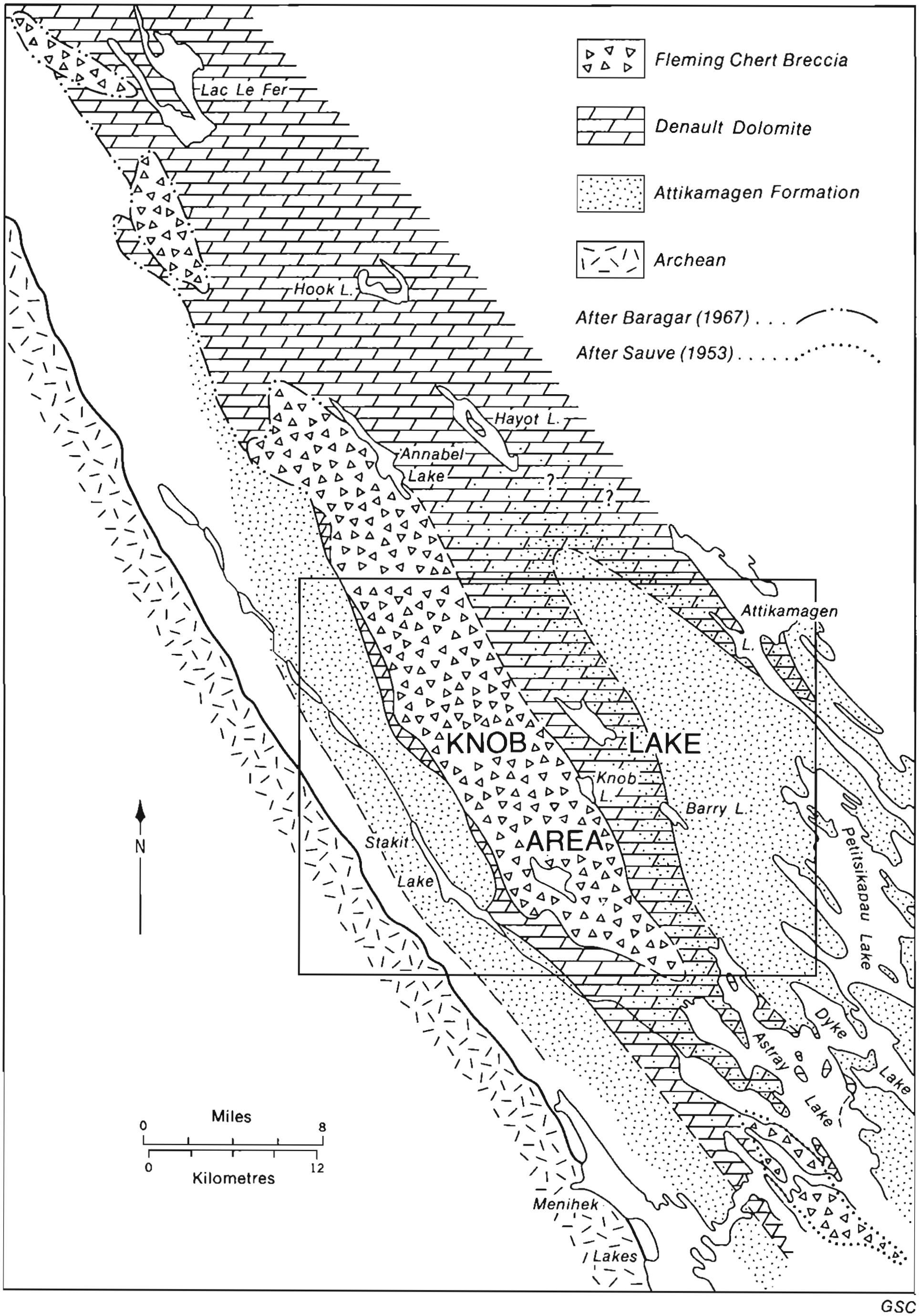


Figure 4. Distribution of Attikamagen, Denault and Fleming Formations.

small part of the formation. Quartzite and dolomite interbeds are most common in the Stakit Lake area. Basic volcanics are interbedded with Attikamagen rocks in the eastern part of the Trough (Frarey, 1961; Baragar, 1963; Donaldson, 1966) and in a few places near Dyke Lake (Perrault, 1955). No volcanic rocks are known in the Knob Lake area west of Montreal Bay, but a few of the coarser argillaceous beds may be tuffaceous.

The Attikamagen and Denault Formations are interbedded as shown in Figure 3. In the west, the interfingering of Attikamagen and Denault rocks is seen in numerous outcrops along the contact between these two formations. West of Evelyn Lake, the Denault Dolomite becomes increasingly more interbedded with the Attikamagen Formation and disappears among the shales within a few miles of the present margin of the Labrador Trough. In the central part of the area the interbedding of the two formations is most convincingly established by following the shales as a continuously thickening unit above the Denault Dolomite, from Annabel Lake southeast to Barry Lake (Fig. 4), where the dolomite disappears and the overlying and underlying shales and slates merge imperceptibly with one another. Similarly, a thin shale horizon above the Denault Dolomite near Abel Lake can be traced along Gilling Lake southeast into the eastern part of Astray Lake area, where the shales and slates above the dolomite are 100 to 300 feet thick. Here, likewise, the lower and upper part of the Attikamagen Formation join to form one single unit once the intervening dolomite pinches out.

The extent to which the upper part of the Attikamagen Formation overlaps the Denault Dolomite is shown in Figure 4. The northern limits of this overlap are not certain, but evidently it does not extend north of Annabel and Hayot Lakes, since in most parts of the Wakuach Lake area (Baragar, 1967) the Denault Dolomite is directly overlain by, and in places interbedded with, the Wishart Quartzite.

Denault Dolomite

This formation is mainly a light grey, buff-weathering, fine-grained dolomite. Chert is common and occurs as intimate mixtures with dolomite; as nodules, thin conformable beds and lenses; as fracture fillings; and as replacement veinlets.

Much of the dolomite is thick bedded and massive with minor, erratically distributed dolomite sands, intraformational conglomerates and breccias. Sedimentary textures and structures such as grain gradations, scour and fill features, crossbedding, and rare ripple-marks, are present in a few places.

Chemically the carbonate is close to pure dolomite. Howell (1954), on the basis of optical determinations, identified appreciable quantities of ankerite, but the 10 available chemical analyses (Dufresne, 1952; Stubbins et al., 1961; Donaldson, 1966), which show 2.91 to 0.38 per cent combined FeO-Fe_2O_3 and less than 0.2 per cent MnO, do not support his data. The silica content of the formation varies between 0.5 and 37.91 per cent, with an average figure for the 10 analyses of 17.11 per cent.

The Denault Formation is 200 to 600 feet thick in the ore zone but thins out both east and west (Fig. 3). It is absent in the eastern part of the Knob Lake area (Fig. 4). Where it appears again, 35 miles east of Knob Lake, it is more than 3,000 feet thick (Donaldson, 1966). From Knob Lake the formation apparently thickens northward and becomes just over 4,000 feet thick in the Wakuach Lake area (Baragar, 1967).

The stratigraphy and lithology of the Denault and Attikamagen Formations reveal the first broad outlines of the depositional environment in the south-central miogeosynclinal part of the Labrador Trough. Denault Dolomite indicates the shallower and nearshore high energy areas of the basin, whereas the shales denote the deeper offshore areas and also marginal areas of the basin unaffected by strong currents or wave action. (See Table 1 and Figs. 3 and 4).

Fleming Chert Breccia

The present distribution of the Fleming Formation is shown in Figure 4. With the possible exception of the Marion Lake area (Donaldson, 1966), 35 miles east of Knob Lake, these are the only occurrences of this peculiar rock in the Labrador Trough.

The main body of Fleming Chert Breccia is approximately 42 miles long, 5 miles wide (7 to 9 miles before deformation) and is 300 to just over 400 feet thick in its central part. Wherever its lower contact is seen in the Knob Lake area, it grades sharply into cherty and sandy shales or rarely into a silty and shaly dolomite. Its contact with the overlying Wishart Quartzite is usually sharp, and locally, as north and east of Elizabeth Lake, marked by a well-sorted and rounded chert-pebble conglomerate.

The breccia, which makes up 90 to 95 per cent of the formation, is a dense, massive rock with little evidence of bedding. It consists of chert fragments which are cemented by chert, quartzite, or various mixtures of the two. Bedded chert, lenses of quartzite, rare slabs and lenses of dolomite, and thin beds of shale are the other minor constituents of the formation.

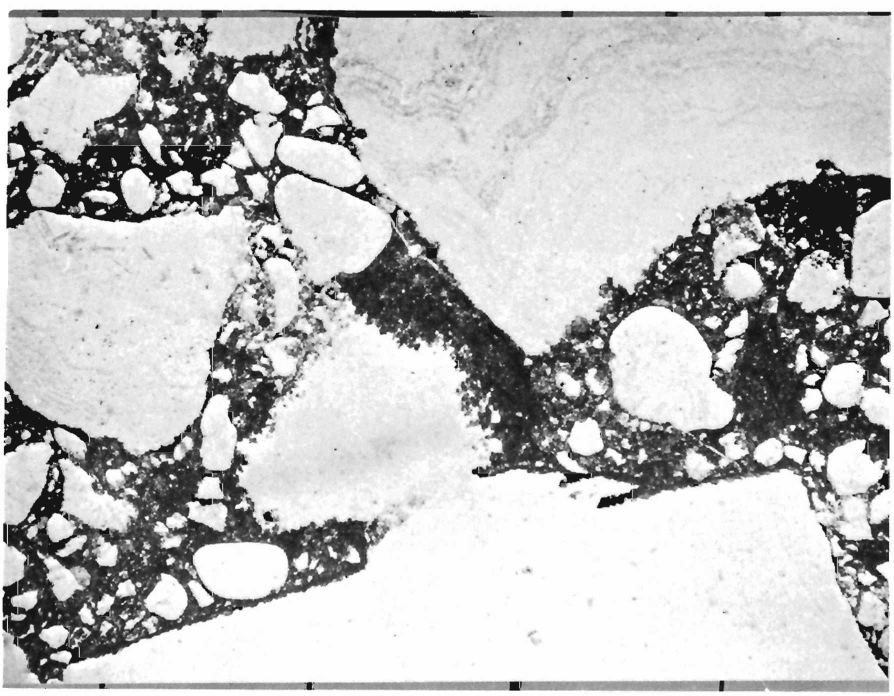


Figure 5. Typical texture of Fleming Chert Breccia – angular fragments of chert in a cherty, ankeritic (?) quartzite matrix. (Carbonate is not common in Fleming Chert Breccia.)

Table 1. Depositional environment of Attikamagen and Denault Formations

	Stakit Area	Ore Zone	Knob Lake-Dolly Ridge	Dolly Ridge-Iron Arm	Iron Arm Attikamagen Lake
Depth	shallow	shallow	intermediate	deep	intermediate
Energy	low	high	intermediate	low	intermediate
Dominant lithology	shale	dolomite	dolomite; shale	shale	shale; dolomite

Figure 6. Same as above with crossed nichols. Note the massive and banded-crustiform types of chert.

Chert cement of the breccia predominates in the lower half and quartzite in the upper half of the formation in the eastern parts of the area (north and east of Ruth Lake and east of lac de l'Hématite). Twenty-five feet of chert-cemented breccia are also present near George V Lake (Perrault, 1955). In other parts of the area there appears to be little difference between the top and bottom of the formation, most of the matrix being medium- to coarse-grained quartzite. The predominantly well-rounded quartz grains of the matrix are surrounded by chert and less commonly by fine-grained argillaceous material. Rare dolomite (and ankerite?), where present, is seen to replace the chert fragments and some of their cement (Fig. 5).

Most fragments of the breccia are sharply angular, a fraction of an inch to several inches in size, and commonly make up 30 to 60 per cent of the rock. They are mainly of two types, massive microcrystalline chert, and variously banded chert and quartz. The latter fragments are wavy bedded, laminated, and in places colloform. Many show all gradations from very fine grained, granular, lamellar or radiating chert to medium- or, rarely, coarse-grained quartz. The quartz is locally crustiform and well zoned. All bands and laminae end abruptly at the edges of the fragments, indicating that the banding as well as the cherts existed before the breccia was formed.

The origin of the breccia is not clear. It is generally agreed that the deposit is an intraformational breccia, but the origin of the chert and the manner in which the brecciation occurred are still uncertain.

Wishart Quartzite

The Wishart Quartzite is a blanket type of deposit which covers successively, from west to east, the Archean basement, the Attikamagen Formation, the Denault Dolomite, the Fleming Chert Breccia and then again the Attikamagen Formation (Fig. 3).

In the Knob Lake area the formation is mainly a light grey, massive- to well-bedded, medium- to coarse-grained quartzite, which ranges in composition from very pure orthoquartzite to argillaceous quartzite with 5 to 10 per cent of argillaceous cement. Other rock types in decreasing order of abundance are: feldspathic quartzite, arkose, feldspathic greywacke, siltstone, shale, dolomite (ankeritic in places) and conglomerate. The medium- to coarse-grained quartzites are composed of very well rounded and in most places well-sorted quartz grains cemented by authigenic, optically continuous quartz, and less commonly by chert. The argillaceous cement of the majority of quartzites is a fine-grained mixture of mica and chlorite with other unidentified clay-size constituents.

The feldspathic rocks range from very coarse grained arkoses to fine-grained, thin-bedded feldspathic greywackes. The coarser varieties, most common in the upper part of the formation, are texturally similar to the previously described quartzites. Some are also very pure and could be classified as feldspathic orthoquartzites. The fine-grained types, most common in the middle and lower part of the formation in the eastern part of the area, are composed of angular, to sharply angular, quartz and feldspar exposed by abundant argillaceous cement. Rock fragments, which make up 5 to 15 per cent of the greywackes, are mainly chert. The characteristic feldspar of all quartzites and arkoses is microcline, but untwinned orthoclase and perthite also appear in some of them. Plagioclase (albite to andesine), which is rarely present in the medium- and coarse-grained rocks, makes up one-fourth to one-half of the feldspars in the greywackes. It is well twinned, angular, and as most other feldspars, little altered.

Shale is common only near the lower contact of the formation in the vicinity of Stakit Lake and in the northeastern and southeastern parts of the Knob Lake area. Near Attikamagen Lake the shale is closely associated with abundant light greenish grey siltstone and the similar rocks of the Attikamagen Formation.

Figure 7. Quartzite cobble and boulder conglomerate at the top of Wishart Formation, Gill Mine Ridge.

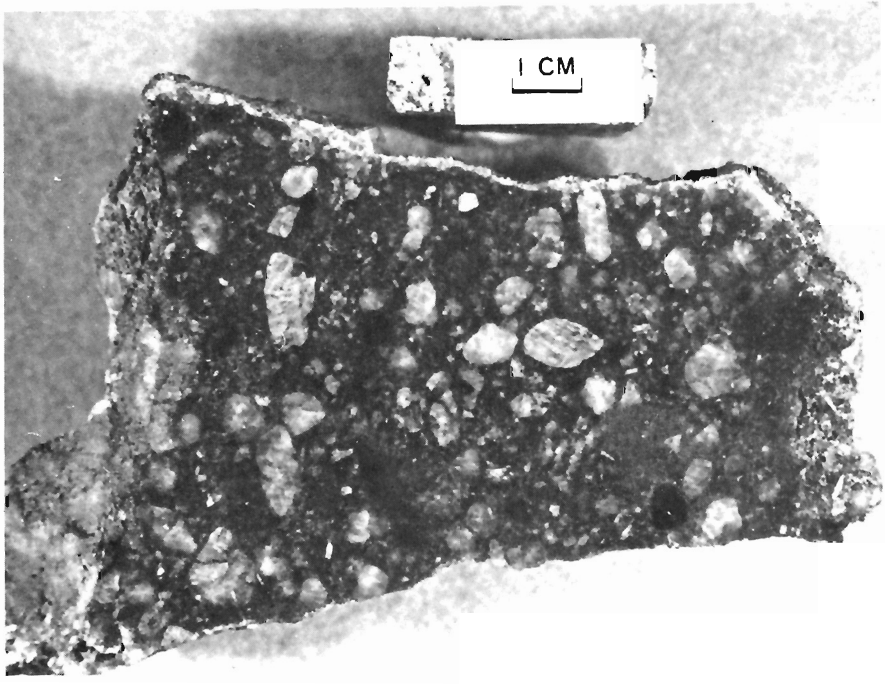


Figure 8. Quartz-feldspar pebble conglomerate at the top of Wishart Formation, Hope Lake.

Brown to dark rusty weathering carbonate, some of which is probably ankerite,[1] forms very few, widely separated but conspicuous, lenses in the lower part of the formation. These lenses are present west of Fleming Lake where the Wishart Formation lies on the Archean basement; in several places above the Fleming Chert Breccia in the ore zone (Harrison, 1952; Howell, 1954); and at three localities northwest of Dyke Lake. A 30-foot-thick lens of ankeritic (?) dolomite on the ridge between Petitsikapau Lake and Dyke Lake is notable for the only well-formed stromatolites in the area.

The conglomerates are most numerous in the lower and upper part of the Wishart Quartzite, particularly in the western part of the area. A few thin, discontinuous beds of pebble conglomerates are present locally within the lower 30 feet of the formation, mainly where it overlies the Fleming Chert Breccia. Similarly located conglomerates are also present near the western margin of the Labrador Trough (Harrison, 1952; Perrault, 1955). The upper conglomerates occur along the very sharp contact between the Wishart Quartzite and the overlying sediments of the iron-formation. They range from boulder conglomerates to pebbly quartzites and are composed mainly of pebbles, cobbles and boulders of quartzite embedded in a very similar quartzite matrix (Fig. 7). Conglomeratic quartzites with various amounts of chert, quartz and feldspar pebbles (Fig. 8) are less common. Where the conglomerates are covered by a thick layer of chert, some of the boulders and pebbles are entirely surrounded by chert and appear to be suspended in it. The conglomerates at the top of the Wishart Quartzite are confined mainly to the western part of the area. The predominant rocks at the same stratigraphic horizon east of Knob Lake are medium- to coarse-grained orthoquartzites. The presence of conglomerates and orthoquartzites at the top of the formation indicates a very shallow, high energy depositional environment throughout most of the Knob Lake area during the last stage of Wishart sedimentation.

The conglomerates and possible unconformities are not always equally emphasized or similarly interpreted. The lower conglomerates, as noted by Sauve (1953) and Baragar (1967) are probably minor and local erosional breaks which formed in the shallower parts of the basin near its western shores. A major unconformity at the base of the Wishart Quartzite is unlikely, because in most places where the Wishart Quartzite directly succeeds the Attikamagen Formation the contact between them is gradational and shows no evidence of a break in sedimentation. Local interbedding between Denault and Wishart rocks (Harrison, 1952; Gross, 1968; Baragar, 1967) is additional evidence against a major unconformity at the base of the formation.

A major unconformity at the top of the Wishart Quartzite is strongly suggested by the coincidence of the upper conglomerate horizon with an abrupt change from coarse clastic to chemical and fine clastic sediments (conglomerate and orthoquartzite to chert and shale). However, as pointed out by Gross and Baragar (pers. comm.), the quartzite is a very continuous formation which does not show any evidence of complete or very deep erosion. The thinness of the upper conglomerates and their monolithic composition also argue against prolonged erosion of the formation. It may be that even this well-defined unconformity does not represent a major time break, but is merely another local, nearshore unconformity formed by an extensive but short-lived regression of the area.

---

[1] Calculations, based on one chemical analysis of a composite sample from the locality north of Elizabeth Lake described by Howell (1954), show that the carbonate minerals contain 14.50 per cent FeO.

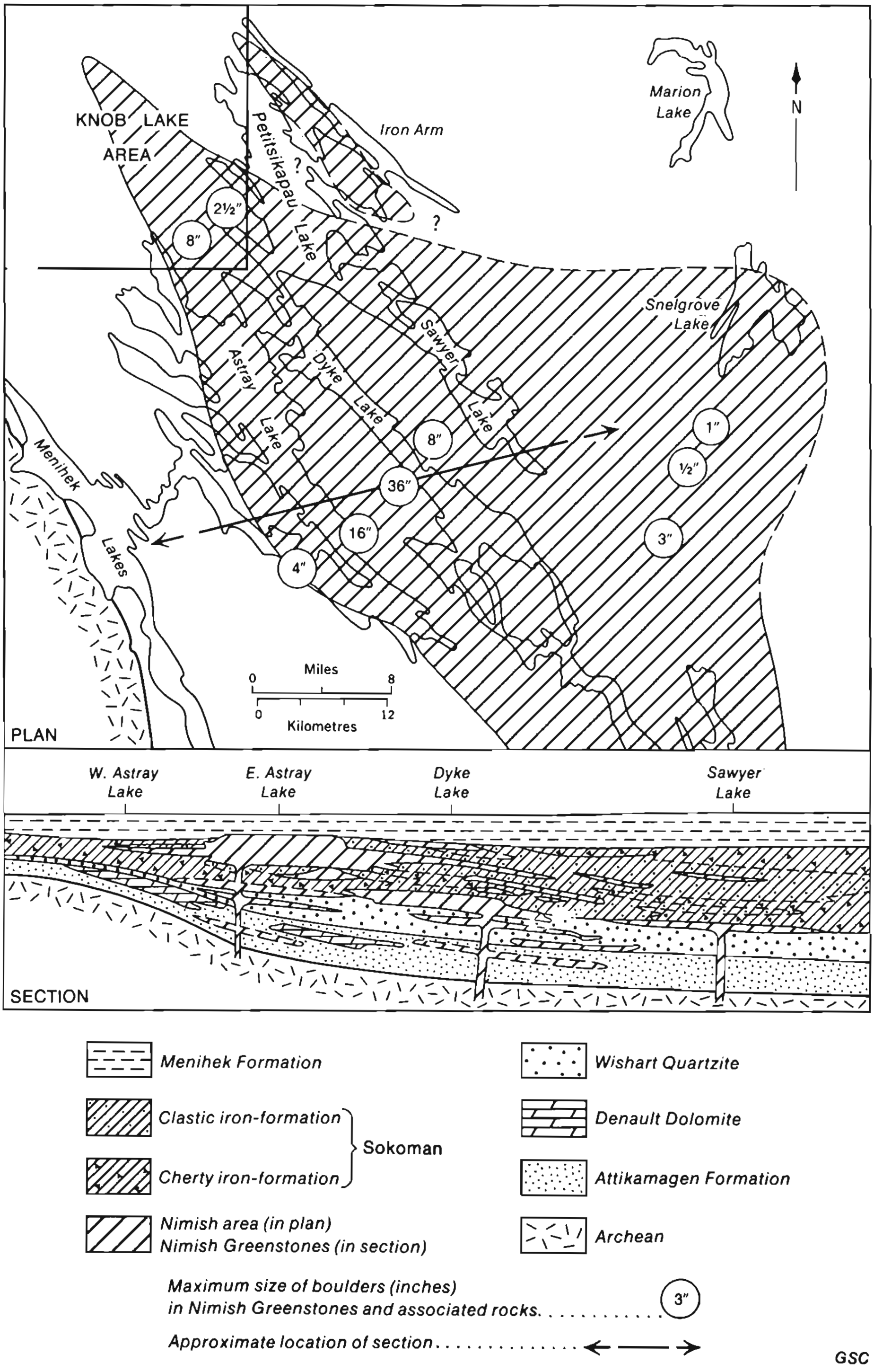


Figure 9. Distribution of Nimish Greenstones. Adapted from Sauve (1953).

Nimish Greenstones

Igneous activity in the Labrador geosyncline was confined almost entirely to its eastern half. During a part of its history, however, a centre of such activity also appeared near the western shores and is now represented in a well-defined area near Dyke Lake. This area as outlined in Figure 9 is here referred to as the "Nimish Area" and the variously altered basic volcanic and intrusive rocks that outcrop within it are collectively designated "Nimish Greenstones".

The writer is familiar with only small parts of this area and much of what is presented here is based on the work of Sauve (1953) and Perrault (1955).

Volcanic Rocks

Lavas and pyroclastics are most abundant in the immediate vicinity of Dyke Lake and Astray Lake where they are over 600 feet thick (Sauve, 1953). They are most closely interbedded with the iron-formation, but a few tuffs and lava flows also appear in the Wishart and Attikamagen formations (Perrault, 1955). None are known in the Menihek Formation.

In the central part of the Nimish area, described by Sauve (1953) and Perrault (1955), the lavas are fine grained, rarely porphyritic, greenish grey rocks composed of up to 50 per cent of plagioclase and a little augite. The groundmass is chlorite, minor sericite, sphene, leucoxene and various amounts of carbonate. Magnetite is locally common and a few flows are strongly magnetic (Sauve, 1953). The plagioclase ranges in composition from albite to andesine and is in many places replaced by a mixture of chlorite, sericite, carbonate, and locally by potassium feldspar. In some lavas orthoclase "is common as phenocrysts and may occur as laths in the groundmass with apparently almost complete exclusion of plagioclase" (Sauve, 1953).

In outcrop, the lavas are massive, pillowed, and less commonly amygdaloidal or vesicular. Pillowed lavas, predominant near Dyke Lake and Petitsikapau Lake, suggest a more frequent subaqueous deposition in the eastern parts of the Nimish area.

The pyroclastic rocks, which are interbedded with lavas at various stratigraphic levels, are chloritic tuffs, conglomerates, agglomerates and breccias. All gradations between sandy conglomerates and breccias occur. In addition to various volcanic and greenstone fragments, these rocks locally contain abundant sand- to boulder-size fragments of chert, jasper, iron-formation, and quartzite. Distribution of the maximum size of boulders in the Nimish greenstones and associated rocks, as noted by Sauve (1953), points to the Dyke-Astray area as the site of the most frequent uplift and volcanism. (See Fig. 9.)

A few unusual crystal tuffs appear among the lavas of Astray Lake. Large orthoclase laths and fragments of large crystals which are in a trachytoid arrangement within a predominantly chloritic groundmass make up from 30 to 40 per cent of the tuffs (Sauve, 1953).

In the Knob Lake part of the Nimish area the greenstones are mainly pyroclastic rocks. Near Petitsikapau Lake the greenstones are not more than 100 to 300 feet thick and are most abundant immediately on top or a few feet above the Wishart Quartzite. To the west, only two agglomerate and lava horizons are present at approximately 250 and 400 feet above the quartzite.

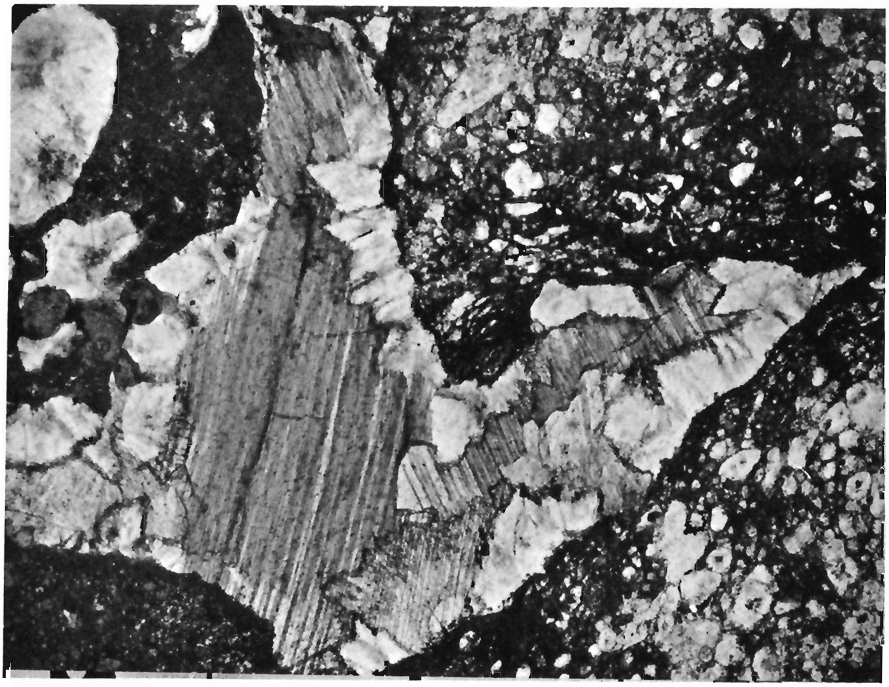


Figure 10. Texture of adularia (white to light grey) in thin section of lapilli tuff from the Nimish area (northwest of Dyke Lake). Calcite is grey and chlorite is dark grey to black.

Mineralogically and texturally the lavas are similar to those of the Dyke-Astray area. Most are pillowed, amygdaloidal and composed essentially of feldspar, chlorite and calcite. The main metallic constituent is ilmenite and less commonly pyrite or magnetite.

The pyroclastic rocks are crudely bedded tuffs and agglomerates. The bedding, well-developed crossbedding in some places, and common association with pillowed rocks indicate that most of the pyroclastic rocks are waterlain materials. The bedded and massive pyroclastic rocks are composed mainly of greenish grey, moderately to strongly amygdaloidal and in places almost spongelike fragments. Recognizable nonvolcanic fragments in the pyroclastic rocks are a few small slabs of black shale, dark grey chert, and iron-formation.

Potassium feldspar is a common constituent of the Nimish rocks in the Knob Lake area. Some of the pyroclastic rocks contain as much as 20 to 30 per cent of the feldspar. Most of the feldspar is probably adularia. It occurs most commonly as partial or complete fillings of vesicles and as cement between rock fragments (Fig. 10). Replacement of plagioclase, and of the fine-grained chloritic groundmass, by potassium feldspar is also clearly evident in some of the volcanic fragments.

Most of the potassium feldspar is very fine grained and difficult to identify even in thin section. It could be mistaken for albite, quartz or chert. In many instances staining with cobaltinitrite and X-ray methods are the only means by which the feldspar can be identified. In hand specimen most of the feldspar is white to colourless, although some of the translucent varieties appear grey to dark grey. Less common colours are pink and pale yellow to

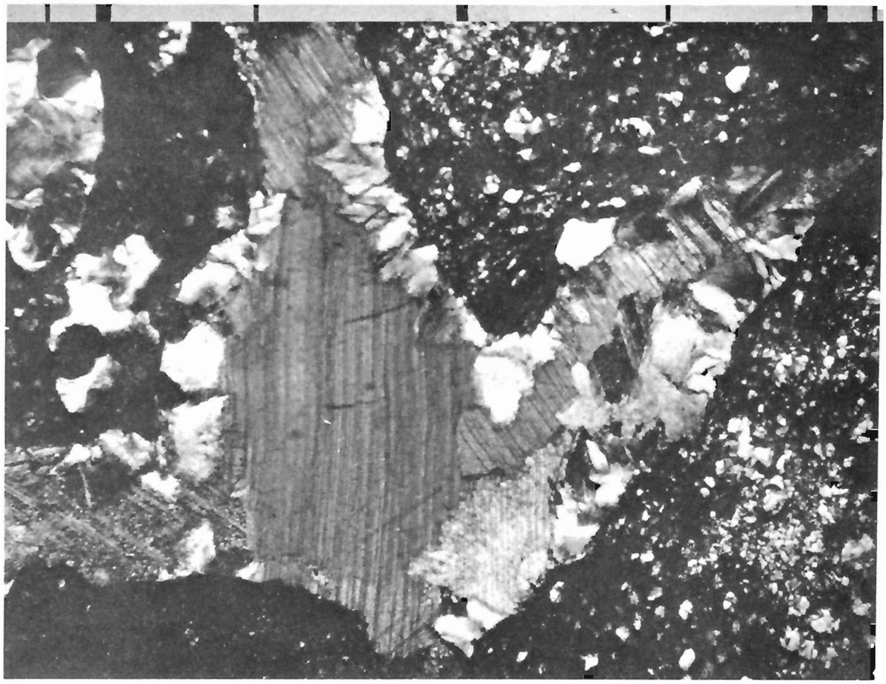


Figure 11. Same as above with crossed nichols. Note the great range of grain size of the feldspar.

brown. In thin section the feldspar shows a great variety of textures ranging from subparallel, partly radiating aggregates of elongate to spear-shaped crystals which form incrustations around volcanic fragments, to irregular, massive or patchy aggregates of randomly oriented, anhedral grains of variable size (Figs. 10 and 11).

Intrusive Rocks

Nimish intrusives occur as sills in the Wishart Quartzite and to a lesser extent in the Attikamagen Formation. The intrusives are predominantly medium to coarse grained, and composed mainly of feldspar and augite. Thin sections show large, slightly to strongly altered laths of plagioclase surrounded by ophitic, remarkably fresh augite. Scattered patches of chlorite and serpentine may represent remnants of olivine. Potassium feldspar, which seems to be characteristic of Nimish rocks, occurs as intergrowths in plagioclase and as rims around plagioclase crystals. The rims locally exhibit a subparallel arrangement of grains which protrude outward from the plagioclase, a texture very similar to some of the adularia encrustations in pyroclastic rocks.

Not all greenstones in the area can be definitely classified as volcanic or intrusive. Scattered outcrops of fine- to medium-grained diabasic greenstone may be fine-grained portions of sills or coarse lavas.

The intrusive and extrusive rocks occur in the same general area and are similar in composition. There can be little doubt that they are consanguineous. Some intrusives, as suggested by Perrault (1955), may be offshoots from the same conduits which supplied the extrusive lavas.

Summary of Events in the Nimish Area

Volcanism in the Nimish area began during deposition of the Attikamagen shales, continued intermittently, and reached a climax during deposition of the Sokoman Formation. It ended at the time when Menihek sediments began to accumulate.

According to Sauve (1953), the igneous and tectonic activity that led to volcanism in the Nimish area resulted in local uplift of the area and formed a number of volcanic islands in the shallow seas. The most intense volcanic activity occurred at the beginning and towards the end of the Sokoman sedimentation. Volcanic centres in the present Dyke-Astray area were most active and produced the largest quantities of volcanic materials. This area also appears to have been the most frequently uplifted and eroded, particularly in the later stages of Sokoman sedimentation. Volcanics, associated iron-formation, Wishart Quartzite, and probably some of the shallow intrusives, were partly eroded and redeposited. Erosion of the rugged islands produced coarse conglomerates and breccias which were dumped within a short distance of the elevated areas. Finer clastics, however, fanned out eastward and northwestward for tens of miles beyond the Nimish area.

Sokoman Formation

The bands of iron-formation that appear among the miogeosynclinal sediments throughout the length of the Labrador Trough belong to the same stratigraphic unit which in the central part of the geosyncline is known as the Sokoman Formation.

The Sokoman Formation, composed mainly of alternating silica-rich and iron-rich beds, is a typically cherty Precambrian iron-formation of the Superior-type (Gross, 1965, p. 91). The granules, oolites, and the predominantly thin but commonly irregular bedding, features which are characteristic of the Superior-type iron-formations, are typical of the Sokoman. The intraformational conglomerates, crossbedding and abnormally thick beds, which are also present in many of the iron-formations in the Lake Superior region, are particularly common in the Sokoman Formation. The mineralogy of the Sokoman and other Superior-type iron-formations is basically similar. Chert, magnetite, minnesotaite and siderite are the most common minerals. The chert-iron oxide units (oxide facies) of the Sokoman Formation, however, are more extensive, better differentiated, and contain more primary hematite than most of the iron-formations in the Lake Superior region.

The Knob Lake area is one of the few places in the Labrador Trough where the iron-formation shows obvious signs of having been affected by penecontemporaneous volcanic activity. The iron-formation in the central and south-central part of the Knob Lake area contains a number of stratigraphic units which contain tuffaceous and detrital material that has been derived from the nearby volcanic Nimish area during deposition of the iron-formation.

Menihek Formation

The Menihek Formation, the youngest division of the Knob Lake Group, is composed almost entirely of grey to pitch black, carbonaceous and locally pyritic shales, slates, and minor siltstones. Chert and dolomite are rare, and fine-grained, feldspathic greywackes are common only in the eastern parts of the area. The thickness of the formation is estimated to be more than 1,000 feet.

The contact between Menihek and the underlying iron-formation is sharply gradational to abrupt. The two formations appear to be quite conformable in the Knob Lake area, but in a few places in other parts of the Labrador Trough (Harrison, 1952; Sauve, 1953; Gross, 1968) discordant relationships, sandy beds, and rare conglomerates at the contact between the Sokoman and Menihek formations, indicate at least very local unconformities in this part of the sequence.

Diabase Dykes

Several north-south trending diabase dykes extend across the central part of the area. They are fine- to medium-grained, dark greenish grey rocks, composed mainly of labradorite, augite and minor olivine. The chief accessory is ilmenite. Pyrite, pyrrhotite and chalcopyrite are rare.

The dykes cut across the folded and faulted rocks in the area without any evidence of being affected by the structure of these rocks. They are clearly the youngest igneous rocks in the area, probably upper Proterozoic.

STRUCTURE

The structure of the area is characterized by northwestward-trending folds and faults. This trend is reflected by the topography and drainage of the area and can be recognized on most maps and aerial photographs.

The complexity of structure is variable, but tends to increase eastward. Near the western margin of the Labrador Trough, the Proterozoic sediments are essentially undeformed. The strata undulate very gently and maintain an average dip of five degrees to the east. East of Stakit Lake the structure is more complex. Folds are more numerous, larger and sliced by many faults.

The laminated and very thin bedded rocks have undergone the greatest amount of deformation and in some places are intricately folded. Drag folds and tight, doubly plunging folds are also seen in the more competent rocks such as quartzite and cherty iron-formation. However, they are only minor, parasitic structures which were formed locally as complements of major folds and faults. The major folds in most of the area are broad and open, although many of them are asymmetrical with axial planes inclined to the west. In places, the western limbs of folds are nearly vertical but are rarely overturned.

By far the most numerous faults are high angle reverse faults that strike northwest and dip east. Vertical displacements along these range from a few feet to several hundred feet, but rarely exceed 1,000 feet. Imbricate sets of eastward-tilted fault blocks are very common and account for much of the easterly dip of bedding in the area.

The regional structural pattern indicates clearly that the prevalent orogenic forces which affected the central and northern parts of the Labrador Trough acted from the northeast. The Hudsonian orogeny is generally

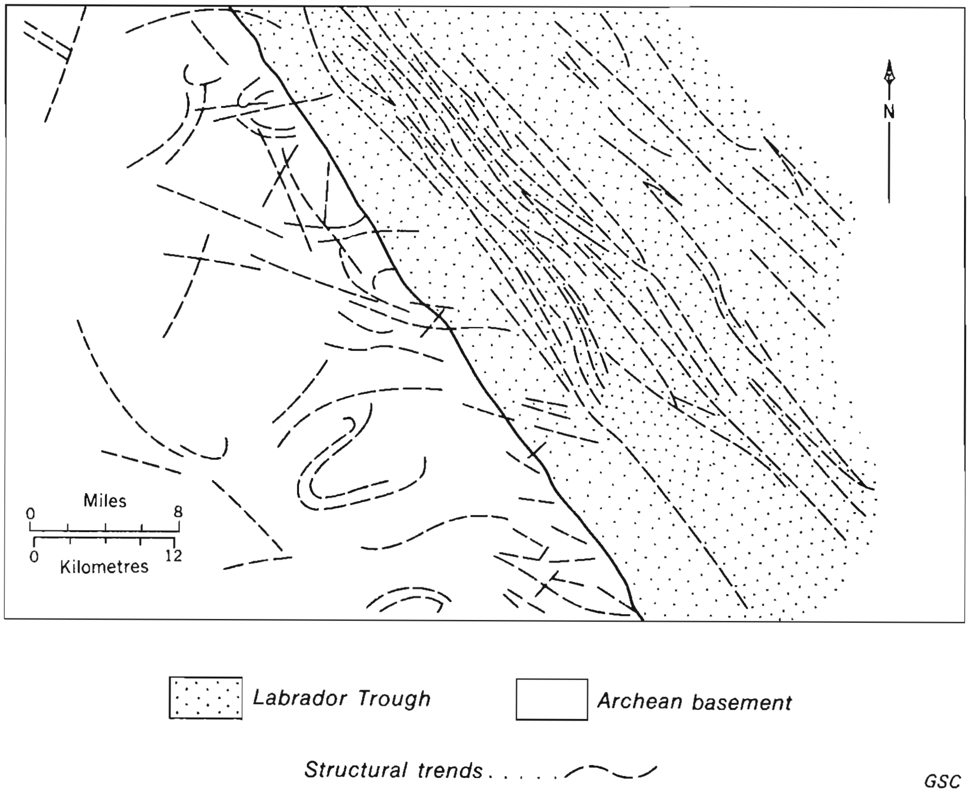


Figure 12. Comparison of structural trends in Labrador Trough and the Archean basement. Based on maps by Frarey (1961) and Stevenson (1962).

recognized as the main orogenic period. Pre-Hudsonian and post-Hudsonian structural readjustments have also occurred in the history of the Labrador Trough but they are comparatively minor tectonic events. The Archean basement - at least that part of it which underlies the western Labrador Trough - was apparently little affected by the Hudsonian orogenic forces since it does not show the same predominant structural trends as the Proterozoic rocks (Fig. 12).

Harrison (1952) estimated that the compressive forces foreshortened the original 75-mile-wide western segment of the geosyncline to its present width of 15 miles, a shortening of 80 per cent. In view of the present knowledge of the structure in the area, this amount of crustal shortening seems excessive. Seguin (1961) on the basis of 25 detailed sections (200 feet to 1 inch) in a mile-wide strip in the central part of the ore zone obtained an average shortening of 44 per cent. The writer's estimates (Fig. 13) agree more closely with those of Seguin. The estimates of minimum shortening (stippled blocks in Fig. 13) are based on 1,000 feet to 1 inch structural cross-sections of the central part of the Knob Lake area. To estimate the maximum possible shortening, the minimum estimates were compared with calculations based on Seguin's sections and detailed sections of the mines. Twenty per cent was

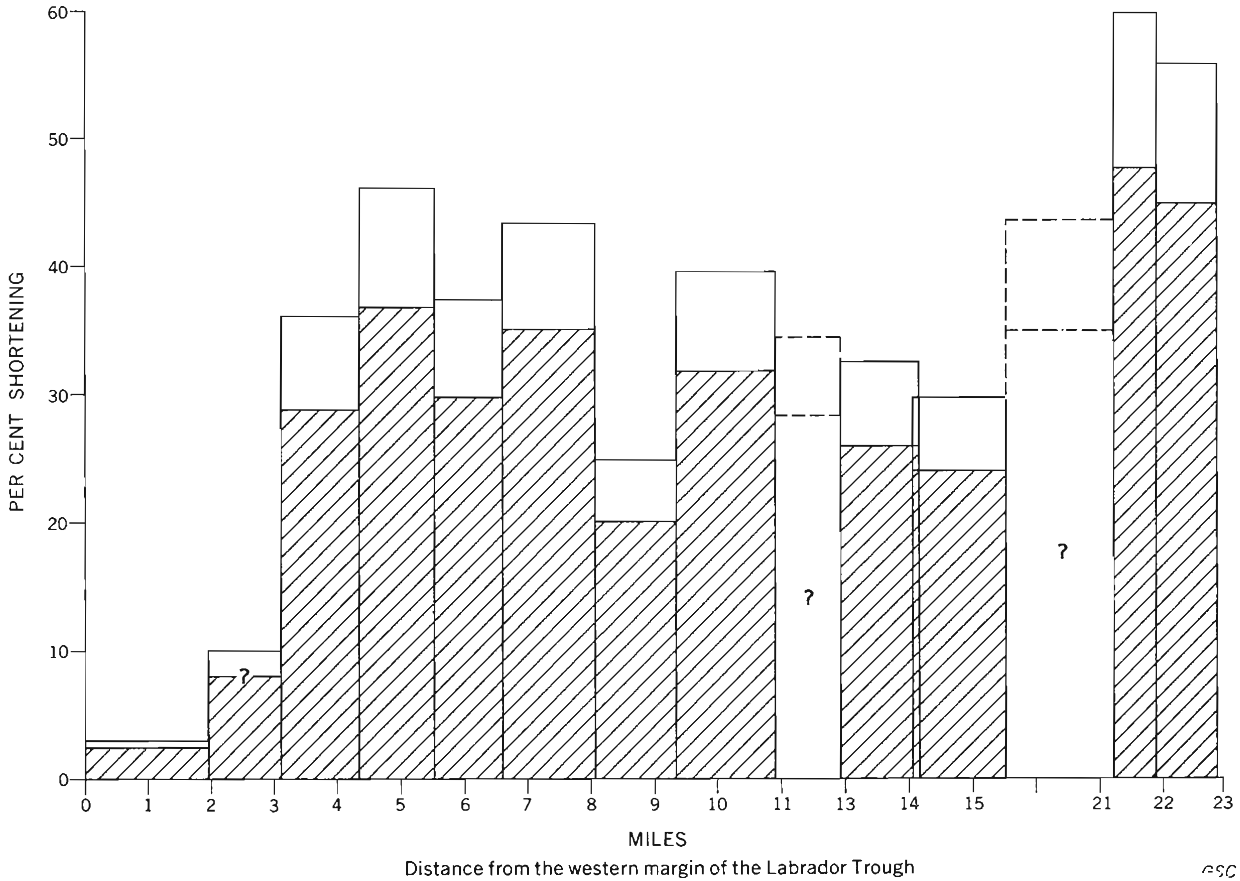


Figure 13. Minimum (shaded) and maximum (blank) crustal shortening in various parts of the Knob Lake area.

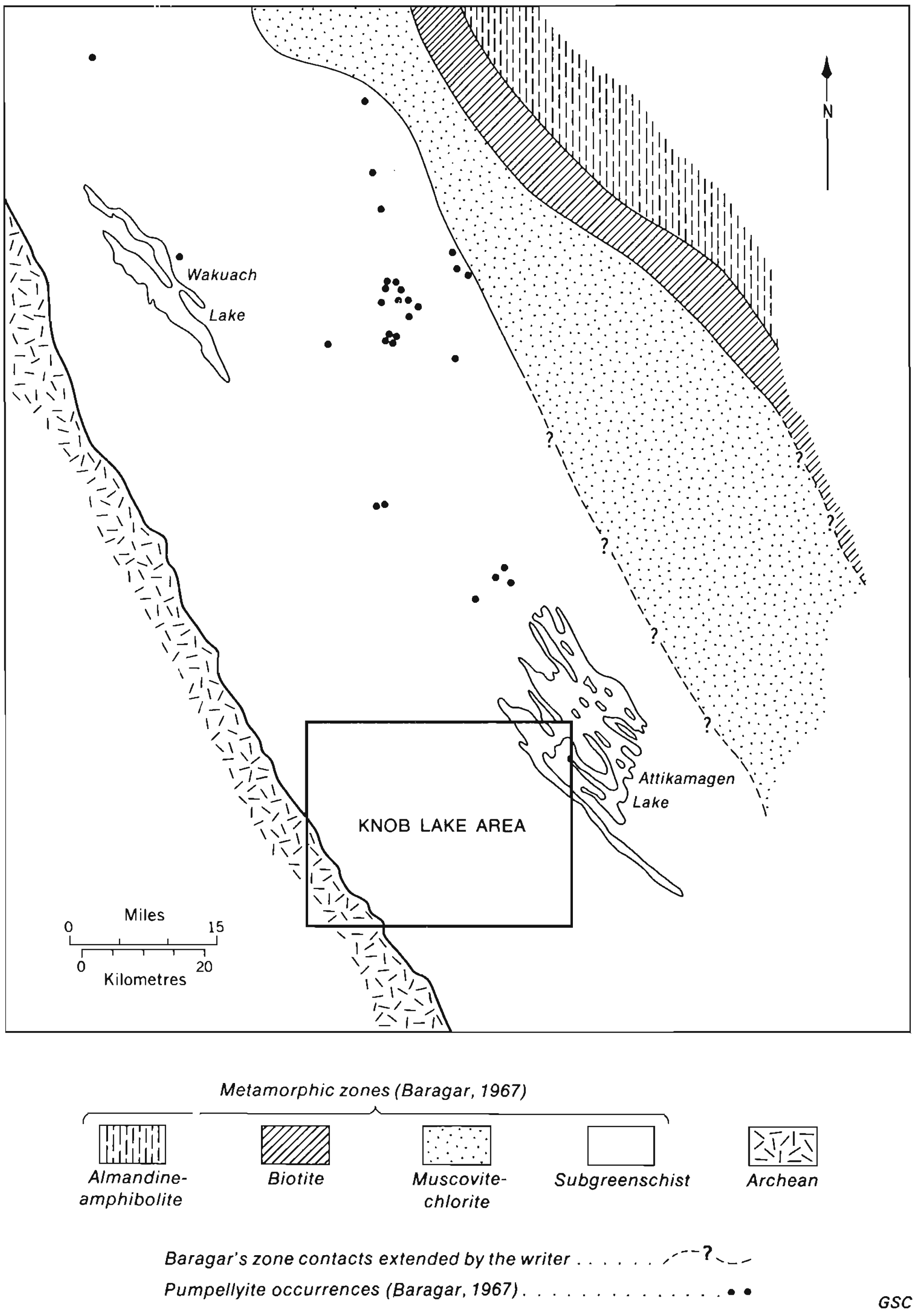


Figure 14. Zones of regional metamorphism.

the maximum discrepancy between the estimates for sections one mile wide. This figure was then used to calculate the maximum shortening in different parts of the area. The average minimum and maximum shortening for the entire 23-mile length of the section was found to be 28 and 35 per cent respectively.

METAMORPHISM

The intensity of metamorphism in the central Labrador Trough increases eastward, from slightly metamorphosed sediments in the west to strongly altered schists, gneisses and amphibolites in the east, with isograds extending approximately parallel to the long axis of the Trough (Harrison, 1952; Baragar, 1963, 1967; Gross, 1965, 1968). Baragar (1967) recognized four metamorphic zones in the central Labrador Trough ranging from subgreenschist to amphibolite (Fig. 14). The subgreenschist zone is the zone of incipient metamorphism (approaching the grade of zeolitic facies?).

The Proterozoic rocks in the Knob Lake area are well within the subgreenschist zone and show little evidence of metamorphism. The iron-formation is in places obviously recrystallized; however, most of the changes are probably diagenetic rather than metamorphic. As in the area described by Baragar, "no systematic geographical variation in mineralogy or textures that might be related to regional metamorphism can be detected" (Baragar, 1967, p. 56). Pumpellyite, which Baragar considered characteristic of the subgreenschist zone, could not be found even in rocks of the appropriate composition. Chlorite, albite, calcite and locally adularia are the characteristic secondary minerals of the fine-grained volcanic and intrusive rocks – not pumpellyite. Metamorphism in the Knob Lake area was probably not sufficiently intense to form pumpellyite.

The temperature of metamorphism in the area is estimated to be less than 300°C, and in all liklihood below 200°C, on the assumption that the boundary between the greenschist facies and facies of lower metamorphic grade is approximately 300°C (Turner and Verhoogen, 1960).

STRATIGRAPHY OF THE SOKOMAN FORMATION

Despite considerable work done in the central Labrador Trough, the stratigraphy of the Sokoman Formation is still not well known in many areas. The attention of most investigators in the Knob Lake area has been focused mainly on the rich deposits of soft iron ore and the partly altered iron-formation associated with them. Unaltered iron-formation, particularly that outside the areas of economic interest, has received comparatively little attention.

PREVIOUS SUBDIVISIONS OF THE SOKOMAN FORMATION

The sediments bounded by the Wishart Quartzite and the Menihek Formation were originally subdivided into two formations - Ruth and Sokoman.

According to the old field manuals of the Iron Ore Company of Canada and the early reports of Gross (1951) and Harrison (1952), the name Ruth Formation was intended for the ferruginous black shales on top of the Wishart Quartzite. A comparison of various areas covered by different maps and reports shows, however, that subsequent usage of the name has not been consistent.

In the mines, where the iron-formation and associated sediments are strongly altered, the name Ruth Slate has been applied to all red-altered sediments immediately above the Wishart Quartzite. This has generally included not only the "slates" (shales) but also the laminated and thin-bedded, siderite-rich iron-formation which is visually indistinguishable from the "slates" when intensely altered. In the field, the distinction between "slates" and other fissile cherty units of the iron-formation has likewise not always been made. In many places, it is difficult to tell exactly which parts of the iron-formation were assigned to the Ruth and which to the Sokoman. In general, it seems that most fissile rocks at the base of the sequence, particularly if they were of red or brown colour, were collectively designated as Ruth Slate or Ruth Formation.

The definition of the Ruth has recently become even more complicated. The work of the geologists of the Iron Ore Company of Canada during 1963 to 1966 and the writer's investigations have shown that the Ruth of the main ore zone (the type locality for the formation) is stratigraphically equivalent to the lowermost carbonate, silicate and oxide facies of the Sokoman Formation in Stakit Lake and other areas. The company geologists, in order to conform with the terminology established in the main ore zone, have chosen to retain the name Ruth but redefined it so that it now locally includes 100 feet or more of the iron-formation that was previously known as Sokoman.

The writer believes that the name "Ruth Formation" should be abandoned. The Ruth and Sokoman are in part time equivalent stratigraphic units and are chemically, mineralogically and genetically related. To maintain them as separate formations is misleading and creates unnecessary problems in correlation and interpretation. Furthermore, the name Ruth has been used in so many ways by various geologists that by now it has little meaning except in the mines of the main ore zone where, through the years, it has acquired its own specialized definition and use. Ruth Formation should therefore be omitted from the formal stratigraphic terminology used in the central

Table 2

Subdivisions of Sokoman Formation

Harrison 1952	Dufresne 1952	Iron Ore Co. of Canada 1952	Schwellnus 1957	Iron Ore Co. 1958-1961 Mines	Iron Ore Co. 1958-1961 Ore Zone	Seguin 1961	Iron Ore Co. of Canada Mines 1962	Iron Ore Co. of Canada Stakit L. 1963, 1965	Iron Ore Co. of Canada Snow L. Hematite L. 1966	Zajac This Rept. Table 3
Slaty Member				LC	LC	LC	LC	LC	LC	X
	Massive lean chert		LC							
Lean Chert				RUIF	RUIF Magnetic greywacke	RUIF	RUIF	RUIF	RUIF	IX
		Lean chert								
Massive Cherty Member			RUIF			LLC				VIII
	Banded carbonate		YUIF							
Cherty Iron Carbonate Member			GUIF	GUIF	GUIF	GUIF	GUIF	GUIF	GUIF	VII
Cherty Metallic Member	Thick-banded Jaspilite		URC	URC	URC YMIF	URC	URC	URC	URC	VI
		Metallic iron-formation	BC	BC	BC	BC				
Thick-banded Jasper Member	Cherty metallic iron-formation		GC	GC	GC	GC	PGC	PGC	PGC	V
			PC	PC	PC	PC				
Banded Cherty Member	Silicate-Carbonate Iron-formation		LRC	LRC	LRC	LRC	LRC	LRC	LRC	IV
Thin-banded Jasper Member	Thin-banded jaspilite	Silicate carbonate iron-formation	SCIF	SCIF	SCIF	SCIF	SCIF	LG	LIF: LG	III
			?		?			RUTH: Ruth chert	LIF: LIF	II
Banded Silicate Member	Banded lean chert							RUTH: Jaspilite	RUTH: Jaspilite	
								RUTH: Ruth chert		
Ruth Formation	Ruth Slate	Ruth Slate	Ruth Slate	Ruth Slate	Ruth Slate	Ruth Slate	Ruth Slate	RUTH: Ruth Slate	RUTH: Ruth Slate	I
										Basal chert

BC = brown cherty
GC = grey cherty
LC = lean cherty
LG = lower grey

LLC = lower lean chert
LRC = lower red cherty
PGC = pink grey cherty
PC = pink cherty

RUIF = red Upper Iron-Formation
SCIF = Silicate Carbonate Iron-Formation
YMIF = yellow middle Iron-Formation
YUIF = yellow Upper Iron-Formation
GUIF = grey Upper Iron-Formation
LIF = lower Iron-Formation

Labrador Trough, and the ferruginous shales, for which the name was originally intended, relegated to their proper place, a facies of the Sokoman Formation.

The name Sokoman Formation was originally applied to the cherty, iron-rich sediments above the Ruth Formation. The first geologists in the area subdivided the Sokoman into various rock types on the basis of colour, texture and composition. The rock types, although generally listed in sequence of their most common occurrence, were not restricted to any particular part of the stratigraphic column. The correlation of such generalized sequences in different areas was almost impossible (see Table 2 under Harrison and Dufresne).

The opening of the French, Ruth Lake and Gagnon mines for production helped to resolve some of the stratigraphic problems. The restriction of investigations to a small area eliminated most of the variability of the iron-formation that was the result of facies changes, and the open pit mines allowed a much closer study of the alteration and structure of the iron-formation than was previously possible. With this advantage over their predecessors a number of company geologists, notably Schwellnus (1957), produced a subdivision of the iron-formation which, with a few changes, has persisted in the mine areas to the present time. The subdivision presented by Schwellnus (1957) with some modifications (see subdivisions used by Iron Ore Co. of Canada 1958-62, Table 2) permits a useful classification of the iron-formation in the main ore zone, particularly in the mines, but is difficult to adapt to the iron-formation in other parts of the Labrador Trough. Schwellnus' stratigraphic subdivisions and nomenclature are too inflexible. Both are based on the lithological characteristics of the iron-formation within a restricted area and cannot adequately cope with the extensive facies changes which become apparent in regional studies of the iron-formation.

PROPOSED SUBDIVISION OF THE SOKOMAN FORMATION

In the attempt to extend the stratigraphic sequence established in the main ore zone to other parts of the Knob Lake area, the writer was faced with the problem of either extensively redefining the existing stratigraphic units and adding new ones or introducing an entirely new system for classifying the iron-formation. The latter course was chosen as the simpler and less confusing.

In the new classification the entire sedimentary sequence bounded by the Wishart Quartzite and the Menihek Formation is designated as the Sokoman Formation.

The formation as here defined is subdivided into eleven members. Each member, with the exception of the Basal Chert, is designated by a Roman numeral which indicates the relative position of that member within the stratigraphic sequence.

The most important criterion in the definition of the various members is their position in the stratigraphic sequence as determined by their relation to certain marker horizons. The peculiarities of colour, texture, composition and weathering are important in differentiating between the members in any one location and are useful in short range correlations. Because of facies changes, however, long-range correlations, particularly of such highly variable members as I, V and VII, are practically impossible on the basis of lithological similarity alone. Detail correlation of stratigraphic sequences in widely separated areas can be made only by reference to stratigraphic marker horizons which persist with comparatively little change. The marker horizons from the bottom to the top of the sequence are:

(1) Wishart Quartzite, Basal Chert member
(2) Upper contact of member I
(3) Member IV (LRC); in places member III
(4) Member VI (URC-YMIF)
(5) Member VIII (LLC); in places member IX (RUIF)
(6) Lower contact of Menihek Formation

In places where the marker horizons cannot be distinguished and the members in question cannot be differentiated in any other way, such members must be treated as one unit. This can be done simply by combining the symbols for the individual members. For example, IV-VI indicates that members IV, V and VI are combined as one unit.

The facies of the iron-formation are rock-stratigraphic units and are classified on the basis of their dominant primary iron minerals as oxide facies (O)[1], silicate facies (S), carbonate facies (C) and sulphide facies (Sf). This is the same system of classification as used by James (1954, 1966) and Gross (1965). One modification is made by the writer. The name "clastic facies" (abbreviation symbol x) will be used to denote those units of the iron-formation which contain clastic quartz, feldspar or any other non-iron-formation material. Most Precambrian iron-formations are noted for the scarcity or absence of foreign clastic contaminants. Beds, lenses and larger sedimentary units of shale, greywacke or tuffaceous sediment, which appear locally in some iron-formations, represent unusual events in the history of the iron-formations, outbursts of volcanic activity or local uplifts and erosion of the landmasses bordering the sedimentary basins, and deserve special recognition.

BASES FOR INTERPRETATION OF DEPOSITIONAL ENVIRONMENTS

Interpretations of the stratigraphy and lithology of each member of the Sokoman Formation are directed towards reconstruction of the depositional environment of the formation. The interpretations are based mainly on chemical-mineralogical and physical-textural characteristics of the formation in comparison with ideal sedimentary models.

Chemical-Mineralogical Model

The characteristics of chemical sediments during deposition and diagenesis are controlled largely by the hydrogen ion concentration (pH) of the environment and by its oxidation-reduction potential (Eh). Since sedimentary

---

[1] In parentheses are the abbreviation symbols for the facies used by the writer.

MEMBERS	FACIES	PREVIOUS DESIGNATION OF UNITS (See Table 2)
X	Carbonate Facies X: C Chert Clastic Facies X: x	LC
IX	Silicate-oxide Facies IX: O-S Carbonate Facies IX: C Clastic Silicate-oxide Facies IX: S-O-x	RUIF Mag. Gyke. Mag. Shale
VIII	Chert (Oxide-silicate Facies, VIII: S-O-x)	RUIF LLC
VII	Oxide Facies VII: O Carbonate Facies VII: C Silicate Facies VII: S	GUIF
VI	Silicate Facies VI: S Oxide Facies VI: O	URC (YMIF)
V	Oxide Facies V: O	PGC
IV	Oxide Facies IV: O	LRC
III	Silicate Facies III: S	SCIF LG
II	Oxide-silicate-carbonate Facies II: O-S-C	LIF
I	Oxide Facies Silcate Facies I: S Carbonate Facies I: C Clastic Silicate-sulphide Facies I: S-Sf-x	RUTH or RUTH
BASAL CHERT		Ruth Cht. Wishart Cht.

GSC

Table 3. Proposed subdivisions of the Sokoman Formation.

iron occurs in several minerals with different ratios of ferrous and ferric iron, the Eh-pH diagrams are very useful in interpretation of the natural mineral associations.[1]

A knowledge of the free energy of pyrite, siderite, magnetite and hematite enables their stability fields to be readily calculated (Fig. 15). A problem arises, however, with natural iron silicates for which thermodynamic data are not available. The diagram in Figure 16 was obtained by assuming hypothetical iron silicates (greenalite type) with varying ferrous-ferric ratios. Since no free energy data are available for such silicates, only the slopes of the equilibrium lines could be calculated. The position of the stability fields of the silicates cannot be determined. However, if ferrous-ferric silicates can coexist with magnetite and siderite, and natural mineral assemblages indicate that this is true, then their stability fields have to lie somewhere between the stability fields of magnetite and siderite.

Using Eh-pH diagrams as guides and natural mineral associations as evidence, James (1954, 1966) devised a hypothetical depth-to-shore sequence of mineral facies. In his model the sulphide facies are confined to the deepest stagnant and strongly reducing part of the basin, whereas the oxide facies occupy the well-aerated, oxidizing shallows near the shore. The intermediate parts of the basin, which range from shallower, mildly oxidizing environments to deeper, weakly reducing environments, are occupied respectively by silicate and carbonate facies.

Another possible model is one in which iron silicates do not appear. As far as the Eh-pH controls are concerned (Figs. 15 and 16), the silicate field can be bypassed entirely if the changes from reducing to oxidizing conditions take place at low pH. The ideal depth-to-shore sequence in such a case would be sulphide-carbonate-oxide.

The models outlined above are based on idealized situations and are not likely to be reproduced in such simple form in the more complex natural environments. James (1954, p. 242) explains this as follows:

> It is doubtful if the pattern of precipitation indicated is ever actually obtained in nature because of complicated relationships between depth of basin, height of barrier and details of circulation . . . .

James (1966, p. 15) also observed:

> No example is known of a complete array of major facies grading laterally one into the other . . . . The main reason for this is that the shore-to-depth profile is merely a device for indicating a range in environmental conditions, particularly oxygen availability.

Mineralogical changes during diagenesis and metamorphism may further complicate the ideal sedimentary models. Post-depositional changes, however, with the exception of some strongly metamorphosed iron-formations, are rarely so extensive as to invalidate completely the use of the sedimentary models in the deciphering of depositional environments.

---

[1] Eh-pH diagrams and their use in environmental interpretations of iron-rich sediments is discussed fully by Krumbein and Garrels (1952), James (1954, 1956) and Garrels and Christ (1965).

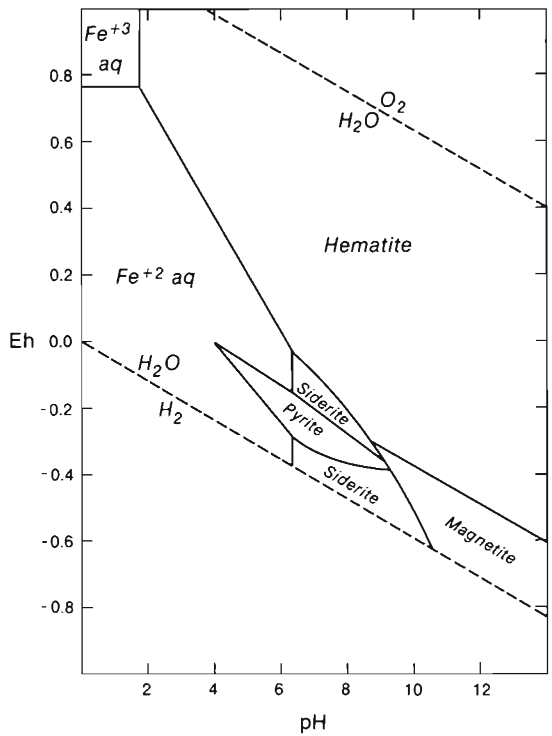
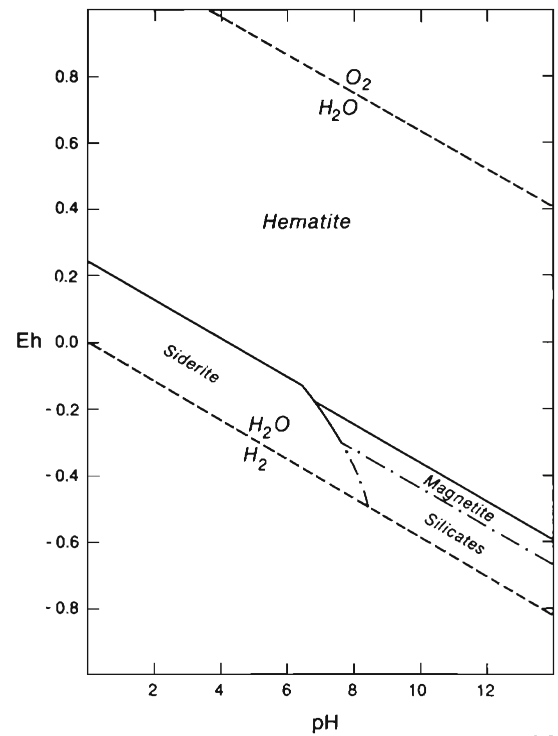


Figure 15. Stability relations of hematite, magnetite, siderite and pyrite in water at 25°C. Total dissolved sulphur = 10-6 M/L. Total dissolved carbonate = 10° M/L. After Garrels (1965, p. 224).



GSC

Figure 16. Stability relations of hematite, magnetite, siderite and hypothetical iron silicates with different Fe+2-Fe+3 ratios in water at 25°C. Total dissolved carbonate = 10 M/L. Only fields of solids are shown.

Physical-Textural Model

Although iron-formations are chemical sediments, many of their textures are independent of chemical controls. Iron-formations which are deposited within range of waves and currents, or are at times partially and temporarily exposed above water, are likely to be, as other chemical sediments in similar situations, at least partly reworked, fragmented, tumbled, shifted and then redeposited as clastics of the originally strictly chemical precipitates. Such reworked precipitates would then show at least some characteristics of clastic sediments.

The Sokoman Formation, and to a lesser extent the iron-formations of the Lake Superior region, contain a number of sedimentary textures and structures which can be interpreted in terms of the physical environment of deposition. Figure 17 is an interpretation of these sedimentary features.

The threefold subdivision of depositional environment is similar to that of Pettijohn (1957, p. 593). Deep refers to depositional areas essentially below wave base; intermediate denotes shallower environments intermittently affected by waves and currents; and shallow applies to environments dominated by turbulent conditions, the shallowest areas of deposition. The correlation of textures and structures with different depth and energy zones is, of course, over-simplified and does not always apply. It is an idealized model which can be used only as a guide in environmental interpretations.

Conglomerates and crossbedding are too well known to require much explanation. Both are most characteristic of the shallow high-energy environments. The conglomerates in the Sokoman Formation and most of those in the Gogebic (Hotchkiss, 1919; Huber, 1959), Mesabi (Gruner, 1946; White, 1954), and Gunflint (Goodwin, 1956). Ranges are of the intraformational type. Slabby to oval fragments and pebbles are clearly of local derivation. Some of the fragments may have formed by desiccation during temporary shoaling of the sea, and others by fragmentation due to strong wave and current activity.

Probably the first interpretation of bedding in iron-formation in terms of depositional environment was made by Hotchkiss (1919, p. 446):

> "It is believed that the wavy-bedded (i.e. thicker-bedded, p. 445) members of the iron formation are relatively shallow water deposits in which the bottom was within reach of waves that disturbed the bottom and produced the wavy-bedded structure . . . . The even bedded portions of the formation are believed to have been deposited in deeper water where the bottom was below the reach of wave action . . . ."

The interpretation of the thickness of bedding shown in Figure 17 is based on essentially the same line of reasoning; namely, the more turbulent, and ideally the shallower the environment, the thicker and more irregular the bedding.

The cause of the banded and laminated textures which are so characteristic of many Precambrian iron-formations is still unknown and widely disputed. Most geologists agree, however, that the thinly layered iron-formations had to be deposited in quiet, even "amazingly still" waters (Trendall, 1965) in order for the fine bands and delicate laminae to be

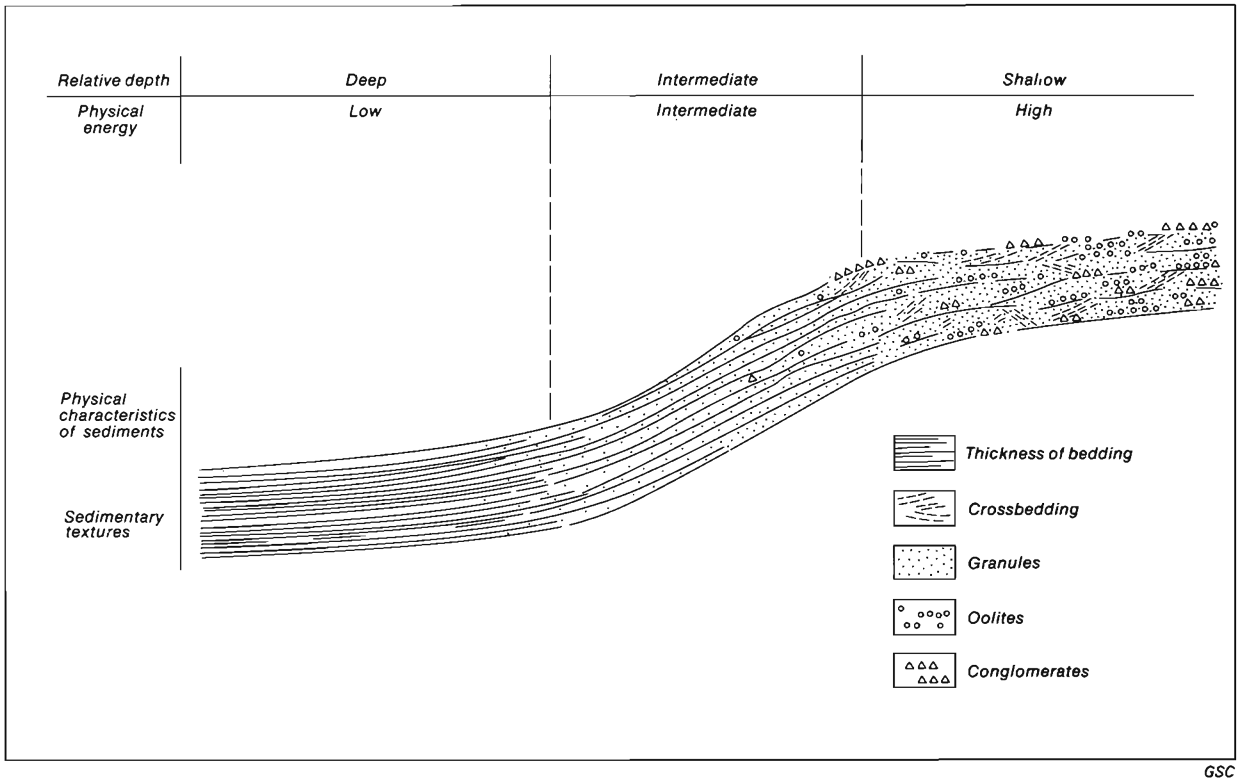


Figure 17. Relation of sedimentary textures to depth and physical energy of depositional environment.

preserved. The most probable places for quiet, low-energy conditions to exist and for thin-, even-bedded sediments to accumulate are the deepest parts of sedimentary basins. The shallowest areas, on the other hand, have the greatest chance of being affected by waves and currents and are therefore the most likely to produce thickly and irregularly bedded deposits.

Oolites are one of the best environmental indicators. They are easily recognized and, in comparison with some of the other structures and textures of iron-formations, are readily interpreted. Although numerous theories have been proposed for the origin of oolites, recent studies indicate that most of them are accretionary bodies formed in shallow, well-agitated waters (Illing, 1954; Newell et al., 1960; Rusnak, 1960).

Granules[1] are similar in shape and size to oolites but lack their concentrically laminated texture. Most of them are without any regular internal structure. The origin and environmental implications of granules are not clear. Field and laboratory evidence suggest at least three possible modes of origin for the structures: (1) gelatinous-colloidal, (2) intraclastic, and (3) organic.

The most popular and widely accepted belief is that the granules in iron-formations formed as globules or rounded aggregates from gelatinous precipitates. The belief is shared by Van Hise and Leith (1911), Gill (1927), Gruner (1946), White (1954) and Goodwin (1956), and is supported by the experiments of Van Hise and Leith (1911) and Moore and Maynard (1929). Van Hise and Leith were successful in duplicating granular structure in precipitates of "greenalite" and Moore and Maynard (1929, p. 519) observed that silica precipitated from carbonated sea water "was not a homogeneous mass but consisted of a great number of elliptical and spherical globules of gelatinous silica . . . ." The delicate granules of very uniform texture and composition (mainly chert) in the upper part of the Sokoman Formation may be of this origin.

The work of Barghoorn and Tyler (1965), Mengel (1965), LaBerge (1967) and Dimroth (1968) indicates that many granules in Precambrian iron-formations are intraclasts formed by penecontemporaneous reworking of the sediments by waves and currents. Barghoorn and Tyler (1965) concluded, after completing their study of the Gunflint cherts, that:

> "The morphology of the granules and the intimate association with clastic grains suggest that the granular particles were not deposited as colloidal globules or aggregates but that they are clasts which were deposited in an environment of strong wave or current activity."

They also observed that the

> "occurrence of cross-laminated units (in the Mesabi Range and Gunflint) indicates that the granules behaved as clasts and that currents were responsible for their transportation and deposition."

---

[1] The nearly perfectly spherical particles generally less than 40 microns in diameter, commonly with centrally located opaque inclusions and in some instances with a suggestion of radial texture are referred to as "spherites", the name used by James (1954).

Similar evidence can be cited from the Knob Lake area. Particularly convincing is the occurrence of granules of different shape and composition in the same rocks, and the close association of granular units with intraformational conglomerates. In some conglomerates a complete transition can be seen from large rounded or subrounded fragments to rounded particles of granule size; in others, the evidence is not as obvious, but the conclusion must be the same. It is difficult to envisage a process which could produce coarse conglomerates without also producing intermediate- and fine-size materials.

The possibility that some granules are of organic origin also exists. Tyler et al. (1957) and Stinchcomb et al. (1965) described oval-rounded shapes in Precambrian carbonaceous shales as possible organic (algal) structures. The similarly shaped granules in black carbonaceous cherts may also represent organically formed aggregates.

Despite the diversity of possible origins, most evidence indicates that the granules formed in agitated waters at moderate to shallow depths. The association of granular beds with intraformational conglomerates, with crossbedded and oolitic units as well as the invariable presence of granules in very irregular and abnormally thick beds, is the strongest evidence that links the origin of granules with turbulent, wave- and current-worked areas of deposition. The almost complete absence of granules in the extensive, thinly banded and laminated iron-formations of Western Australia, South Africa and the Ukrainian Shield, which are generally accepted as quiet-water sediments, also indicates that at least slightly agitated waters are needed in order for the granules to form.

BASAL CHERT MEMBER

The Basal Chert member is a discontinuous layer of chert that covers much of the Wishart Formation over an area of at least 1,800 square miles of the westernmost central Labrador Trough. The stratigraphic status of this chert is uncertain. Geologists have assigned it to either the Wishart or the Sokoman Formation (as part of the "Ruth Formation" or "Ruth Slate"), and Schwellnus (1957) suggested that it should perhaps be considered a separate formation. The writer has included this chert as part of the Sokoman Formation because mineralogically and texturally it is very similar to some of the black chert interbeds in the lower and upper part of the formation (members I and X).

The thickness of the Basal Chert is variable. The chert is generally less than 20 feet thick, although locally (Wakuach Lake area) thicknesses of up to 55 feet have been recorded (Baragar, 1967). In the Knob Lake area a thickness of more than 4 feet is uncommon. In many outcrops only traces of the chert can be recognized.

In outcrop the chert is various shades of grey, and rarely white, pink or red. On fresh surfaces it is invariably dark grey to black. The rock appears to be very uniformly fine grained, is extremely dense, and in most places shows no evidence of bedding. Contacts with the adjoining rock types are very well defined. Along the contact with the Wishart Formation the chert locally incorporates boulders of quartzite, but its contacts in the boulders and the orthoquartzite on which they rest are sharp even in thin section.

Under the microscope the chert is microcrystalline and very commonly segregated into oval granules. Most granules are 0.5 to 1.5 mm in size and are slightly finer grained than the matrix. Many are mottled by black-brown carbonaceous(?) material.

The texture and composition of the Basal Chert appear to be very uniform over wide areas. The uniformity and purity of the chert indicate that, for a period of time after the sea transgressed and covered the upper conglomerates of the Wishart Formation, no clastic sediments accumulated near the western margin of the Labrador Trough. Chemical sedimentation during this time was confined almost entirely to deposition of silica.

MEMBER I

Member I is approximately 80 to 150 feet thick[1] and consists of four laterally gradational facies: (1) clastic silicate-sulphide, (2) carbonate, (3) silicate, and (4) oxide.

Clastic Silicate-Sulphide Facies (I:S-Sf-x)

The distribution and thickness of the facies are shown in Figure 18. Although the true thickness cannot be accurately determined in many places because of intricate small-scale folding, there is little doubt about the regional variation in thickness of the unit. The unit definitely becomes thinner to the north, east and west of the central area. The zero contour (Fig. 18) indicates the limits of the area in which the facies can be distinguished as a distinct stratigraphic unit.

The clastic silicate-sulphide facies is mainly a black, ferruginous, fissile shale with interbeds of chert and chert-carbonate. Tuffs are prominent only in the central part of the area (Fig. 18). Mineralogical classification of the facies, silicate-sulphide (abbreviated S-Sf), indicates the characteristic iron minerals of the dominant rock type (iron-rich shale). The presence of foreign clastic material (clastic quartz, feldspar, tuffaceous material) is one of the distinguishing characteristics of the facies and genetically important. It is indicated by the terms "clastic" (abbreviation symbol x).

The shale is evenly banded or laminated, and has excellent cleavage parallel to the bedding. Slaty cleavage is extremely rare even in strongly folded and faulted rocks. On fresh surfaces the shale is pitch black or less commonly dark green or grey. The weathering colours are various shades of brown. In the mines and other areas where the shale is closely associated with deposits of soft iron ore, its alteration colour is a characteristic dull to bright red.

Most thin sections of the shale are not very informative. Laminated and banded texture defined by thin cherty and silty layers, and in places suggestions of graded bedding, can be recognized, but not much else. Most of the rock is a fine-grained, dark grey to black, nearly opaque mixture, which,

---

[1] Thicknesses of 175 feet measured west of Ruth Lake and 201 feet north of Dyke Lake are unusual and may be due to repetition by faults or tight folds.

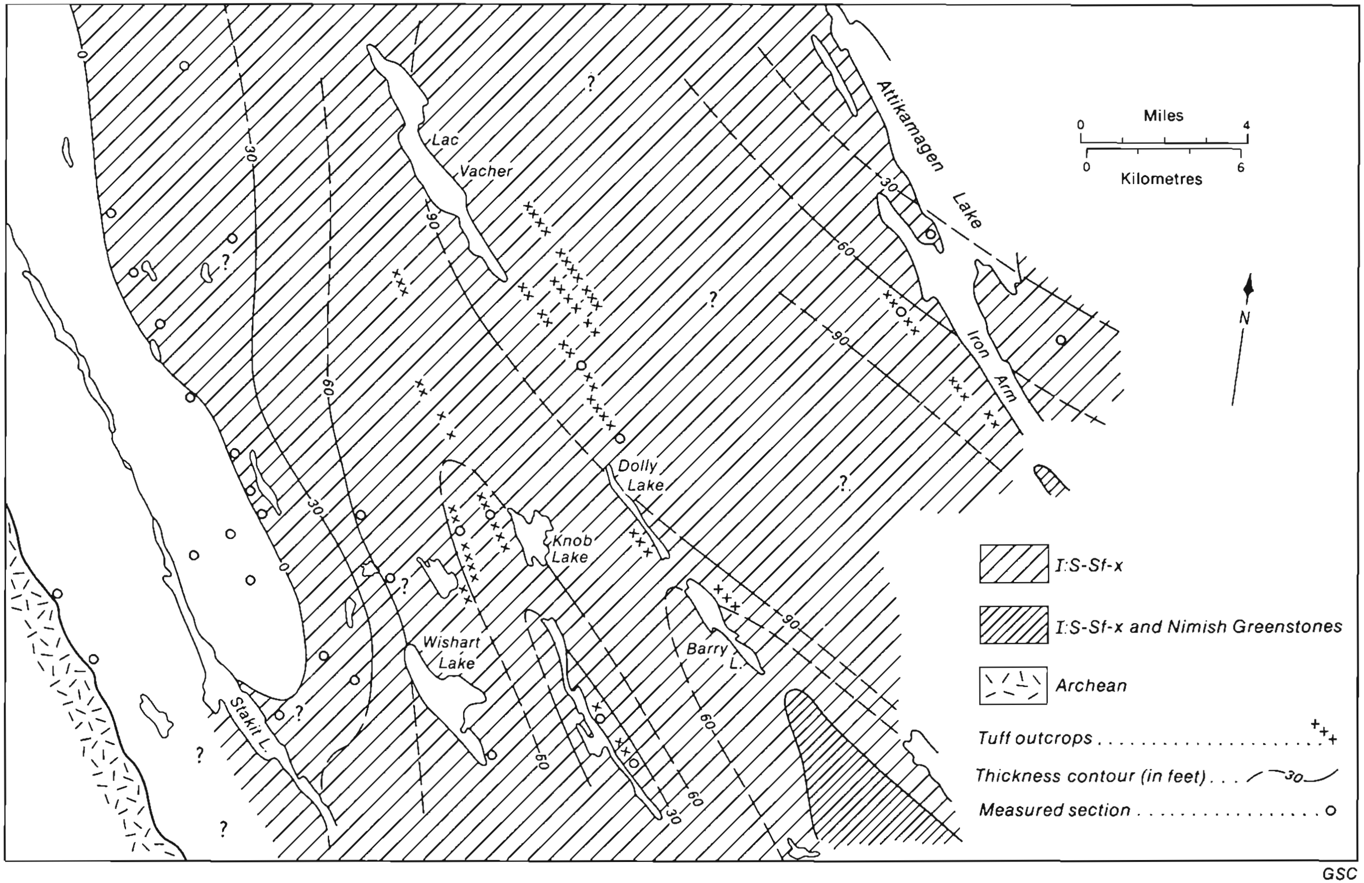


Figure 18. Thickness and distribution of clastic silicate-sulphide facies (I:S-Sf-x).

Figure 19. Outcrop of the carbonate facies of member I, in the central part of Stakit Lake area. Such coherent outcrops of this rock are not common, but the brown weathering and banded to laminated texture is typical.

according to X-ray analyses, is composed mainly of chlorite, potassium feldspar and quartz (chert?). Chemical analyses (Table 4) of composite samples show up to 2.79 per cent of carbon which may be organic (Harrison, 1952).

Iron silicates can only be recognized as minute feathery flakes within the nondescript, fine-grained material. X-ray analyses indicate the silicates to be mainly chlorite (iron septechamosite?). Stilpnomelane (?) is present in some of the more cherty and feldspathic beds.

The presence of sericite, reported by previous investigators, could not be confirmed. X-ray diffractograms obtained even from water-settled clay fractions of the shale show no evidence of sericite. The K_2O content of the shale (see Table 4) is derived mainly from potassium feldspar. The feldspar is microcrystalline and cannot be recognized in thin section. It is, however, readily identified in all diffractograms as one of the major components of the shale.

Most pyrite in the shale is very fine grained (3 to 8 microns) and difficult to identify in hand specimen. Polished sections show, however, that concentrations of several per cent and locally of 10 or 20 per cent are not uncommon. The pyrite appears to be most abundant in the central area where a few thin layers may contain as much as 50 per cent of pyrite (Fig. 20). On Dolly Ridge a pyritic horizon 7 to 13 inches thick with 20 to 30 per cent pyrite can be traced intermittently for a distance of nearly two miles.

Limonite and hematite are present only in weathered or altered rocks. Magnetite is very rare or absent.

The chert interbeds in the shale are microcrystalline, massive, or banded to laminated, and generally not more than a few inches thick. A 30-foot-thick lens of banded and laminated chert just northeast of the Wishart mine is exceptional. The banded and laminated cherts contain variable amounts of carbonate and in places small quantities of silicates. Some of the massive-looking black cherts west of Knob Lake are granular and closely resemble the Basal Chert.

Table 4. Analyses of shale and tuff from member I and other analyses selected for comparison

	Clastic Silicate-Sulphide Facies (IS-Sf-x)					Tuff	"Ruth Slate"		Tuff	Slate and Argillite		Average Shale
	Shale			?	Tuff	Nimish Area	Knob Lake		Cuyuna	Cuyuna Range		
	1	2	3	4	5	6	7	8	9	10	11	12
SiO_2	51.41	51.68	47.92	34.71	44.81	41.99	52.35	54.84	57.27	67.2	43.5	58.38
Al_2O_3	9.78	9.15	11.68	13.07	21.11	12.60	11.56	10.62	23.76	13.1	22.3	15.47
Fe_2O_3	19.46*	21.06*	22.61*	29.16*	3.01	1.45	13.00	13.39	1.63	9.1	16.2*	4.03
FeO	**	**	**	**	10.07	8.24	3.98	7.50	0.62	0.45	**	2.46
MnO	0.13	0.39	0.97	0.11	0.12	0.34	0.15	-	-	0.04	0.02	Trace
MgO	0.82	0.73	2.10	2.36	3.58	6.32	-	1.79	1.38	0.78	0.74	2.45
CaO	0.65	0.57	0.57	0.04	0.62	9.64	0.34	0.52	0.05	0.08	0.23	3.12
Na_2O	0.10	0.10	0.10	0.10	0.11	0.57	1.32	0.13	0.07	0.10	0.11	0.31
K_2O	5.98	5.04	4.71	5.27	7.86	5.94	4.54	3.06	6.43	4.5	3.7	3.25
H_2O-	2.09	1.70	1.05	0.51	0.17	0.26	1.33	0.97	0.05			1.34
H_2O+	4.91	5.22	5.33	4.11	4.49	3.52	7.34	4.46	4.63			3.68
TiO_2	1.58	1.34	1.12	0.39	2.18	1.26	1.86	0.42	3.38	1.8	2.3	0.65
Co_2	0.00	0.04	0.04	0.02	0.14	7.40	-		0.02			2.64
P_2O_5	0.01	0.00	0.00	0.10	0.22	0.19	0.27	0.11	0.23	0.08	0.23	0.17
S	0.25	0.65	1.98	15.30	1.68	0.01			0.05			0.65
BaO									0.16			0.81
F									0.11			0.05
C	2.79	2.01	0.86	0.01	0.19	0.05	1.78	2.18				
Total					99.73	104.51	99.82	99.99	99.84	92.72		100.46

Analyses 1 to 6 by J.L. Bouvier, Geol. Surv. Canada, 1967.

1. Black shale, Elizabeth Lake. Sample Z-1013 (composite of chip-samples across 30-foot-section).

2. Tuffaceous black shale, Gill Mine Ridge. Sample Z-1012 (composite of chip-samples across 40-foot section).

3. Tuffaceous black shale, Dolly Ridge. Sample ZA-1011 (composite of chip-samples across 50-foot-section).

4. Composite of several samples of 1-foot pyritic layer, Dolly Ridge. Sample ZA-1009.

5. Composite of several samples of tuff, Dolly Ridge. Sample ZA-1007.

6. Composite of several samples of calcite cemented tuff-agglomerate, Nimish area.

7-8. Gross (1951).

9-11. Schmidt (1963).

12. Clarke (1924).

* Total Fe expressed as Fe_2O_3.

** FeO not determined due to high content of carbon or sulphur.

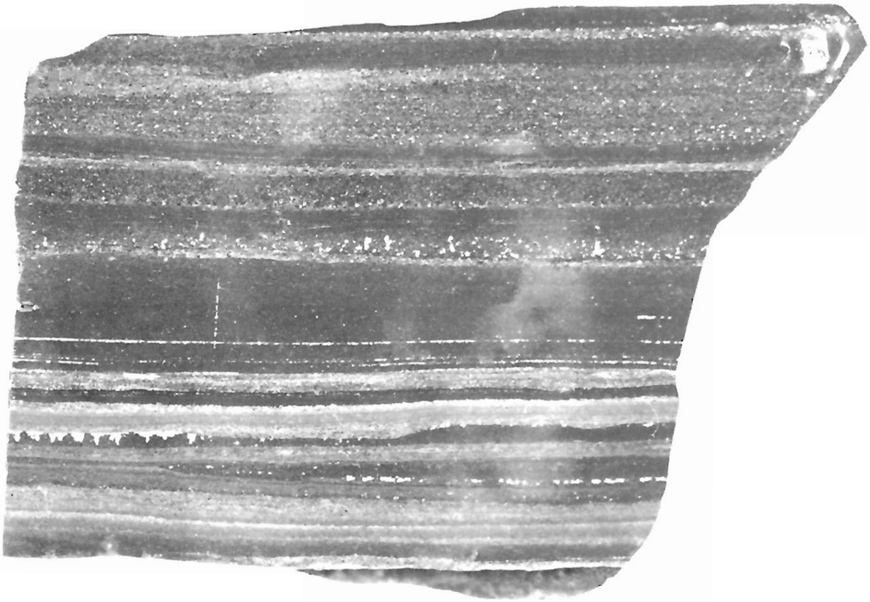


Figure 20. Pyritic tuffaceous shale in silicate-sulphide facies of member I (Dolly Ridge). Arrows point to layers with 30 to 50 per cent pyrite.

Figure 21. Tuffs in member I. Specimens A and B are from Dolly Ridge; C is from just west of Gill Mine Ridge. Graded bedding and massive character are typical of the tuffs.

Tuffs and tuffaceous rocks have not been previously described in this part of the stratigraphic sequence (Ruth or lower Sokoman) in the Knob Lake area. Frarey (1961) mentions that the shales (Ruth Slate) are in places tuffaceous, but does not describe them. The tuffs are confined to the central part of the area (Fig. 18) and occur at several irregularly spaced horizons in the lower 35 to 40 feet of member I. They are light greenish grey to dark grey rocks that stand out among the interbedded shales as massive, more competent beds. The beds are single or composite (Fig. 21), generally less than 2 inches thick, and have sharp contacts with the adjoining beds. Graded bedding, revealed by differences in colour and size of the particles (Fig. 21) can be recognized in most tuffs. Large particles (1/4 to 1/16 inch) at the bottom of the beds grade rapidly upwards, commonly in less than 1 inch, into silt- or clay-size material.

The remarkably well-developed graded bedding, the absence of cross-bedding, and the association with laminated shales, which show no evidence of normal currents sufficiently strong to transport the abundant coarse particles of the tuffs, strongly suggests that the tuffs were deposited by turbidity currents.

Fine-grained portions of the graded beds cannot be differentiated from ordinary carbonaceous shales or argillites, but microscopic examination of the coarse layers leaves little doubt about the volcanic parentage of the beds. The coarse fractions are composed mainly of feldspathic and chloritic volcanic fragments (30 to 50 per cent) and potassium feldspar (20 to 35 per cent). The feldspathic fragments, made up of small feldspar laths and fine-grained chlorite, exhibit ophitic, porphyritic and trachytic textures. The chloritic fragments are very fine grained and are so crowded with dust-like inclusions that they appear grey to nearly opaque in thin section. X-ray powder patterns of this fine-grained material correspond to a mixture of chlorite with smaller amounts of potassium feldspar. Approximately half of the chloritic fragments have a sponge-like texture and are identical to many of the vesicular-amygdaloidal volcanic rocks in the Nimish area. Other chloritic fragments are massive to mottled oval aggregates which may be accretionary lapilli.

The large feldspars (0.5 to 2.5 mm) in the tuffs are seen in thin section as subhedral, blocky or embayed crystals or as sharp, angular fragments of such crystals. Vacuole-like inclusions of chlorite in some of them probably represent altered inclusions of volcanic glass. Under crossed polars many of the feldspars have a distinct feathery mosaic texture which is similar to the texture of some of the feldspar in the Nimish volcanic rocks. The size, shape and quantity of the feldspar crystals is comparable to the "orthoclase" (Sauve, 1953) in the crystal tuffs of Astray Lake.

Quartz which occurs as angular or rounded grains is a minor (3 to 5 per cent) but persistent constituent of the tuffs. It was probably picked up from the Wishart Quartzite at the source of the tuffs.

Carbonate Facies (I:C)

Distribution and thickness of the carbonate facies are shown in Figure 22. These rocks are close analogues of the carbonate facies of the Lake Superior region. They are composed mainly of alternating layers of chert and chert-carbonate.

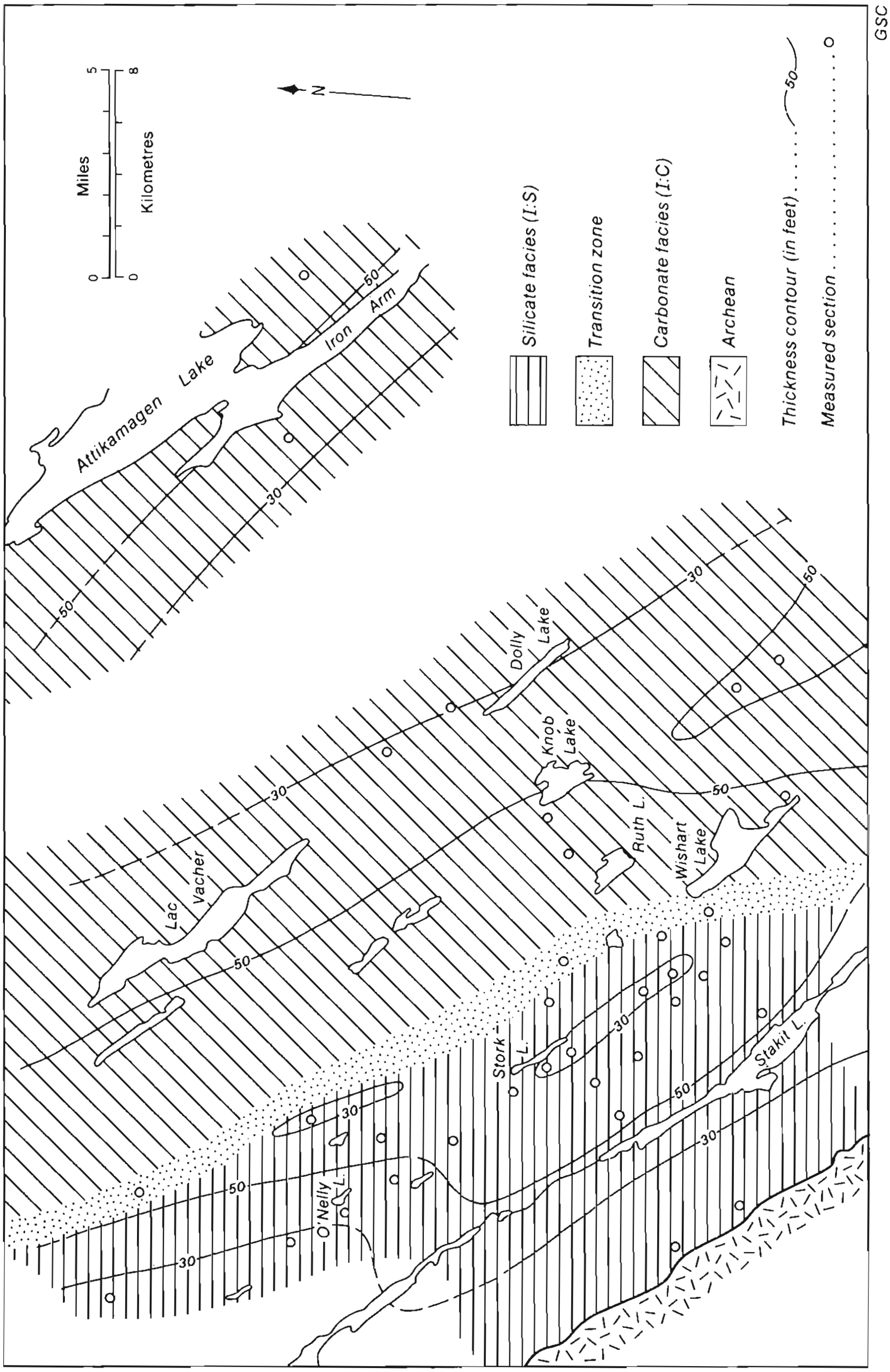


Figure 22. Distribution of carbonate facies (I:C) and silicate facies (I:S).

A banded to laminated appearance is typical of the rocks (Figs. 19, 23). In general, the beds are less than one-half inch thick, parallel, and well defined, but some are finely corrugated by stylolitic indentations and small irregularities caused by deformation and compaction of the originally soft sediment. Massive chert beds up to 4 inches thick are minor but conspicuous interbeds in some of the rocks. Laminated chert-carbonate rocks, in places with very thin partings of shale or carbon, are most common east of the Stakit Lake area. Many of these develop good cleavage parallel to the bedding when partly decomposed, and in badly broken up outcrops are difficult to distinguish from shales. Shingle-like exposures of such rocks are commonly referred to as "slates" and in some areas have been previously mapped as such. The name "slate", however, is a misnomer as the rocks have no slaty cleavage and in many places contain little or no argillaceous, clastic material.

Where fresh, the chert layers are dark grey to black, but are nearly white when abraded or altered. The ordinarily grey carbonate-rich beds are in various shades of brown on weathered surfaces – yellow-brown if stained by limonite to nearly black if coated by manganese oxides. In the vicinity of soft iron ore deposits the predominant colour is red. The weathering and alteration colours are very similar to those of the closely associated shales, and in the mines, where both are strongly altered, it is impossible to differentiate between them.

Thin sections of the chert layers show, as a rule, a uniform microcrystalline mosaic of quartz grains. Some chert layers are speckled by finely disseminated carbonate and what may be carbonaceous material. In others, wisps of these impurities outline thin (less than 0.5 mm), parallel and commonly undulating laminae. The intervening carbonate bands contain 20 to more than 95 per cent of fine-grained siderite, small amounts of ankerite(?) and disseminated carbonaceous material. Iron oxides are absent except in weathered or altered rocks. Silicates are common only in the transition zones between the adjacent facies.

Silicate Facies (I:S)

The silicate facies of member I forms a distinct unit to the west of the transition zone shown in Figure 22. The transition zone indicates areas where the silicate and carbonate facies are so closely interbedded with one another that they cannot be differentiated.

The iron-formation comprising this facies is green to grey-green, thin bedded and composed mainly of chert, minnesotaite and variable amounts of siderite and magnetite. Siderite is most common in the lower and magnetite in the upper part of the unit. The green colour of chert and the yellow-orange weathering of minnesotaite-rich beds is characteristic of the unit (I:C) and in the field distinguishes it from the associated carbonate and oxide facies (I:C and I:O respectively). Similar silicate rocks occur in two other units of the Sokoman Formation (III:S and VI:S, or the so-called "SCIF" and "YMIF" of the ore zone), but their stratigraphic position and geographic locale are distinctly different.

The bedding in the silicate facies (I:C) is on the whole thicker and more irregular than in the previously described chert-carbonate rocks (I:C). Many of the irregular minnesotaite-rich beds are granular. The granules, 0.3 to 2.0 mm in size, are composed chiefly of minnesotaite. Chert of

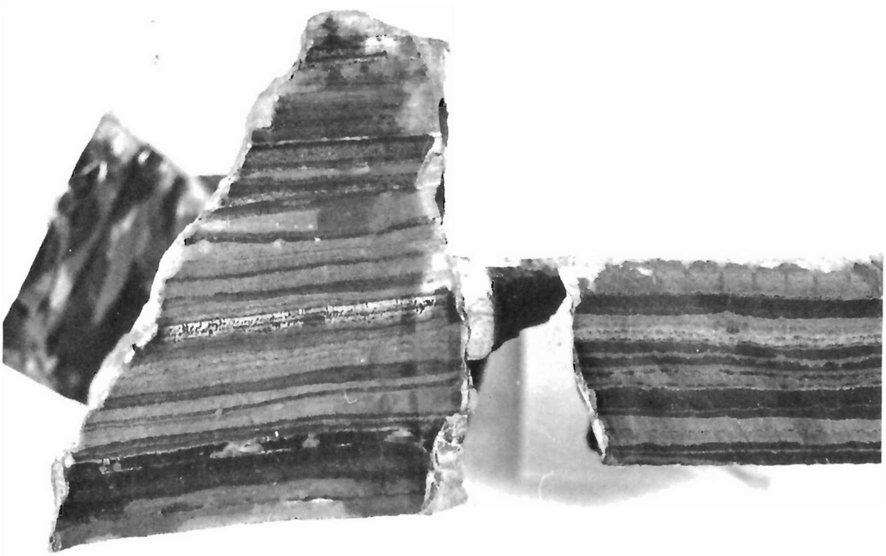


Figure 23. Fresh surface (cut and wetted) of carbonate facies (I:C). The light layers are composed mainly of siderite, and the dark layers, almost entirely of chert. Specimen on the left is from Stakit Lake area; on the right from Dolly Ridge.

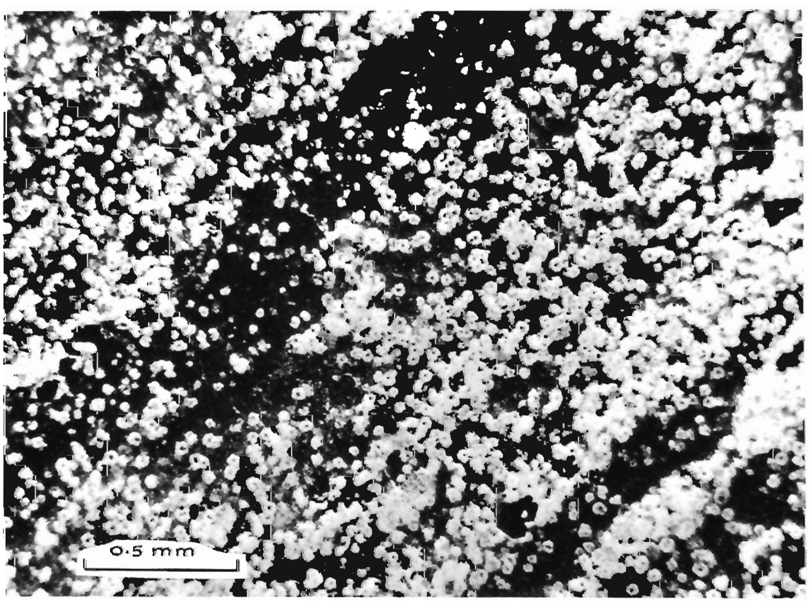


Figure 24. Typical chert spherites in jaspilite, Stakit Lake area. Most of the black material is hematite.

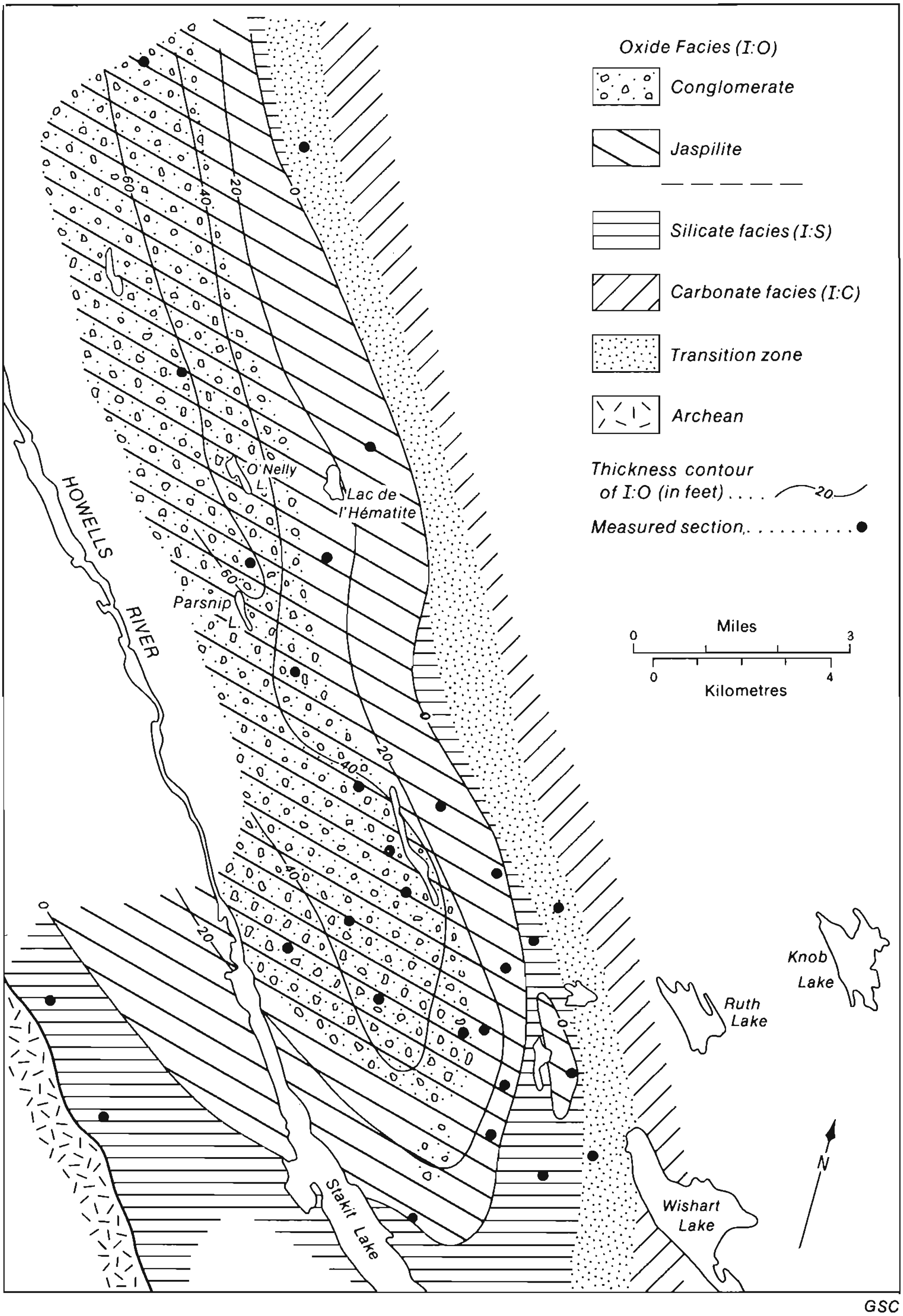


Figure 25. Thickness and distribution of oxide facies (I:O).

variable grain size is the most common cement of the granules. The fine-grained siderite occurs mainly in nongranular beds with chert of minnesotaite. Rocks made up essentially of alternating beds of magnetite and green chert characterize the upper part of the facies in many places. The rocks are thinly bedded, nongranular, and contain little or no carbonates. The silicates which impart the green colour to the rocks occur as disseminated, fine-grained to nearly submicroscopic inclusions in chert.

Oxide Facies (I:O)

This unit contains the most easily recognizable rocks in the entire Sokoman Formation. They are mainly of two types: jaspilites and pisolitic pebble conglomerates.

The red, banded jaspilites which make up the bulk of the facies are unmistakable. With the exception of small areas near Gilling Lake, where somewhat similar jaspers appear in member IV, the jaspilites are confined to the northwestern part of the Knob Lake area (Fig. 25) and are not repeated again in any other part of the stratigraphic sequence.

The jaspilites consist largely of alternating red layers of chert and steel grey layers of iron oxides. The chert layers are usually 1/4 to 1 inch thick, microcrystalline, and for the most part finely laminated. The laminae, spaced a fraction of a millimetre to several millimetres apart, stand out clearly on wetted surfaces and are one of the identifying characteristics of the jaspilites. They are fairly even and roughly parallel to the bedding, but small irregularities are not uncommon. In places sets of these laminae converge or sharply intersect one another forming what may be micro-cross-bedding.

The oxide bands in the jaspilites are generally less than 1/2 inch thick and are composed mainly of magnetite (in various stages of martitization) and smaller amounts of fine-grained hematite. The bands commonly pinch and swell or bifurcate; in places they coalesce and enclose lenticular to angular, disjointed layers of jasper.

In thin section, the most conspicuous feature of the jaspilites is the abundance of spherites (Fig. 24) which make up 20 to 70 per cent of many of the jasper beds. The spherites are clear, nearly spherical patches of microcrystalline, massive chert outlined by densely packed dust-like hematite inclusions of the otherwise similar chert matrix. Most spherites are 30 to 45 microns in diameter and contain minute inclusions of hematite or magnetite near their cores. The layers without spherites are even grained and massive or patchy due to irregular distribution of iron oxide inclusions, or to variable chert grain size. Chert-filled microfractures are numerous.

In addition to the bright coloured jasper, the jaspilites also contain grey to pink interbeds. The grey, brown-weathering interbeds (most common in the area between Stork and O'Nelly lakes) consist of chert, siderite, magnetite and small amounts of hematite. They are just as thinly bedded as the jaspilites but lack the fine laminae and spherites of the jaspilites. The pink-grey beds are composed almost entirely of chert and iron oxides (mainly magnetite). They are distinctly coarser grained and thicker bedded than the jaspilites (beds are up to 4 inches thick) and contain numerous chert-iron oxide granules.

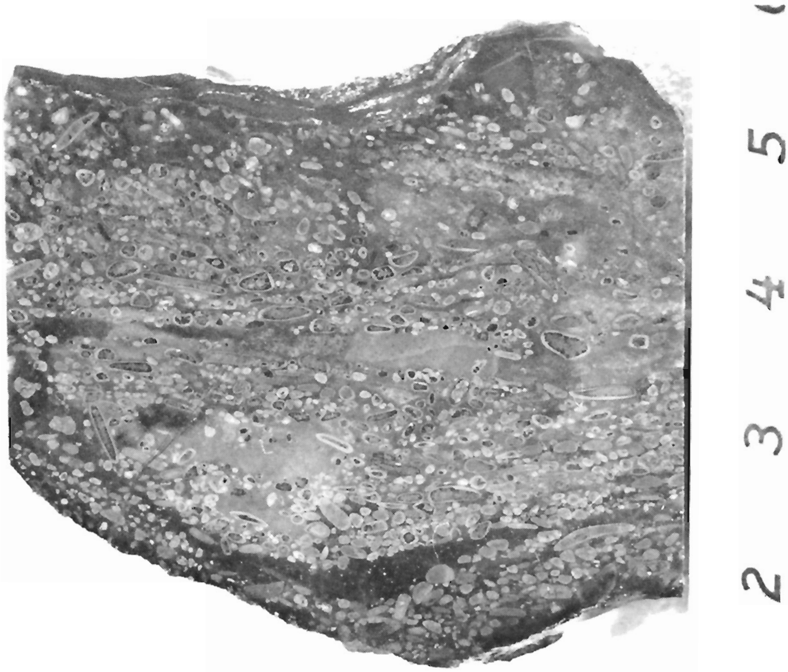


Figure 26. Jaspilite pebble conglomerate in the oxide facies of member I. Specimen from the central part of Stakit Lake area.

The conglomerates are restricted to the thickest part of the facies - the areas approximately subtended by the 35-foot contour (Fig. 25). They constitute one prominent horizon at the very top of the facies and also a few widely separated lenses in the upper two-thirds of the unit.

These conglomerates are as distinctive as the jaspilites. They are composed mainly of jaspilite fragments set in a pink or grey to nearly white chert matrix (Fig. 26). Most fragments are oval to tabular and well rounded, but some are irregularly shaped or subangular. Fragments with numerous chert-filled syneresis cracks are common. Many fragments show a vague concentric zoning defined by 2 to 3 shells of chert and iron oxides or by diffuse colour gradations that die out towards the centres of the particles. The majority of fragments are less than 1/2 inch in size, although a few slabby fragments are as much as 2 1/2 inches in length. Small-scale crossbedding and crudely developed graded bedding are seen locally in some of the conglomerates.

The extent of the upper conglomerate, which occurs along a very sharp contact between the oxide facies of member I and member II, is shown in Figure 25. The conglomerate ranges in thickness from less than 1 inch to a maximum of 2 feet. The pisolitic, pebbly texture and monolithic composition are typical of the conglomerate, except in a few places where it grades into granular, grey to pink chert-iron oxide beds with a few scattered, irregularly shaped jaspilite lenses. The thinness, distribution and composition of the conglomerate suggest that it is a wave-worked lag conglomerate. The presence of irregularly shaped, syneresis-cracked fragments indicates that not

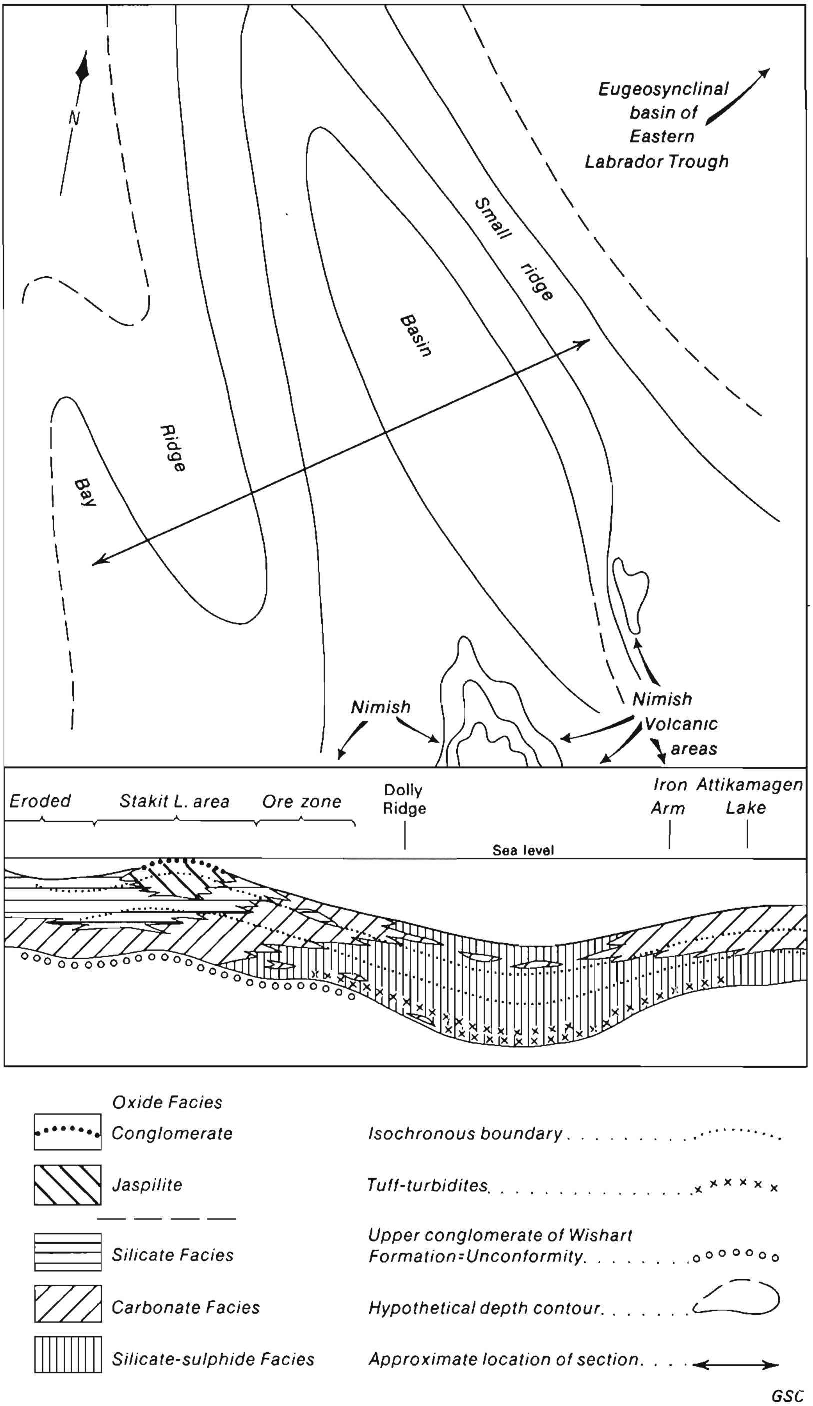


Figure 27. Interpretation of the depositional environment of member I.

all of the jaspilite was consolidated at the time when the conglomerate formed. Curved, tabular fragments, on the other hand, may be desiccation fragments that formed during temporary exposure of the jaspilite to air.

Depositional Environment

The shape and distribution of the mineralogical facies outline an elongate northwestward-trending basin (Fig. 27) the deepest portions of which correspond to the present area between Dolly Lake and Attikamagen Lake, and the shallowest, to the jaspilite areas north of Stakit Lake. Textural characteristics of the facies agree with this interpretation. The finely laminated shales occupy the central and the deepest part of the basin. They show practically no evidence of wave or current activity. On the other hand, the jaspilite conglomerates (the only conglomerates in member I) occur in the shallowest part of the basin where the current and wave action were strong enough to rework and round the jaspilite fragments of the conglomerates.

The distribution of jaspilites and associated conglomerates delineates a northward-trending, elevated area, probably a nearshore ridge, which partly isolated a shallow bay (west of the present Stakit Lake) from the main basin. In the east the increasing abundance of carbonate facies suggests an elevated area, possibly a small ridge, as the eastern flank of the basin (Fig. 27). The evidence, however, is not as conclusive as for the jaspilite ridge. The southeastern part of the basin was very likely enclosed, at least in part, by the active volcanic centres of the Nimish area. The basin also shallowed northwestward as in the area of lac Le Fer member I consists mainly of silicate and silicate-carbonate iron-formation rather than of pyritic, carbonaceous shale.

The mineralogical composition of member I was probably determined mainly by the oxidation-reduction potential (Eh) of the depositional environment. The change from silicate-sulphide through carbonate and silicate to oxide facies indicates the change from deep, reducing to increasingly shallower and more oxidizing environments of deposition. Physical energy played only an insignificant role. The predominance of very thin bedding and the scarcity of granules, oolites and intraformational conglomerates attest to the low physical energy of the environment during deposition of member I.

Source of Clastic Material

A number of previous investigators have considered the possibility that the black shales and coarser clastics at the base of the Sokoman Formation are tuffaceous. In the final analysis, however, most of them have indicated or implied that these sediments were derived from the same source as the Wishart Quartzite which they overlie.

Frarey (1961) was the only one to actually state that the shales are locally tuffaceous. The present study shows that the coarse-grained interbeds of the shales are definitely tuffs.

The composition of the tuffs and their proximity to the Nimish volcanic area leaves little doubt that the tuffs were derived from that area. The transport of the coarser volcanic material was most probably accomplished by turbidity currents which were initiated on the steep slopes of the volcanic

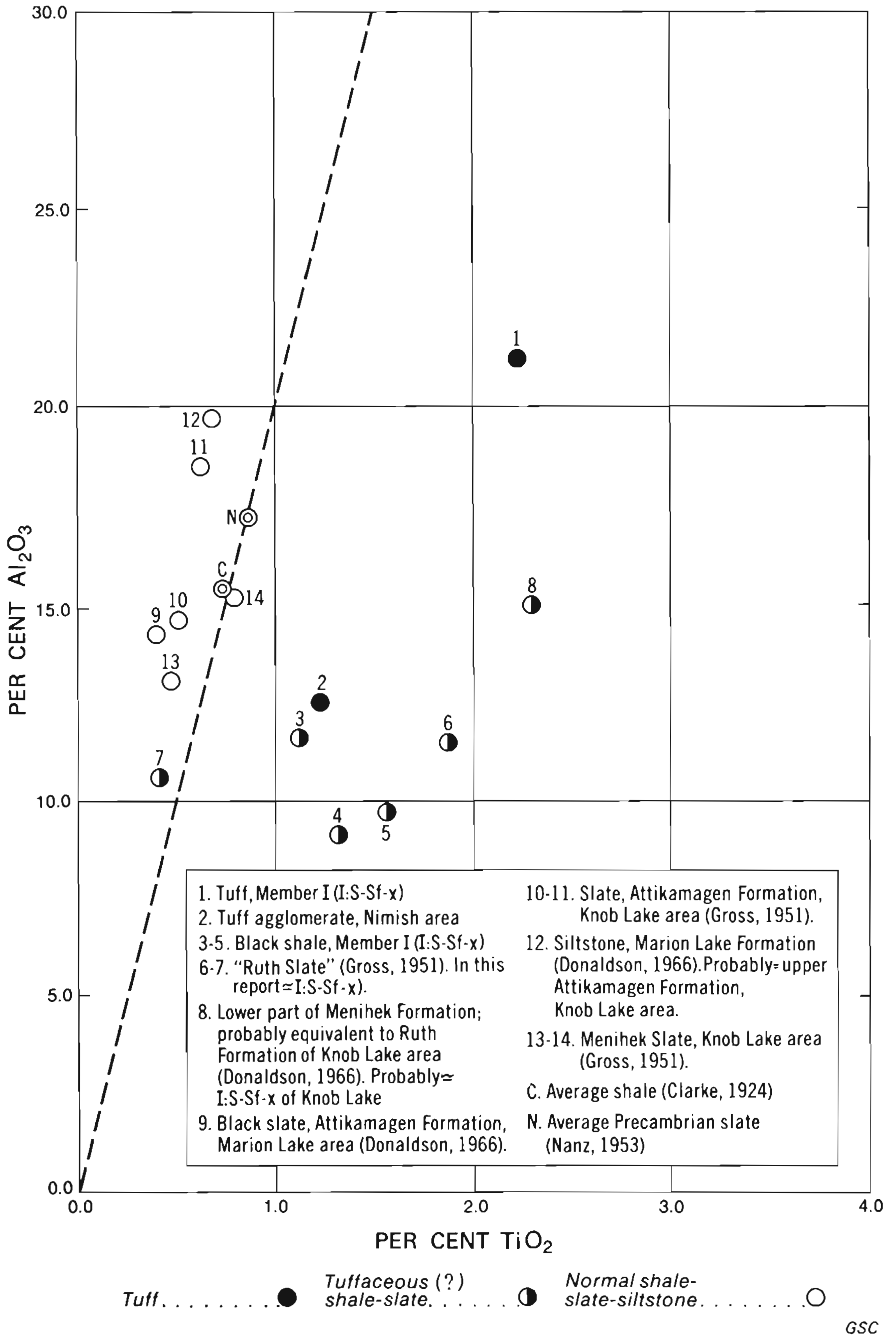


Figure 28. Relative abundance of TiO_2 and Al_2O_3 in normal and tuffaceous sediments. (Diagram after Schmidt, 1963.)

islands and descended northwestward along the flanks and axis of the basin. The finest materials may have been also transported by wind and any sluggish water currents that might have existed in the basin.

The origin of all of the fine-grained components of the shales cannot be definitely established. Since coarse-grained tuffs are present, however, it is only reasonable to assume that some of the fine-grained material is also tuffaceous. A comparison of the relative abundance of TiO_2 and Al_2O_3 in tuffs and normal argillaceous sediments (Fig. 28)[1] strengthens the argument. As Figure 28 shows, the relative abundance of Al_2O_3 and TiO_2 in the Attikamagen and Menihek sediments, which are generally believed to have been derived mainly from the Archean craton, is distinctly different from the Al_2O_3-TiO_2 content of the tuffs and of most of the shales in member I. Since most of the shales are so similar in this respect to the tuffs, the logical conclusion is that the clastic material of the shales has been derived largely from the same source as the tuffs; namely, the volcanic terrain of the Nimish area.

Some fine-grained clastic material may have been derived from the craton. If any fine-grained sediment originating in the craton was introduced into the Knob Lake area from the northwest, the clay-size particles could have bypassed the comparatively higher energy areas of the offshore ridge and the embayment west of it, then settled in the central, secluded part of the basin.

MEMBER II

Member II has not been studied in detail. Its composition and stratigraphy are only superficially known.

Carbonate-rich iron-formation is the dominant lithology of the member northeast of Ruth Lake. The iron-formation is typically laminated or evenly bedded, nongranular and composed mainly of chert and carbonate with smaller quantities of silicates. Minnesotaite-rich beds are prominent in a few places. Magnetite is scarce or absent. With the exception of a few outcrops in the ore zone (Gill Mine Ridge), the rocks are nonmagnetic. In the area, which extends from Ruth Lake to the jaspilite conglomerate outlined in Figure 25, the texture and composition of member II is variable, especially along the periphery of the jaspilite area. Despite local complexities there is a general trend from banded carbonate-silicate rocks in the east to irregularly bedded silicate-oxide rocks in the west. North of Stakit Lake, where member II directly overlies the jaspilite conglomerate of member I, the iron-formation is made up almost wholly of irregular, 1- to 8-inch-thick, greenish grey to pink-grey beds. These massive-looking but invariably granular beds consist of chert and iron oxides locally with subordinate amounts of silicates. The facies relationships in member II, on the whole, agree well with the previously determined outlines (Fig. 27) of the sedimentary basin. The thin-bedded carbonate-rich rocks occupy the deepest part of the basin, whereas the irregularly bedded oxide facies occur in the shallowest part of the offshore ridge, the area underlain by the jaspilite conglomerates.

---

[1] This method has been used successfully by Schmidt (1963) to distinguish between tuffaceous and normal sediments in the Cuyuna Range of Minnesota.

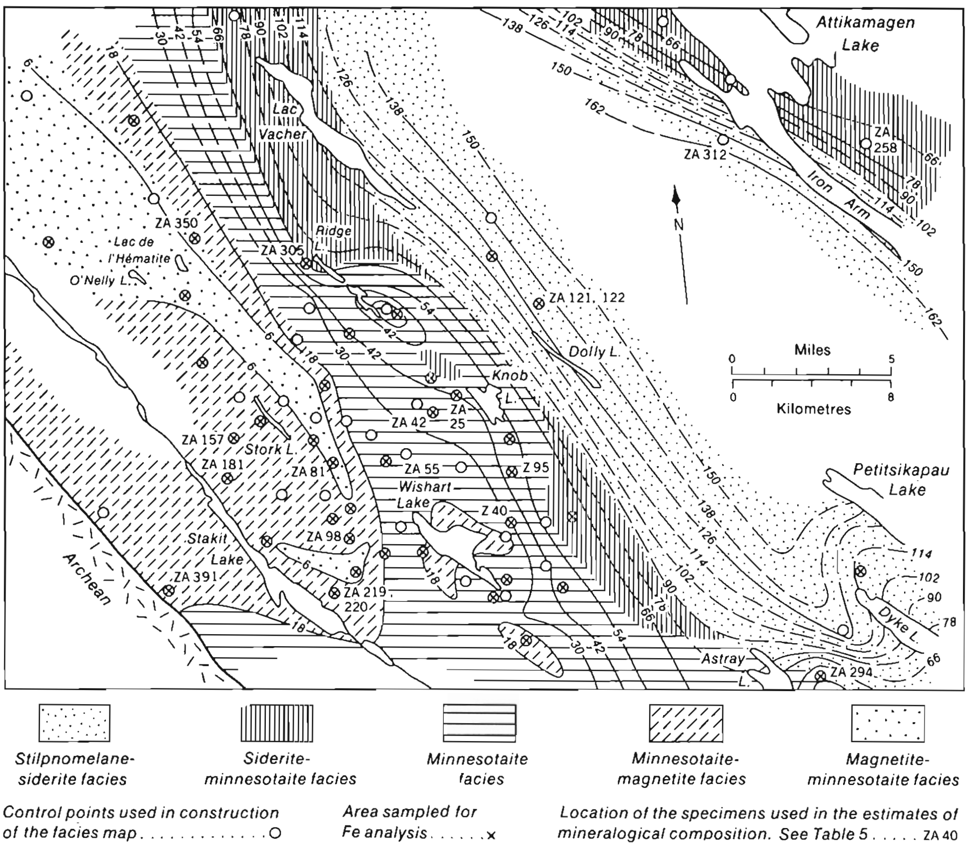


Figure 29. Facies map of member III. The contours indicate the average number of beds per foot of the iron-formation.

Figure 30. Outcrop of the stilpnomelane-siderite facies on the ridge between Dyke and Petitsikapau Lakes. Note the thin bedding, brown weathering and finely crackled appearance (indicated by arrows) of the rocks.

MEMBER III

Member III is composed mainly of thin-bedded iron-formation which is characterized by the abundance of silicates, particularly minnesotaite. The mineralogical and textural changes within the member are gradual but distinct when examined in detail over large areas. In the Knob Lake area it is possible to subdivide the member into at least five laterally gradational units as shown in Figure 29: stilpnomelane-siderite facies, siderite-minnesotaite facies, minnesotaite facies, minnesotaite-magnetite facies and magnetite-minnesotaite facies.

The thickness of the member ranges from less than 30 feet in the Stakit Lake area to as much as 80 or 90 feet in the area between lac Vacher and Dolly Lake. In the area near Attikamagen Lake the thickness could not be measured accurately because of complex structure and the scarcity of outcrops, but probably does not exceed 75 feet.

Stilpnomelane-Siderite Facies

The stilpnomelane-bearing rocks underlie an area which has been almost completely ignored since the very early investigations of this part of the Labrador Trough by company geologists. They occupy the east-central part of the Knob Lake area (Fig. 29) and are best exposed on Dolly Ridge between lac Vacher and Dolly Lake. The northwestern boundary of the

Table 5

Analyses of shale and tuff from member I and other analyses selected for comparison

Facies	Stilpnomelane-siderite				Siderite-minnesotaite		Minnesotaite						Minnesotaite-magnetite					Magnetite-minnesotaite	
Specimen Numbers	ZA-122T	ZA-121TP	ZA-312TP	ZA-294TP	ZA-305TP	ZA-258TP	ZA-293T	Z-95TP	Z-40T	ZA-25TP	ZA-45TP	ZA-55T	ZA-157TP	ZA-98TP	ZA-182T	ZA-219TP ZA-220T	PLGTP	ZA-81TP	ZA-350TP
Chert	79	2	3	2	48	56	56	34	43	5	15	2	61	62	81	68	60	55	62
Hematite										Tr.						Tr.			5?
Magnetite	?	1	3	4	?	1?	2	3	2	17	19	1?	17	15	8	11	28	41m	22m
Minnesotaite	?	10	5	53	6	17	41	58	55	77	63	96	19	23	9	8	11	4	5
Greenalite		5?	?	?											2?	3	1?	?	?
Chlorite																			
Stilpnomelane	14	53	39	16															
Siderite	6	28	49	24	45g?	25	1?	5			3	1?	3			9a			6g?
Pyrite	?	Tr.	Tr.	Tr.	?	?													
Ilmenite		Tr.	Tr.	Tr.															

Composition of each specimen represents the average of visual estimates obtained from thin section (T) and polished sections (P).

Location of the specimens is shown in Fig. 29.

m = mainly martite
g = carbonate partly altered to goethite
a = siderite and ankerite

stilpnomelane-siderite facies is not known, although these rocks probably do not extend more than 5 to 10 miles beyond the limits of the Knob Lake area. Baragar (1967) does not mention any stilpnomelane-bearing rocks in the Wakuach Lake area, nor could the writer find any such rocks in the exposures between lac Hayot and lac Le Fer.

Fresh specimens of the stilpnomelane-rich rocks are light to dark green, but become brown after a few months exposure to air. In outcrop (Fig. 30) the rocks are brown locally with a tinge of yellow-orange. The yellow-orange colour is attributed to the weathering of minnesotaite. A finely crackled appearance formed by numerous ramifying cross-fractures is typical of strongly weathered outcrops (Fig. 30). The rocks are very fine grained and thinly layered. Some resemble shales except that they do not possess good cleavage.

Stilpnomelane, siderite and chert are the dominant minerals of the stilpnomelane-siderite facies (Table 5). The amount of minnesotaite is variable. Stilpnomelane occurs as microcrystalline, green (brown when weathered) felty intergrowths with fine-grained siderite, chert and minnesotaite. Magnetite occurs as small irregularly disseminated grains. Most of it is found in the upper 10 to 20 feet of the member. The other iron oxides, hematite, goethite and limonite, are restricted to altered or weathered rocks.

The stilpnomelane-siderite facies is the only unit of member III that contains ilmenite. The ilmenite is present in the form of minute elongate particles and as lath-like intergrowths in some of the magnetite grains Figs. 31, 32).

Siderite-Minnesotaite Facies

The siderite-minnesotaite facies forms northwestward-trending zones which flank the area occupied by the stilpnomelane-bearing rocks (Fig. 29).

The siderite-minnesotaite facies is a distinctly more cherty, tougher, lighter coloured and somewhat thicker bedded iron-formation than the stilpnomelane-siderite facies. It consists mainly of nongranular, even, alternating layers composed essentially of chert and siderite or of chert and minnesotaite. Magnetite is scarce or absent except for the very top of the unit where it is present in sufficient quantities to make some of the beds magnetic.

The predominant colour of fresh rocks is grey rather than green. In outcrop the cherty, commonly porcellaneous-looking layers are yellow to nearly white. The minnesotaite-rich layers are yellow-orange, whereas the sideritic beds are in various shades of brown. In glacier-polished outcrops some of the goethite-rich sideritic layers have a metallic appearance.

Minnesotaite Facies

The minnesotaite facies of member III is probably the best known silicate facies of the Sokoman Formation. It is the type example of the unit which is known to the geologists of the Iron Ore Company of Canada as SCIF (Silicate-Carbonate Iron-Formation). The facies can be distinguished by the thin but irregular bedding, the yellow-orange weathering (Fig. 35) and the abundance of minnesotaite.

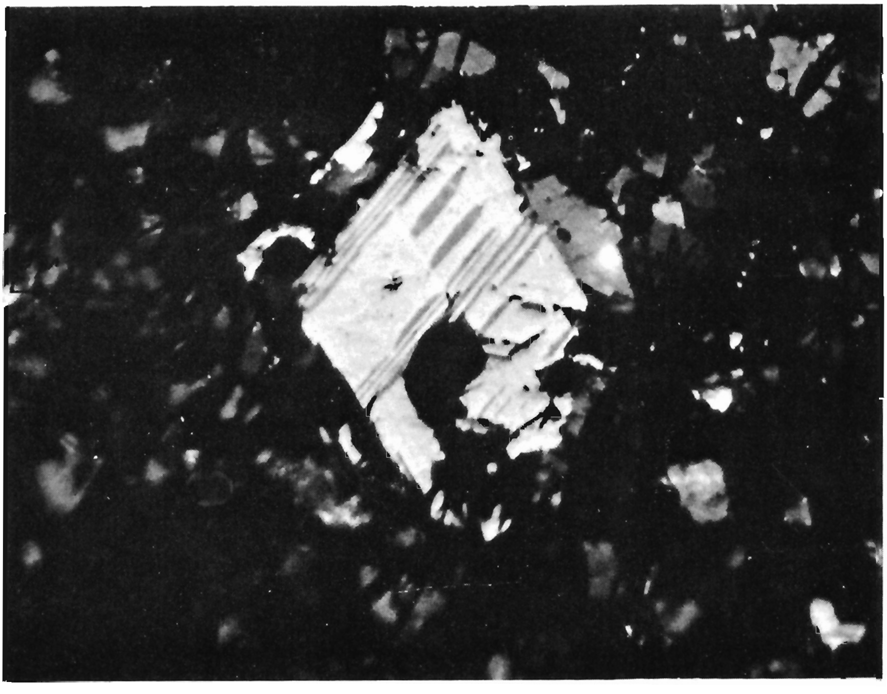


Figure 31. Ilmenite laths in one of the titaniferous magnetite crystals in the stilpnomelane-siderite facies. Reflected light.

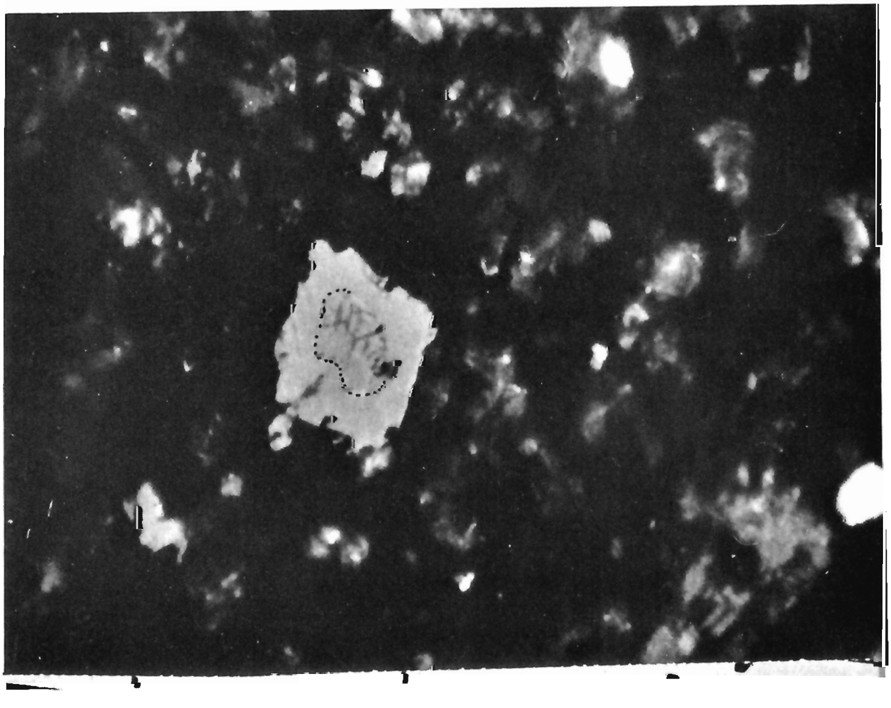


Figure 32. Oval grain of titaniferous magnetite surrounded by ilmenite-free magnetite. Stilpnomelane-siderite facies. Reflected light.

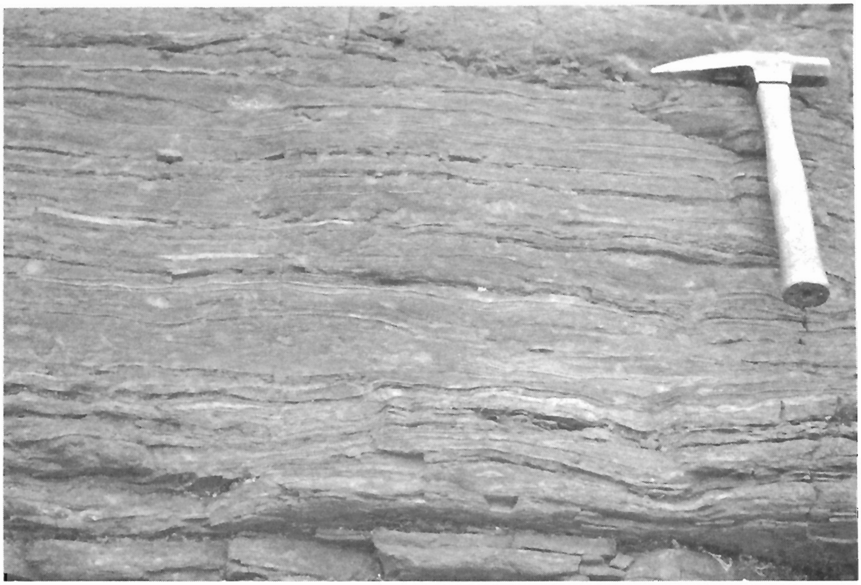


Figure 33. Irregularly banded outcrop, typical of the minnesotaite facies in the eastern part of the ore zone. Knob Lake Ridge.

Figure 34. "Thick-and-thin" bedding characteristic of the minnesotaite-magnetite facies. Wishart Lake area. .

The notable feature of the bedding in the minnesotaite facies rocks is their variable thickness and undulating irregularity. The bedding ranges from predominantly banded type (Fig. 33), in the eastern part of the area underlain by the minnesotaite facies, to predominantly "thick-and-thin" banded type (Fig. 34) in the western part of the area. Individual beds pinch and swell, merge with adjoining beds or pinch out sharply along strike. In a few places some of the thin layers transgress the thicker beds in a way suggestive of crossbedding. In the upper few feet of the member the bedding becomes less distinct and highly irregular. Slabby, angular to subrounded fragments in some of these beds indicate a more vigorous reworking of the sediments by waves and currents. The average thickness of bedding in the minnesotaite facies increases westward as shown in Figure 29, and in many places also from the bottom to the top of the unit.

Minnesotaite is the characteristic mineral of the facies and in most places the only silicate. It commonly accounts for more than 50 per cent of all minerals in the lighter coloured beds, but is also present in smaller amounts in the darker magnetite-rich layers. The minnesotaite occurs as felty masses, as radiating intergrowths with chert and as granules with or without chert. The granules are most abundant in the thicker more irregular beds. The very thin bands and laminae are usually nongranular.

Magnetite probably makes up no more than 10 to 15 per cent of the minnesotaite facies as a whole. It is most common in the upper part of the unit where its abundance may locally approach 30 per cent. Most of the magnetite is restricted to thin beds a fraction of an inch thick. It occurs as massive layers or as euhedral crystals interspersed with various amounts of chert and minnesotaite. The magnetite-rich layers, like the minnesotaite beds of similar thickness, contain very few or no granules.

Although the name Silicate-Carbonate Iron-Formation (SCIF) is commonly applied to the minnesotaite facies, carbonate is but a very minor constituent of the unit as a whole. Most specimens from the middle and upper part of the unit contain very little or no carbonate.

Minnesotaite-Magnetite Facies

The minnesotaite-magnetite facies which underlies most of the Stakit Lake area (Fig. 29) is composed of alternating thick- and thin-bedded units such as those shown in Figure 34. The thicker beds are predominant, particularly in the central and northern part of Stakit Lake area where the layered interbeds are very thin.

The layered interbeds resemble the banded minnesotaite facies (Fig. 33) except that in most places they contain more magnetite than minnesotaite. The thicker more irregular beds, commonly 2 to 4 inches thick, are composed mainly of chert with subordinate amounts of disseminated minnesotaite and magnetite. Greenalite was identified in several specimens from the Stakit Lake area. The greenalite occurs as deep green, microcrystalline, irregularly shaped felty patches in some of the chert-minnesotaite granules or as microscopic elongate particles uniformly disseminated in chert.

Granules are numerous in the thicker beds and in many places can be recognized without magnification. Small scale crossbedding is also seen in some of the thicker beds.

Figure 35. Thick, massive-looking beds of the magnetite-minnesotaite facies. The lack of thinly layered interbeds and a grey colour is typical of the facies. Bar Lake, Stakit Lake area.

Magnetite-Minnesotaite Facies

The magnetite-minnesotaite facies is mineralogically and texturally similar to the minnesotaite-magnetite facies. It differs from the minnesotaite-magnetite facies in being thicker bedded, of lighter colour and in containing more iron oxides than silicates.

The magnetite-minnesotaite facies is predominantly grey, distinctly cherty, granular and almost entirely massive bedded (Fig. 35). The weathering colours in outcrop are light grey to very pale yellow. The yellow-orange colours, so characteristic of weathered minnesotaite-rich rocks, appear locally as streaks in some of the beds and along bedding planes.

The thickness of bedding in most of the magnetite-minnesotaite facies ranges from 2 to 10 inches, but thicker beds are not unusual. The beds are locally streaked by or separated from one another by thin (less than 1/2 inch), irregular and discontinuous layers made up entirely of iron oxides or of chert, magnetite and minnesotaite in various proportions.

Magnetite and hematite are the main iron-bearing minerals. They appear in the form of dense massive bands, as oval to irregular aggregates up to several millimetres in size, or as fine disseminations. Diffuse, irregularly streaked concentrations of iron oxides are also seen within some of the beds. In outcrop the dominant iron oxide is hematite, particularly in the area extending from east of Stork Lake to lac de l'Hématite. Polished sections, however, show that most of the hematite is not primary but formed by partial to complete replacement of magnetite.

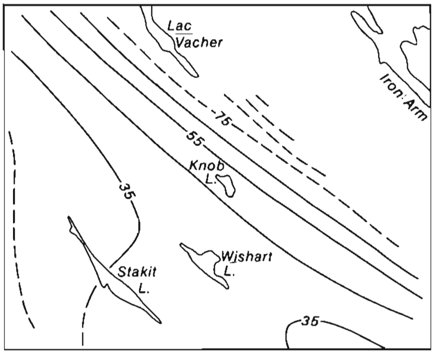


Figure 36a
Thickness (in feet) of member III. Trend surface accounts for 61.4% of the total sum of squares of the variate.

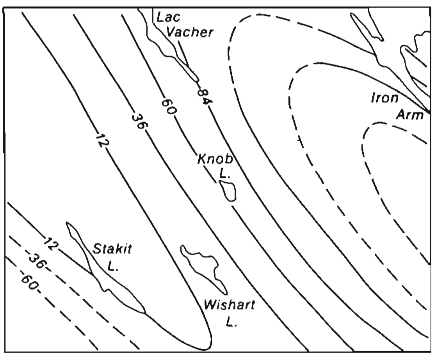


Figure 36b
Thickness of bedding (in number of beds per foot). Trend surface accounts for 85.8% of the total sum of squares of the variate.

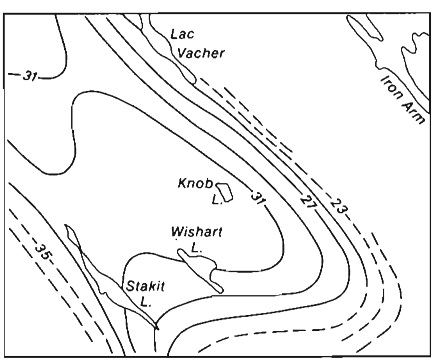


Figure 36c
Per cent total Fe. Trend surface accounts for 45.3% of the total sum of squares of the variate.

GSC

Figure 36. Trend surface maps (linear + quadratic + cubic surfaces) of member III.

Quantitative Measurements of Variability

Thickness of Bedding

Traversing the Knob Lake area in a northwesterly direction one is not likely to notice any major changes in the iron-formation of member III. On the other hand, in westerly traverses (as from Dolly Ridge to Stork Lake, a distance of only 8 miles) one cannot fail to be impressed by the mineralogical and textural changes in the rocks. The most noticeable change and one which is the easiest to express quantitatively is the change in the thickness of bedding.

The contour maps which show the average thickness of bedding in member III (Figs. 29, 36b) are based on the average number of layers (beds, bands, laminae) in one vertical foot of the iron-formation. The counts were made in outcrops by marking off a distance of 1 foot to 3 feet normal to the bedding and then counting the number of layers within that distance. Five to ten such counts were made in randomly selected areas of outcrop in the lower, middle and upper part of the member at each of the control points (indicated as circles in Fig. 29). To avoid the transition zones of member III between the adjoining members II and IV, the beds in the upper 2 to 5 feet and lower 5 to 15 feet of the member were not counted. The total number of beds counted at each locality divided by the total number of feet in which the counts were made was taken as the average number of beds per vertical foot of member III at a particular control point.

The average figures show no highly erratic variations and can be easily contoured by hand (Fig. 29). In general, the thickness of bedding increases outward from the east central part of Knob Lake area, being thinnest in the areas underlain by the stilpnomelane-siderite facies and the thickest in the areas occupied by the magnetite-minnesotaite facies. This trend stands out even more distinctly in the computer contoured trend surface (Fig. 36b).

Two other variables, the distribution of granules and the grain size of chert, should be mentioned in conjunction with bedding. Although no statistical study of these variables has been made, their variability would probably resemble the regional pattern exhibited by the changes in the thickness of bedding.

The granules are a fraction of a millimetre to several millimetres in size and are composed mainly of chert and silicates. Magnetite is present also in some of the granules. The granules are most common in the irregular thicker beds. They first appear in abundance in some of the irregular, "wavy" beds of the minnesotaite facies and become especially numerous in the predominantly thicker bedded minnesotaite-magnetite and magnetite-minnesotaite facies. In general, as the thickness and irregularity of bedding in member III increase, the amount of granules also increases. Similar generalization can be made about the grain size of chert. Comparing thin sections of thinly layered, nongranular specimens with those of strongly granular rocks, it is obvious that in most of the granular specimens the chert is distinctly more variable in texture and on the average coarser grained.

Iron Content

Forty-one composite samples were analyzed for total iron. Each composite sample consisted of a handful of chips taken approximately every

5 feet from top to bottom of member III at each location (shown by x's in Figure 29). The samples from each locality were combined, crushed and analyzed in the laboratories of the Iron Ore Company of Canada in Schefferville. Figure 36c) shows the regional trend in variation of iron in the sampled area.

Interpretation

The trend surfaces of the thickness of member III, the thickness of bedding and the contents of iron are presented together for comparison in Figure 36. Note the coincidence of low values on the three contour maps in the area north of Stakit Lake.

The area north of Stakit Lake corresponds approximately to the off-shore ridge on the western flank of the Knob Lake basin, which was outlined by the jaspilites and pebble conglomerates of member I (see Figs. 25 and 27). The ridge, being more elevated than the surrounding areas, was likely to be within easy reach of waves and currents which could rework and redistribute the accumulating sediments of iron and silica. The thicker bedding and the abundance of granules in the area north of Stakit Lake are attributed to the higher energy environment in this area during deposition of member III. As the intensity of turbulence decreased with increasing depth of water, the sediments became progressively less granular as well as thinner and more even bedded.

The lesser thickness of member III in the area of the offshore ridge suggests that some sediment has been removed from the current- and wave-worked ridge.

The lower content of iron on the ridge suggests that iron was removed preferentially to silica. Iron has apparently remained longer in a more finely divided state than silica. The finer particles would have been retained longer as suspension in the turbulent waters and were more likely to have been removed by currents and transported to the deeper less turbulent environment. The somewhat higher content of iron north of Wishart Lake and northwest of Knob Lake may be due to such preferential "winnowing out" of iron from the higher energy area of the offshore ridge and subsequent deposition in the deeper, quieter area to the east.

Depositional Environment

The mineralogy and sedimentary textures of member III agree very well in the interpretation of the sedimentary environment. Both indicate the deepest part of the Knob Lake basin to be in the area between Knob Lake and Iron Arm and the shallowest area to be north of Stakit Lake (refer to Fig. 29). The area near Attikamagen Lake corresponds to the eastern flank of the basin.

The general configuration of the Knob Lake basin during deposition of members I and III is similar (compare Figs. 27 and 29). The overall depth of the basin and the details of circulation, however, were apparently different. The absence of red or even pink hematitic jaspers and of any widespread conglomerates in member III indicates that even the shallowest part of the basin, the offshore ridge in the Stakit Lake area was not as shallow during deposition of member III as it was at the time when the jaspilite of member I and its associated pebble conglomerates were formed.

The absence of carbon-rich, pyritic sediments in member III shows that the deeper parts of the basin were not as stagnant as they were during deposition of member I. The less reducing conditions in the deeper part of the basin may have been due to decreased availability of carbon, improved circulation within the basin, elevation (rise) of the depositional interface in the deeper part of the basin, or to any combination of the three factors.

The availability of carbon cannot be evaluated. The circulation within the basin, however, was improved as evidenced by the greater abundance and more widespread distribution of the thicker bedded, granule-bearing facies. The deepest part of the basin also accumulated at least 70 to 120 feet more sediment than the shallower areas in the interval of time between the end of deposition of members I and III. The deeper part of the basin was at least that much shallower at the end of deposition of member III.

The increase in the irregularity and thickness of bedding from bottom to top of the minnesotaite facies, as well as the greater amount of magnetite in the upper part and of siderite in the lower part of the unit, indicate that the basin became shallower and better oxygenated with the passage of time.

The Origin of Stilpnomelane in the Stilpnomelane-Siderite Facies

The presence of stilpnomelane in the deeper part of the Knob Lake basin is not what might be expected from theoretically derived depth-to-shore sequences or from the prevalent trends observed in the Sokoman Formation. The sequence of mineralogical changes that occur in member III in the area between the offshore ridge and the deeper part of the basin is: (1) chert-magnetite-minnesotaite, (2) chert-minnesotaite-magnetite, (3) chert-minnesotaite, (4) chert-siderite-minnesotaite. The next expectable assemblage in the sequence is chert-siderite but the assemblage that does appear is chert-stilpnomelane-siderite.

The unexpected appearance of stilpnomelane is believed to result from the reaction of volcanic ash and/or of fine clastic material, derived by erosion of volcanic rocks in the Nimish area, with the iron-rich water in the Knob Lake basin to form stilpnomelane. There are a number of reasons which prompt this conclusion.

1. The volcanic rocks in the Nimish area are the most probable source of the ilmenite and titaniferous magnetite in the stilpnomelane-siderite facies. No other minerals or textures can be traced directly to the volcanic area. However, since ilmenite and titaniferous magnetite were probably derived from volcanic rocks it is likely that some other volcanic material was also deposited along with these minerals.

2. Volcanic ash, or feldspathic-argillaceous material derived by erosion of the Nimish volcanics, would provide a ready source for the alumina necessary for the formation of stilpnomelane in an otherwise alumina-deficient environment.

3. The stilpnomelane-siderite facies is in close proximity to the Nimish volcanic area which lies just southeast of the Knob Lake basin.

4. The geographic distribution of the stilpnomelane-siderite facies is similar to the distribution of other tuffaceous sediments in the Sokoman Formation – the shales of member I and the feldspathic greywackes and shales of member IX.

5. The stilpnomelane-siderite facies of member III is similar to the thinly layered stilpnomelane-rich units in the Precambrian iron-formations of South Africa and Western Australia in which the stilpnomelane has formed apparently by alteration of waterlain tuffs (LaBerge, 1966a, b; Trendall, 1966).

MEMBER IV

Member IV is one of the best stratigraphic markers of the Sokoman Formation. The upper main part of the member is the more conspicuous. It is characterized by abundance of purple to red jasper, distinctive soft-sediment deformation structures, and in places by oolites and high iron content.

The thickness of the member increases eastward from 25 to 35 feet near the western margin of the map-area to as much as 55 to 80 feet in the area between lac Vacher and Dyke-Petitsikapau Lakes. In the Attikamagen Lake-Iron Arm area, the thickness is uncertain owing to the difficulty in determining the contact with the overlying member V which also contains abundant jasper and oolites.

The lower part of member IV is a transition zone between the oxide-silicate and silicate-carbonate facies of member III and the hematitic facies of the upper part of member IV. In the ore zone and near Stakit Lake the transition zone ranges from a few inches to several feet in thickness. It consists of purplish grey or red beds 1 to 2 inches thick containing abundant granules and, in places, poorly formed oolites. Mineralogically the transition zone is a mixture of chert, magnetite, hematite and minnesotaite. Minnesotaite is confined mainly to the granules and cores of oolites. Hematite, dust-like in appearance, is most common in the chert matrix which surrounds the granules and oolites.

Many hematite-minnesotaite oolites are unusual. The cores are oval granules composed of minnesotaite with a little chert and magnetite. Surrounding the cores are thin alternating shells of chert and dusty hematite. The chert-hematite shells rather than the cores show evidence of deformation, and in close-packed oolites, the shells are squeezed out to almost completely fill the interstices between oolites. The cores, which were clearly more solidified than the chert-hematite shells, are probably granules which were picked up by currents from the underlying member III and were coated by chert and hematite in the more oxidizing depositional environment of the hematitic member IV.

In the eastern part of the Knob Lake area, the transition zone is generally thicker and more variable in character than to the west. It ranges from jasper-bearing magnetite-hematite iron-formation near Dyke Lake and Astray Lake to thin-bedded iron-formation near Dolly Ridge, which in addition to jasper also contains beds rich in stilpnomelane, minnesotaite and locally siderite.

The transition zone on Dolly Ridge between lac de la Squaw and lac Vacher is unusual; it contains abundant siderite and iron oxides but virtually no silicates. The lower contact is here defined by a pink, granular jasper bed which contains a few irregular silicate-rich fragments that may have been derived from the underlying rocks. The rest of the transition zone consists of 1- to 4-inch beds of jasper with a few beds of sideritic iron-formation

Figure 37. Typical outcrop of the upper part of member IV in the ore zone. The colour, bedding and general appearance are characteristic of this part of the Sokoman Formation.

Figure 38. Irregular soft-sediment deformation structures common in the upper part of member IV. Outcrop near Wishart Lake.

from a few inches to 20 feet thick. The granular jasper beds are composed almost entirely of chert and hematite; the granules, which make up 25 to 70 per cent of the beds, are composed of fine-grained hematite and chert. Many granules, heavily charged with dust-like hematite, are nearly opaque in thin section. The beds of purplish (brown where weathered) sideritic iron-formation are composed of chert, siderite and magnetite with a little hematite. Unlike the jasper beds, the siderite rocks are evenly bedded and contain no granules. The individual laminae, a fraction of an inch thick, have sharp or gradational contacts with each other but are laterally uniform.

The upper, main part of member IV is an irregularly, thin-bedded iron-formation composed almost entirely of chert, hematite and magnetite. In outcrop, it appears to consist of beds of jasper and massive metallic-looking beds and lenses (Fig. 37). The jasper, composed of chert and finely disseminated hematite, is granular and in places oolitic; it occurs as irregular stringers and lenses. The "metallic" beds and lenses are composed of hematite and magnetite with a subordinate amount of chert. Minnesotaite, brown stilpnomelane and ankerite are rare accessory minerals. Crocidolite is common in only two places in the area. "Metallic" beds with 60 to 80 per cent of iron oxides are most common in the upper 10 to 20 feet of the member.

Characteristic of member IV, especially of its upper part, are soft-sediment deformation structures which range from simple pull-apart lenses (the least common type) to highly contorted stringers and lenses (Fig. 38). These structures are most numerous in the ore zone and the Stakit Lake area. A notable but less obvious feature of the upper beds is the abundance of well-formed granules and oolites. The granules are widespread, whereas the oolites are restricted to certain areas. The quantity, size and perfection of the oolites vary from place to place, but there is a definite regional trend in the distribution of the oolitic rocks (Fig. 39). Typical textures of the granular and oolitic rocks are shown in Figures 40 to 43.

In the highly oolitic zones in the upper part of member IV, the oolites are well formed and reach a size of 2 or 3 mm (Figs. 40, 43). They are megascopically visible in most of the cherty beds. The cores of the oolites are oval granules, fragments of granules or oolites (Fig. 43). The surrounding shells are made up of as many as 40 concentric layers of chert and hematite (Figs. 40, 43). Magnetite, where present, appears as irregular to crudely concentric crystal aggregates (Fig. 41). In the initial stage of oolitic growth, the shells closely follow the outline of the cores which acted as seed grains. In later stages the oolite approaches a spherical shape unless another granule or oolite was captured, in which case the oolite assumes an elongated or lobate (grapelike) shape (Figs. 40, 43). Such composite oolites make up 20 to 30 per cent of the oolites in strongly oolitic rocks.

In the weakly oolitic zones, most oolites are not well formed and composite oolites are very rare (Fig. 41). The oolites are less than 1.5 mm in size and can rarely be recognized without the aid of a microscope.

With the exception of a few jaspers (particularly common near Abel and Gilling lakes) which contain no oolites and few or no granules, most of the iron-formation in the upper part of member IV in the non-oolitic areas is strongly granular. Granules up to 1.5 mm in size, composed of chert, and iron oxides, make up 50 to 75 per cent of most of the cherty beds (Fig. 42). Despite the great number of granules, oolites appear to be completely absent from these rocks.

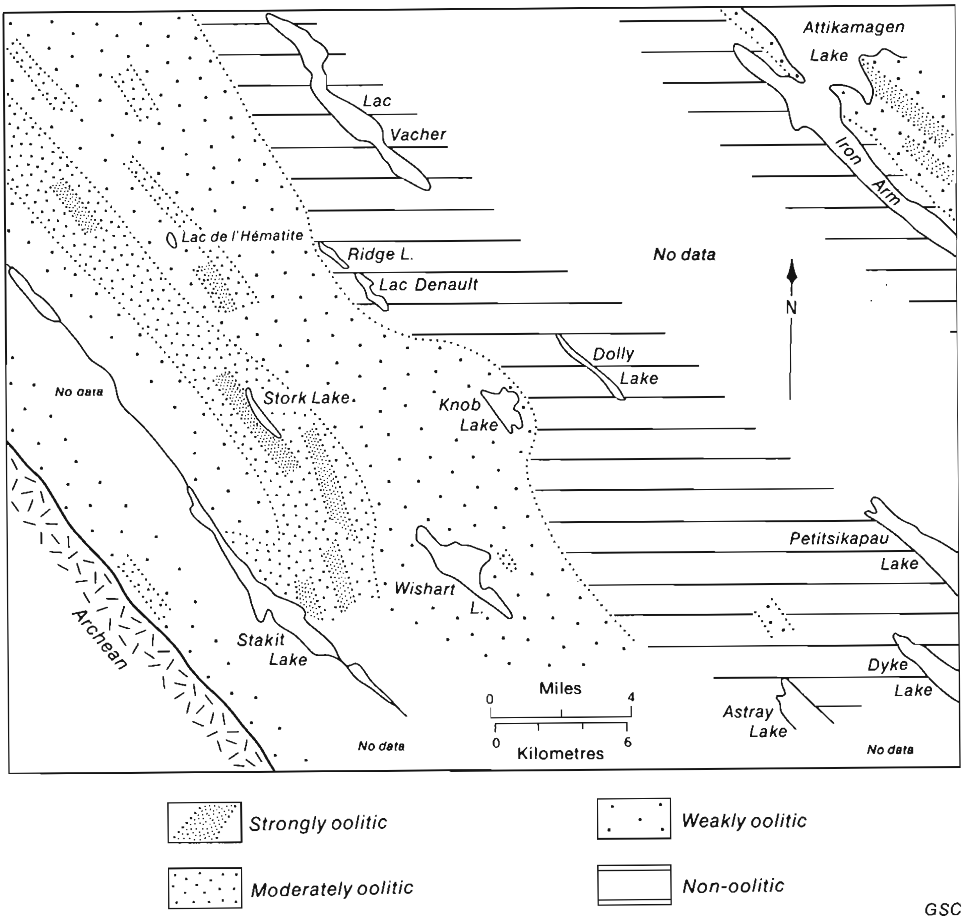


Figure 39. Distribution of oolitic rocks in the upper part of member IV.

Depositional Environment

After deposition of member III the Knob Lake basin continued to become shallower and better oxygenated. This is evident by the general increase of iron oxides and of oolitic and granular beds from bottom to top of member IV. The abundance of jasper, iron oxides, oolites and granules indicates a well oxygenated, shallow environment throughout the basin in the last stages of sedimentary history of member IV.

The general configuration of the Knob Lake basin outlined by the changes in sedimentary textures and mineralogy of members I and III can also be recognized from the distribution of oolitic rocks in the upper part of member IV. The strongly and moderately oolitic rocks were deposited just seaward of the previously deduced offshore ridge north of Stakit Lake. Eastward of this shallow, high energy area the oolites became generally smaller and fewer in number as the result of less frequent tumbling of the particulate sediment by waves and currents which dissipated into the deeper parts of the basin now represented by the granular, non-oolitic rocks.

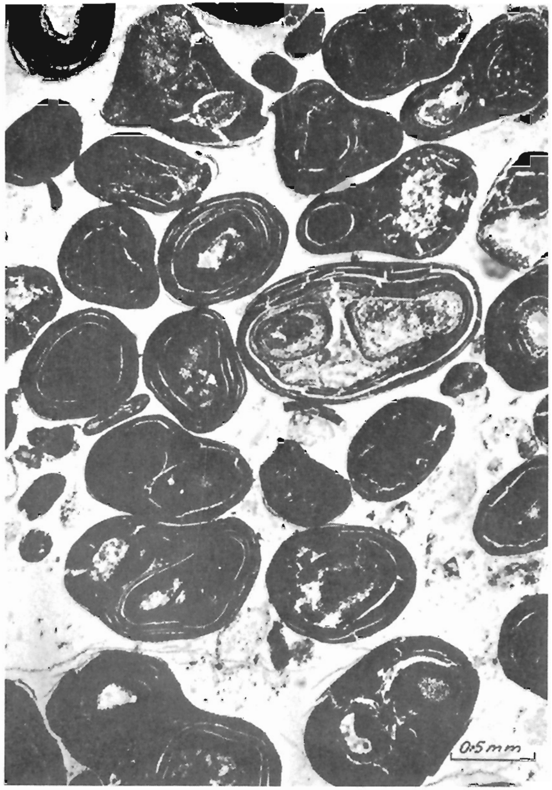


Figure 40. Hematite oolites in the strongly oolitic zone in western part of Knob Lake area (Bar Lake, Stakit Lake area). Hematite is black. Chert is white.

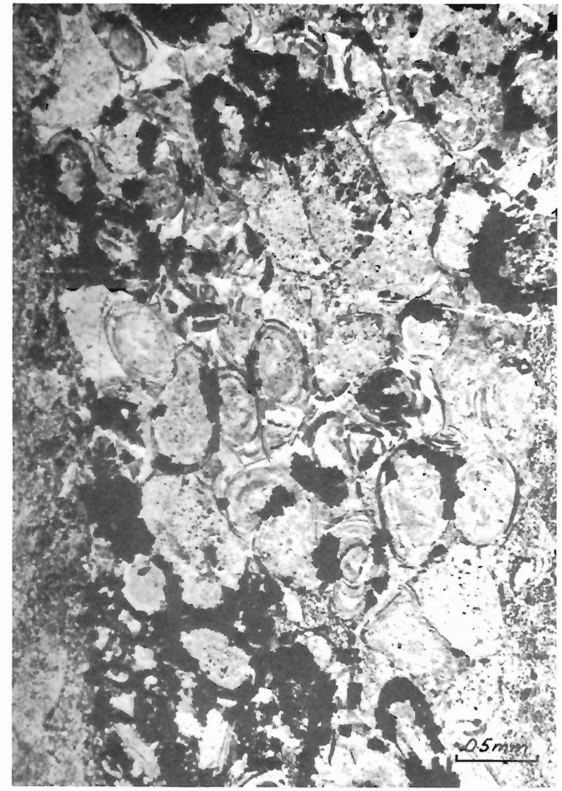


Figure 41. Poorly formed oolites in the weakly oolitic zone in eastern part of ore zone just west of Knob Lake (Knob Lake Ridge). The large crystals and crystal aggregates are magnetite. The finely disseminated opaques are hematite.

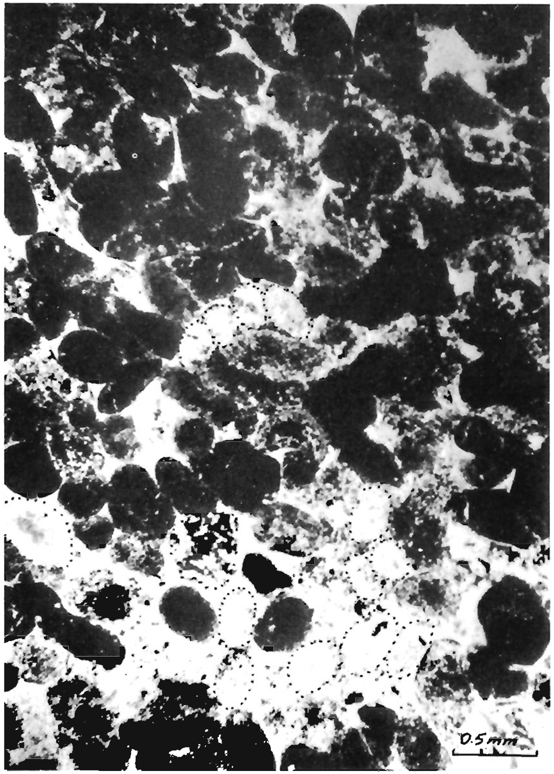


Figure 42. Granules in the non-oolitic zone in central part of Knob Lake area (Dolly Ridge). Most of black opaques are hematite. Chert is white. Note the different size, shape, texture and composition of the granules. Outlines of some of the cherty granules have been inked in.

Figure 43. Hematite oolites in the strongly oolitic zone in eastern part of Knob Lake area (Joyce Lake, Attikamagen Lake area). Note the broken off outer oolitic shell (1) and the elongate oolite (2) formed around a cuspate fragment.

MEMBER V

Member V is one of the more variable units in terms of its thickness, textures and general appearance in outcrop. Mineralogically the member is essentially an oxide facies although the mineralogical composition is variable and locally complex.

Near Attikamagen Lake member V consists of thin-bedded iron-formation which is composed almost entirely of chert and iron oxides. Red and pink jasper beds are abundant and granular, oolitic and locally conglomeratic beds are common. The abundance of jasper and locally of strongly oolitic rocks makes it difficult to distinguish member V in this part of the Knob Lake area from the underlying member IV.

In the areas just west of Iron Arm and on the ridges between Petitsikapau Lake, Houston Lake and lac Vacher, member V consists of iron-formation which is similar to some of the clastic (shale, greywacke type) iron-formation. Its distinguishing features in outcrop are dull grey to purplish brown colours and an "argillaceous" rather than cherty appearance. The only distinctly cherty rocks are a few thin layers and lenses of massive or granular jasper. Bedding in the argillaceous-looking iron-formation is irregular and generally not well defined. Granules of chert and hematite and less commonly of other minerals are numerous, although they can rarely be recognized without a microscope. The main minerals are chert, a felty to pasty material of uncertain composition and fine-grained, well-crystallized or dust-like hematite. Magnetite and brown stilpnomelane(?) are locally abundant. South of Iron Arm and southeast of Dolly Lake a small amount of feldspar and rare clastic grains of quartz are visible in some of these argillaceous rocks.

In the area between lac Vacher and Dolly Lake the member is 40 to 70 feet thick but thins rapidly southeastward so that in the area between the northern arms of Dyke and Astray lakes it is not more than a few feet thick or is absent entirely.

The best exposed and most clearly defined part of the member is in the western part of Knob Lake area (Fig. 44). Its upper contact is gradational but generally sharp and easy to locate. The lower contact throughout most of the western area is abrupt. The major exceptions are the exposures in the mines where the sharpness of the contacts is in many places obscured by alteration. The iron-formation between these well-defined stratigraphic boundaries is medium bedded to banded, grey to pink in colour and composed mainly of chert and smaller, variable amounts of magnetite and hematite. Ankerite is locally abundant. Minnesotaite is a common accessory of the grey cherty beds, but the amount rarely exceeds 10 per cent. The rare exceptions are a few yellow-orange weathering beds containing up to 50(?) per cent of minnesotaite. These are found in several places between Wishart Lake and Star Lake, near Stakit Lake, and at one locality north of lac de l'Hématite. Crocidolite is present in two places, just south of Knob Lake and east of Denault Lake. Stilpnomelane is scarce.

Differences in the mineralogical composition of member V in the western part of the Knob Lake area are common, but they are mainly small-scale variations of local rather than regional character. Regional differences in sedimentary textures are more prominent and on this basis the member can be subdivided into two laterally gradational units, which for descriptive purposes are designated as "conglomeratic-banded facies" and "massive-bedded facies" (Fig. 44).

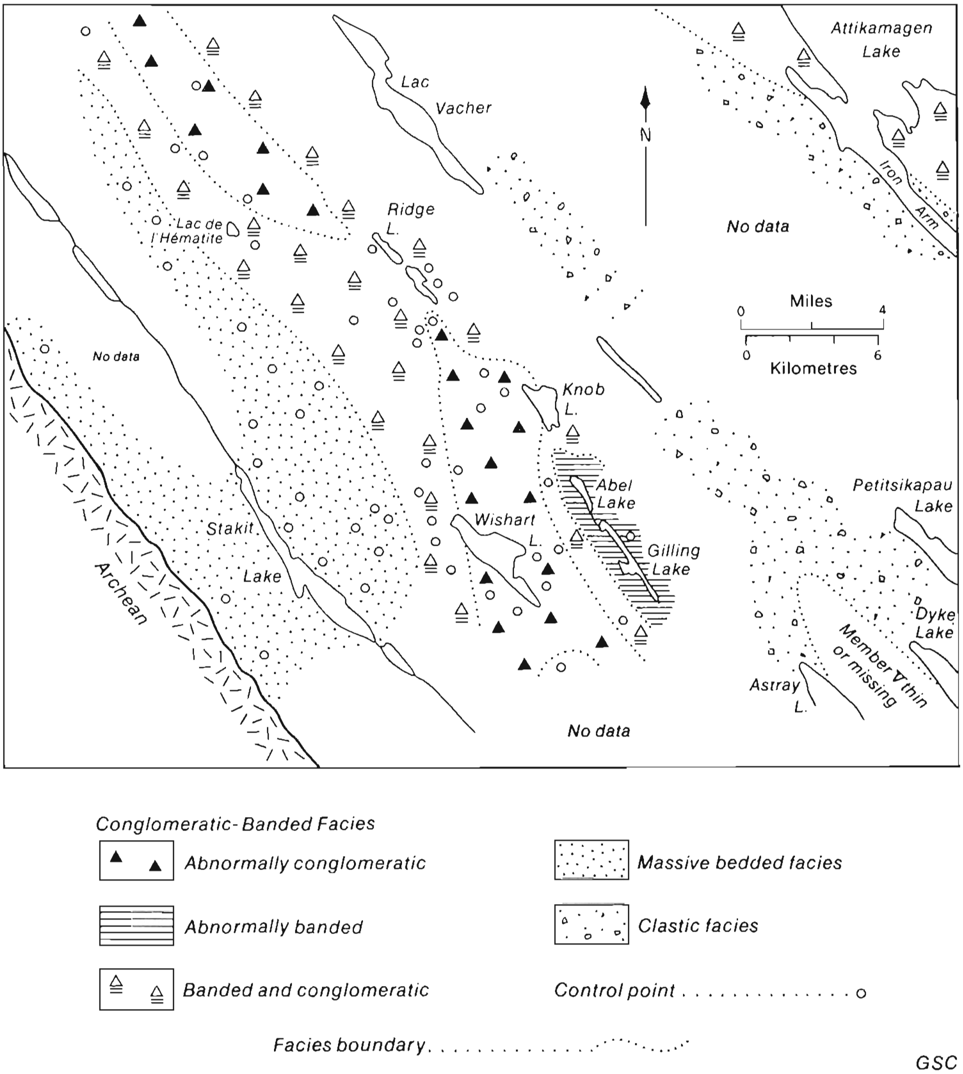


Figure 44. Facies map of member V.

Conglomeratic-Banded Facies

This unit consists of irregular, granule-bearing and commonly conglomeratic cherty beds (Fig. 45) interspersed at irregular intervals with thin "metallic" layers (bands) of iron oxides (Figs. 46, 47) and a few thin, uniform bands and lenses of chert (mainly pink jasper).

The irregular, granule-bearing and conglomeratic beds predominate the conglomeratic-banded facies. The thickness of these beds ranges from 1/4 inch to 1 1/2 feet. Beds thinner than one inch are particularly abundant southeast of Knob Lake (in the area of Hope, Abel and Gilling lakes). In other areas 2 to 8 inches is the usual range. Most fragments and pebbles in the conglomeratic beds (Fig. 45) are less than 2 inches in maximum dimension. They are angular to well-rounded slabby pieces which appear to have

Figure 45. One of the conglomeratic beds as seen in a plane approximately parallel to the bedding. Note the slabby shape of the larger pebbles. In bottom left hand corner is a remnant of one of the "metallic layers" which overlies the pebbly bed. Outcrop on Gill Mine Ridge.

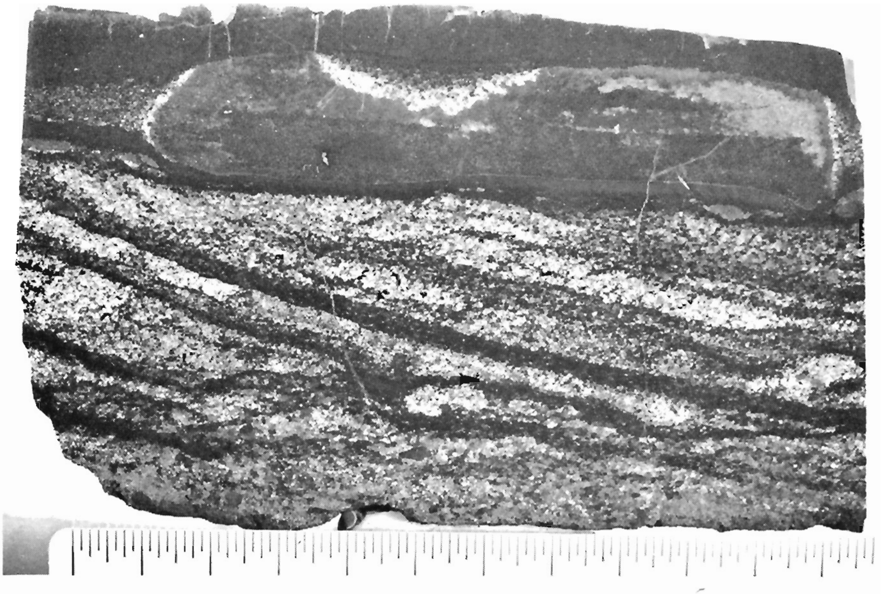
been derived by intraformational reworking of the thin layers of chert and iron oxides, the remnants of which are still present in the same general area. The fragments and pebbles composed mainly of iron oxides are more common than those made up of chert. In most outcrops east of Wishart Lake the iron oxide pebbles and fragments constitute 70 per cent or more of all recognizable pieces.

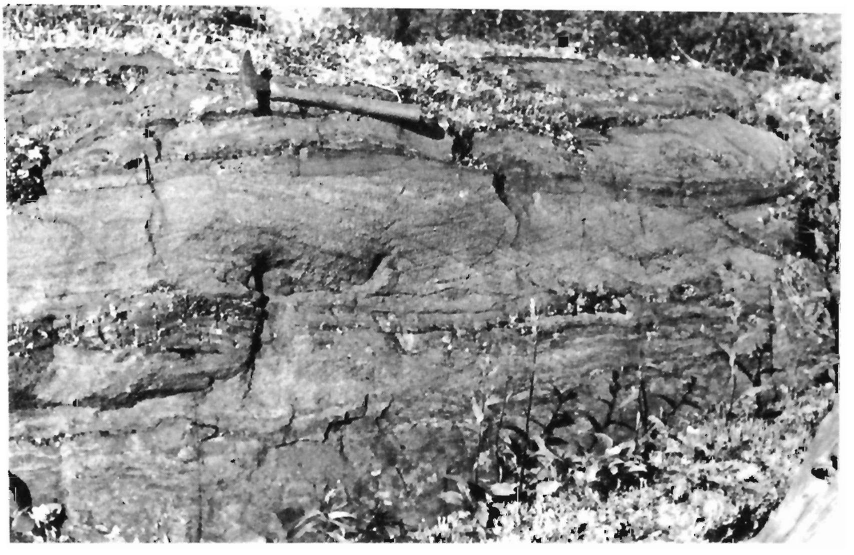
Granules are abundant in the irregular and conglomeratic beds. Most are oval, less than 3 millimetres in size, and composed essentially of chert with a smaller quantity of iron oxides. In some beds, the abundance of granules, as well as their size, shape, texture and composition, are highly variable. Oolites and fragments of oolites are present locally. Crossbedding is common, although such ideal examples as those shown in Figures 48 and 49 are rare. The irregularity of bedding, and the presence of pebbles, granules and crossbedding are all indicative of the turbulent, wave- and current-worked environment in which these sediments were formed.

The thin metallic layers, which are interbedded at irregular intervals with the granular-conglomeratic beds, are mineralogically and texturally different. They are composed mainly of iron oxides, rather than of chert, and are distinctly more uniform in texture.

The metallic bands consist largely of magnetite, hematite or a mixture of both. The total amount of iron oxides in any band ranges from 60 to more than 90 per cent. The nonmetallic component is mainly chert but carbonate and/or minnesotaite is also present in some bands. The thickness of the metallic layers ranges from 1/8 inch to nearly 2 inches. Individual bands in the predominantly thin-bedded parts of the member southeast of Knob Lake can be traced for tens of feet with little apparent change. In other areas

Figures 46, 47. Outcrops (Fig. 46 west of Ruth Lake; Fig. 47 Knob Lake Ridge) of conglomeratic-banded facies of member V showing "metallic bands" composed mainly of iron oxides, interbedded with more cherty, granular and locally conglomeratic beds.





Figures 48, 49. Crossbedding in conglomeratic-banded facies of member V west of Wishart Lake.

where the thicker bedded, granular-conglomeratic beds predominate, the metallic bands pinch out or end abruptly within a few feet. Small irregularities at the contacts of the metallic bands are not uncommon, some are probably due to local intraformational erosion and others to compaction and soft-sediment deformation. Where the bands are in contact with conglomeratic layers, they swerve around and drape over the pebbles. Similar relationships are seen under the microscope. In places the granules as well as the pebbles clearly indent the adjoining metallic bands.

Most metallic bands appear massive in outcrop. In cut specimens and especially on polished surfaces, however, many are seen to consist of two or more parallel and usually sharply defined layers of different thickness and composition. There appears to be no systematic or rhythmic arrangement in the stacking of layers within any one metallic band. The individual metallic layers are much more uniform in texture and composition than the associated granular-conglomeratic beds. Most contain no granules, oolites or any other pellet-like structures. The mineralogical and textural differences that can be recognized under the microscope are mainly gradual changes in the proportion of chert and iron oxides from top to bottom of a band. In many very thin layers the texture and mineral distribution is uniform throughout. The bedding planes that demarcate major changes in composition are as a rule very well defined (Fig. 47) even under the microscope, although they may be irregular where in contact with granular or pebbly beds. Nongranular, thin and equally well-defined bands and lenses of microcrystalline chert are locally associated with the metallic bands. The quantity of these chert layers, however, is on the whole subordinate to the metallic bands.

The absence of granules, oolites and the lateral uniformity of both the metallic and the cherty bands undoubtedly reflect intermittent periods of sedimentation in quiet, nonturbulent waters.

Massive-Bedded Facies

As the number of fragments, pebbles and thin nongranular bands decreases, member V becomes distinctly more thickly and massively bedded. The massive-bedded facies in the area indicated in Figure 44 is a monotonous assemblage of irregular, massive-looking beds in various shades of grey (Fig. 50). The beds, most of them 2 to 10 inches thick, are composed mainly of chert and subordinate amounts of hematite and magnetite. Carbonate (ankerite), the presence of which is revealed in outcrop by pitted surfaces, is locally abundant. Near Stork Lake and in the areas northwest of it where carbonate is most common, it is seen as irregular, patchy beds and lenses composed of tightly packed carbonate granules. Siderite was identified in only one specimen of an ankeritic bed west of Stork Lake. The silicates are scarce or absent except minnesotaite which is common in some of the beds near the north end of Stakit Lake.

The comparatively uniform and monotonous character of the massive-bedded facies is probably the result of a more thorough and continuous mixing of the original sediment by turbulent waters than in the areas now occupied by the conglomeratic-banded facies where the turbulence was periodic.

Figure 50. Typical massive appearance of beds in the massive-bedded facies east of Stakit Lake. Note the absence of pebbles, fragments and metallic bands. Compare with Figures 45 to 47.

Quantitative Measurements of Variability

The abundance of excellent outcrops in the western part of the Knob Lake area provided the opportunity to obtain some quantitative data on the variability of texture and composition of member V. Samples were collected for the study of three variables: (1) the quantity of pebbles and fragments, (2) the number of metallic bands, and (3) the amount of total Fe.

Sampling Method

The estimates of the quantity of pebbles and fragments and of the number of metallic bands were made in randomly selected outcrops in the lower, middle and upper parts of the member at each of the locations indicated by circles in Figure 44. To estimate the number of pebbles and fragments at a particular sampling point, a strip 6 inches wide and 1 to 2 feet long was marked off at right angle to the bedding and all particles larger than 1/4 inch within that strip were counted. The 1/4-inch size is the lower limit at which the pebbles and fragments can be easily discerned in outcrop. In all, 8 to 14 such counts were made from top to bottom of the member at each sampling locality. The counts for each locality were averaged and expressed as the number of pebbles and fragments per square foot.

The estimate of the number of metallic bands was made in the same outcrops in which the pebbles and fragments were counted. The estimates were obtained by counting all well-defined metallic bands over a distance of 1 foot to 3 feet normal to the bedding. In making the counts each metallic band was considered as a single unit, even where some internal layering could be recognized. The average number of metallic bands per vertical foot was then determined for each locality.

The samples collected for total Fe analyses consisted of chip samples taken approximately every 5 feet from top to bottom of the member at each locality. The samples were analyzed in the laboratory of the Iron Ore Company of Canada in Schefferville.

Interpretation

Regional trends in the variability of member V show up most clearly in trend surface maps which minimize the erratic variations due to sampling error and to real but local differences in the member.

The low values in all four trend surface maps (Fig. 51) coincide approximately in a northwesterly-trending area near Stakit Lake which

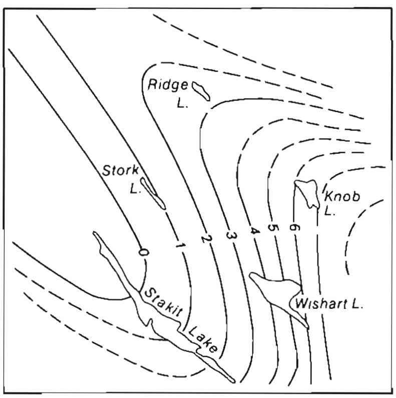


Figure 51a
Number of "metallic bands" per foot.

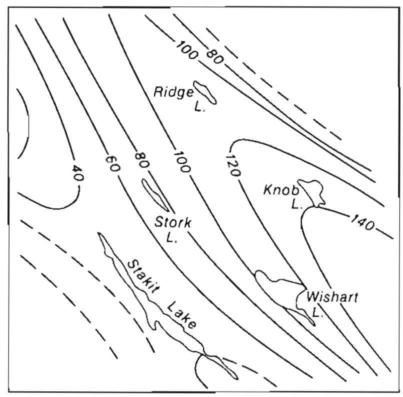


Figure 51b
Thickness (in feet) of member V.

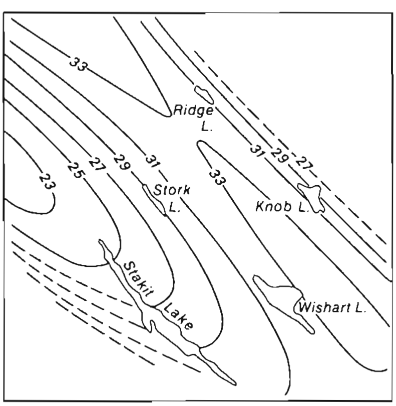


Figure 51c
Per cent total Fe.

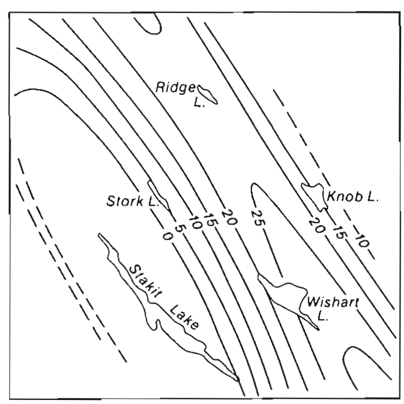


Figure 51d
Number of fragments per square foot.

Trend surface 51a accounts for 76.3% of the total sum of squares of the variate, surface 51b for 57.0%, surface 51c for 56.7%, and surface 51d for 58.1%.

Figure 51. Trend surface maps (linear + quadratic + cubic surfaces) of member V.

corresponds to the offshore ridge - a shallow, high energy area - on the western flank of the Knob Lake basin. The thin, uniform layers of iron oxides, which were probably deposited in quiet water, become more numerous outward from this area (Fig. 51a). The increasing number of these layers indicates the more secluded and likely deeper areas, less frequently accessible to strong wave and current action. The deeper northwesterly-trending area just west of Knob Lake and Ridge Lake accumulated a greater thickness of sediment than the shallow area near Stakit Lake (Fig. 51b). Some of the sediment deposited in this deeper area was probably derived by intraformational erosion from the offshore ridge. As in the case of member III, the iron has evidently been winnowed out preferentially to silica (Fig. 51c). The iron oxides, being finer grained than the granular chert, remained longer in suspension in the turbulent waters on the offshore ridge, and as a result were transported farther out into the deeper areas where they were deposited as thin uniform layers during quiet periods of sedimentation.

The pebbly-fragmental beds in member V are intraformational conglomerates which were very probably formed by periodic, local reworking of the partly consolidated sediments by wave and currents. The distribution of pebbles and fragments (Fig. 51d) indicates that the most intense, periodic wave and current action was on the seaward (eastern) slope of the offshore ridge. Field observations show that large iron-rich (metallic) pebbles and fragments are more common than silica-rich (cherty) pebbles and fragments. Hence the partial coincidence of the Fe high in Figure 51c with the conglomeratic-fragmental high in Figure 51d.

Depositional Environment

The mineralogy and sedimentary textures of member V indicate a shallow and fairly well oxygenated environment throughout most of the Knob Lake basin during deposition of this part of the Sokoman Formation. The conditions under which sedimentation occurred, however, were slightly more reducing than during deposition of the underlying member IV which contains more bright red, hematitic jasper and virtually no silicates. The more reducing environment of member V is probably due to a slight increase in the depth of water in the Knob Lake basin.

The variability of member V can be attributed to the changing configuration of the depositional interface and the resulting changes in the details of circulation within the basin during deposition of member V. Some irregularities in the shallow basin were probably created by local redistribution of the accumulating sediment, and others by structural readjustments within the basin.

The presence of the "argillaceous" clastic facies in the central, southern part of Knob Lake area can be related to the renewed uplift and erosion in the Nimish area just southeast of the Knob Lake basin (Petitsikapau-Dyke-Astray lakes area).

MEMBER VI

Member VI in most of the Knob Lake area is a jasper-bearing oxide facies which has been used for a long time by geologists of the Iron Ore

Company of Canada as a horizon marker. The unit, however, is not everywhere as simple or as distinctive as it is in the mine areas where it was first described by the company geologists and given the name Upper Red Cherty (URC).

The least distinctive and the most troublesome part of the member is in the area of Attikamagen Lake and Iron Arm. Near Attikamagen Lake the member consists of interbedded jaspers and grey to pink cherty beds which are similar in composition and appearance in outcrop to the iron-formation comprising members IV and V. In the area just west of Iron Arm the member contains, in addition to jasper beds, numerous interbeds of clastic

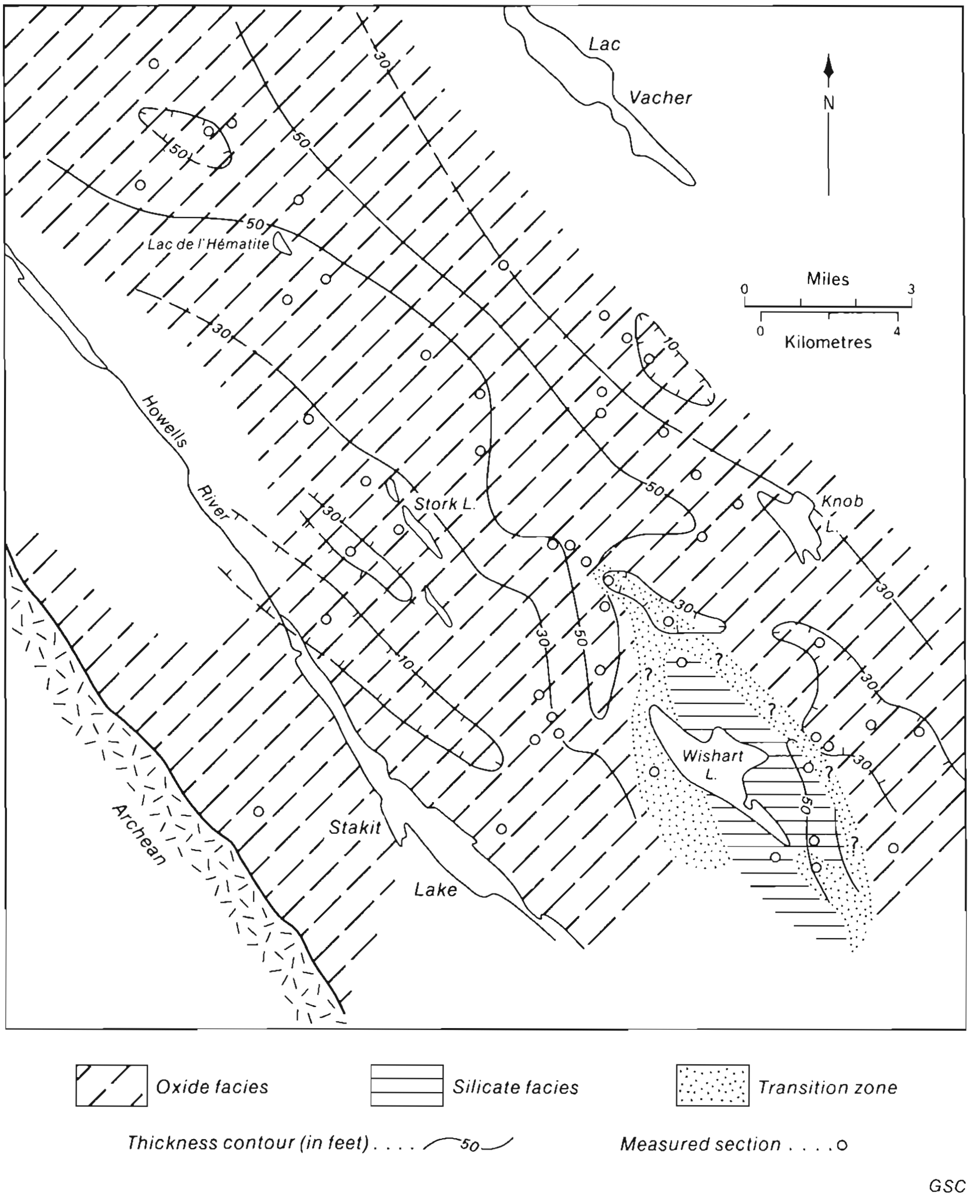


Figure 52. Facies map of member VI.

silicate-oxide facies which are difficult to differentiate from the shale-greywacke type iron-formation in the middle and upper part of the Sokoman Formation in this area.

In the area between Dolly Lake and lac Vacher, member VI is composed mainly of 2- to 4-inch-thick, granule-bearing beds of jasper. What the member is like southeast of Dolly Lake is unknown since this part of the iron-formation is not exposed. The member apparently pinches out near Houston Lake since in outcrops just over a mile southeast of this lake, members VIII and IX rest directly on member V or member IV.

In the western part of the Knob Lake area (Fig. 52), member VI is well exposed, and with few exceptions easily distinguished from the adjoining units of the iron-formation (members V and VII) with which it is in abrupt or gradational contact. This western part of member VI consists mainly of two mineralogically and texturally distinct units: a dominantly thin-bedded silicate facies and a fragmental-conglomeratic and distinctly thicker bedded oxide facies.

The oxide facies, known to the company geologists as Upper Red Cherty (URC), is a medium- to thick-bedded iron-formation (beds are commonly 1/2 to 1 1/2 feet thick) characterized by red, scattered fragments and lenses of jasper within pale pink or grey to dark metallic grey beds (Fig. 53). The predominantly grey beds are granular, although they do not always appear to be so in outcrops. The beds are composed of about equal proportions of chert and iron oxides locally with minnesotaite as an accessory. Carbonates are scarce or absent except in the area between Knob Lake and Ridge Lake and one locality near Wishart Creek where thin lenses, irregular to rounded patches and fine disseminations of carbonate (mainly ankerite) are common. Most jasper fragments are angular, but well-rounded fragments (pebbles) make up conglomeratic lenses in some of the beds (Fig. 53). The majority of fragments and pebbles are less than 2 inches in maximum dimension with a predominant slabby shape. They appear to be nothing more than pieces broken off thin jasper layers, the remnants of which are still preserved in some of the thick beds (Fig. 54).

The silicate facies (YMIF or Yellow Middle Iron-Formation in the terminology of company geologists) is composed essentially of chert, minnesotaite and smaller variable amounts of magnetite. Hematite is generally absent. Carbonates are also scarce except in the upper part of the unit where it is transitional to the carbonate facies of member VII. The bedding in the silicate rocks is thin but irregular. Beds less than two inches thick are especially numerous so that many outcrops are banded in appearance. The banded texture is accentuated by weathering and by the presence of thin layers composed mainly of magnetite. The thin minnesotaite- or magnetite-rich layers are usually uniform in texture. The thicker and more irregular beds, on the other hand, are granular. Elongate fragments and lenses of the thin layers, and rare small-scale crossbedding are also seen in some of the thicker beds. The most persistently thin-bedded rocks near the south end of Wishart Lake resemble the banded minnesotaite facies of member III Fig. 33).

The transition zone between the silicate and oxide facies consists mainly of a drab, pale yellow to grey hybrid iron-formation which combines the mineralogical and textural features of both the silicate and oxide facies. Interbedding of jasper-bearing and minnesotaite-rich rocks occurs near the northwestern margin of the transition zone and in the southeast part of the area.

Figure 53. Typical texture and massive character of the beds in oxide facies of member VI in the central part of the area between Knob Lake and Stork Lake.

Figure 54. Jasper lenses in one of the beds in the oxide facies of member VI, northwest of Knob Lake.

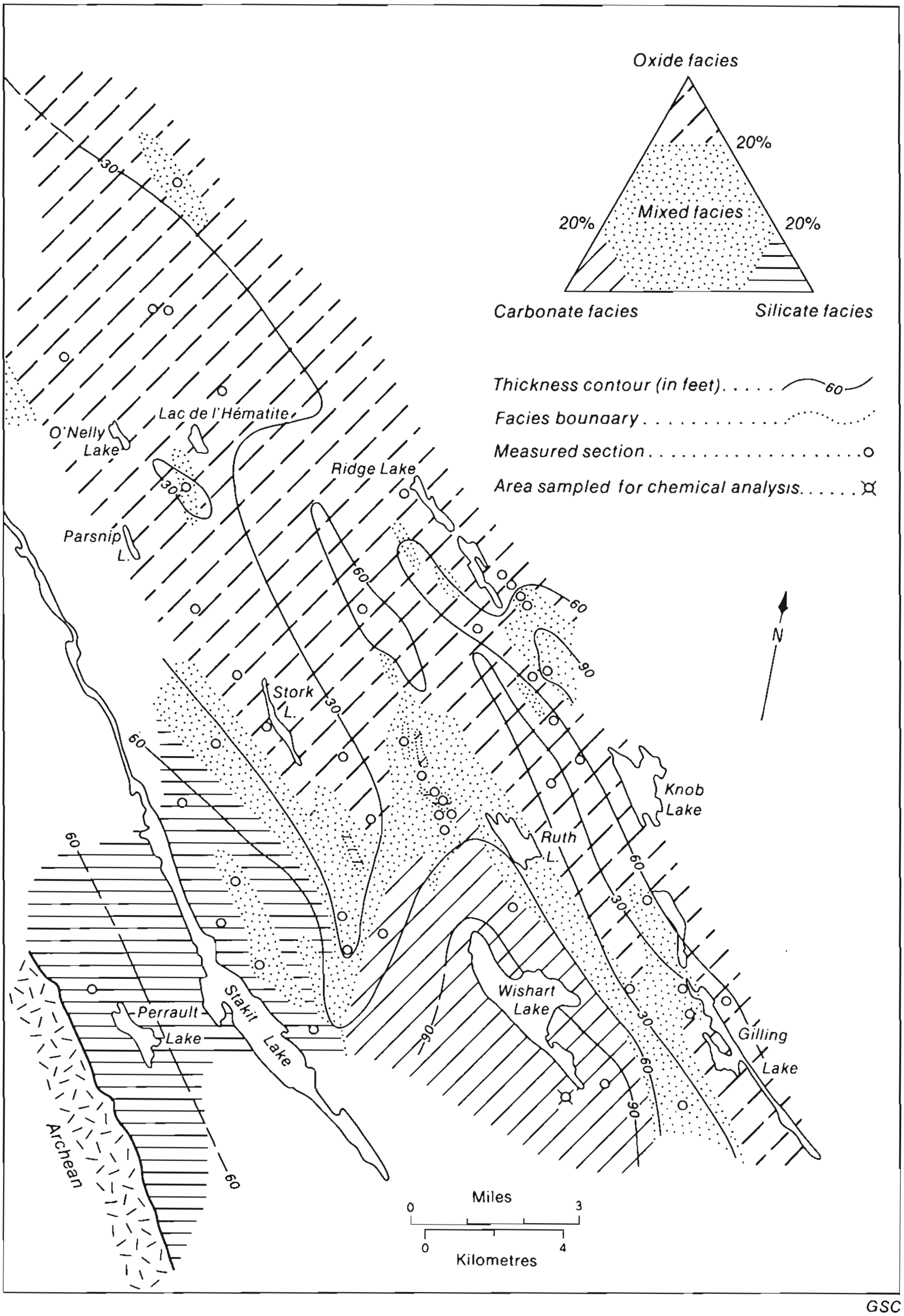


Figure 55. Facies map of member VII.

Depositional Environment

Mineralogy and the sedimentary textures of member VI all indicate a shallow, well-oxygenated environment throughout most of the Knob Lake area during deposition of this part of the iron-formation. The area underlain by silicate facies of member VI is an exception. During deposition of the member this area was probably a shallow depression in the gently sloping western flank of the basin. The deeper waters of this depression, sheltered from the open sea to the east and north by comparatively more elevated areas now represented by the conglomeratic-fragmental oxide facies of member VI, provided a favourable environment – low physical energy, moderately reducing to weakly oxidizing conditions – for deposition of the thin-bedded minnesotaite-rich rocks.

MEMBER VII

Member VII is the most variable unit of the Sokoman Formation. The mineralogical differences are not as extreme as they are in member I, but the relationships between the different facies are much more complex. Member VII was studied in detail only in the western part of the Knob Lake area (Fig. 55) where the member and the two stratigraphic markers which delimit it (members IV and VIII) are well exposed. The stratigraphically equivalent iron-formation in other areas is not well known.

In the eastern part of the Knob Lake area near Attikamagen Lake, the member is an irregularly bedded, grey to pink iron-formation composed of chert and smaller amounts of magnetite, some silicates, and locally hematite. In the area between lac Vacher and Petitsikapau Lake, the iron-formation is thin to medium bedded, predominantly grey to purplish brown and commonly argillaceous rather than cherty in appearance. In addition to chert and iron oxides, fine-grained silicates (mainly brown stilpnomelane?) and rare scattered grains of feldspar and detrital quartz are also present in the more argillaceous-looking rocks.

In the western part of the Knob Lake area, member VII consists of three mineralogically distinct but complexly interfingered facies; carbonate, silicate and oxide. The facies map (Fig. 55) shows the relative abundance of the three facies in various parts of the area. Measured stratigraphic sections (shown as circles) from which the per cent thickness of each facies was calculated provided the control for construction of the map.

Carbonate Facies

The carbonate facies of member VII is an unusual and rare type of iron-formation. Mineralogically and chemically the facies is similar to other chert-siderite rocks in Precambrian iron-formations, but texturally it is very different. Most sideritic rocks in Precambrian formations (James, 1954, 1966) are thin bedded to laminated and composed, in addition to chert, of fine-grained massive siderite. The carbonate facies of member VII, on the other hand, is thickly and very irregularly bedded, locally crossbedded and composed of medium- to coarse-grained siderite, most of which is not massive but is in granule form.

Figure 56. Typical spotted appearance and light grey weathering of the carbonate facies of member VII. Wishart Lake area. Note cross-bedding.

Figure 57. Outcrop of silicate facies of member VII in the Stakit Lake area showing the characteristic weathering colours of the rocks.

The carbonate facies of member VII is well exposed in the area shown in Figure 55. The most extensive and least altered outcrops are south of Wishart Lake in the area where the samples for chemical analysis were collected (see Fig. 55). The light grey, pitted outcrops and the thick but irregular bedding are typical of the facies (Fig. 56). Crossbedding, accentuated in outcrop by the weathering of carbonate, is common although poorly defined and not always easy to recognize.

The mineralogy of the carbonate facies is simple - chert, siderite and little else (Table 6). The siderite is medium to coarse grained and buff to grey in colour. Individual grains are up to several millimetres in maximum dimension. Thin sections show that most of the siderite is in the form of oval to irregularly shaped granules which are irregularly mottled by opaque dust-like inclusions. Most of the inclusions could not be identified, but some at least, judging by the red internal reflection, are hematite. The large anhedral siderite grains, which make up the granules as well as much of the matrix, commonly extend across the contacts of granules without a break in crystallographic continuity. The siderite obviously recrystallized without affecting or being affected by the structure of the granules.

The light grey to light greenish grey chert, which makes up approximately half of the carbonate facies, is confined largely to the matrix of the siderite. The chert also occurs as massive lenses, fragments and pebbles irregularly mixed with siderite. The silicates (mainly minnesotaite) occur as minute fibres or small fibrous clusters, most of which are uniformly disseminated in the chert.

Chemically the siderite-rich rocks of member VII are similar to the carbonate facies of the Lake Superior region (Table 7). The sample collected for chemical analysis consisted of a handful of chips taken every 5 feet to represent a true thickness of 68 feet of the middle part of the member in the area shown in Figure 55. Although care was taken to omit weathered surfaces of outcrops, the collected chips were not entirely free of secondary iron and manganese oxides formed by surficial alteration of the carbonate. It is probable that at least half of the Fe_2O_3 and a trace of MnO reported in the chemical analysis are due to such contamination.

Silicate Facies

The silicate facies of member VII is a predominantly thick-bedded iron-formation composed mainly of chert and minnesotaite (Table 6). It is distinguished from other minnesotaite-rich rocks in the Sokoman Formation by the absence of persistently thin-banded or laminated beds and by its stratigraphic position. The properties which distinguish the facies from the closely associated carbonate and oxide facies of the same stratigraphic unit (member VII) are the abundance of minnesotaite and the distinctive buff to yellow-orange weathering (Fig. 57).

Individual beds in the silicate facies are commonly 1 to 2 feet thick. Thin-bedded units are also present locally but are not typical of the facies. The minnesotaite-rich rocks are on the whole more uniform in appearance than the previously described carbonate facies. Pitted or otherwise spotted outcrops are uncommon except in the mixed facies areas (Fig. 55). Well-defined crossbedding is rarely seen.

Table 6. Mineralogical Composition of Specimens
from Member VII

FACIES	CARBONATE		SILICATE			OXIDE		
Specimens	ZA64	ZA68	Z22C	Z177	ZA252	Z262	ZA304	ZA253
Chert	57	61	26	29	50	57	57	70
Hematite		tr.		1	tr.	1	tr.	10
Magnetite	tr.?	?	8	25	11	33	37	18
Minnesotaite	tr.	tr.	55	41	33	tr.	2?	?
Stilpnomelane	tr.	tr.				tr.		
Siderite	42	38	11	4	6	8	3*	2?
Carbon	tr.?							

Composition of each specimen is the average of visual estimates obtained from one thin section and one polished section.

*Siderite and ankerite.

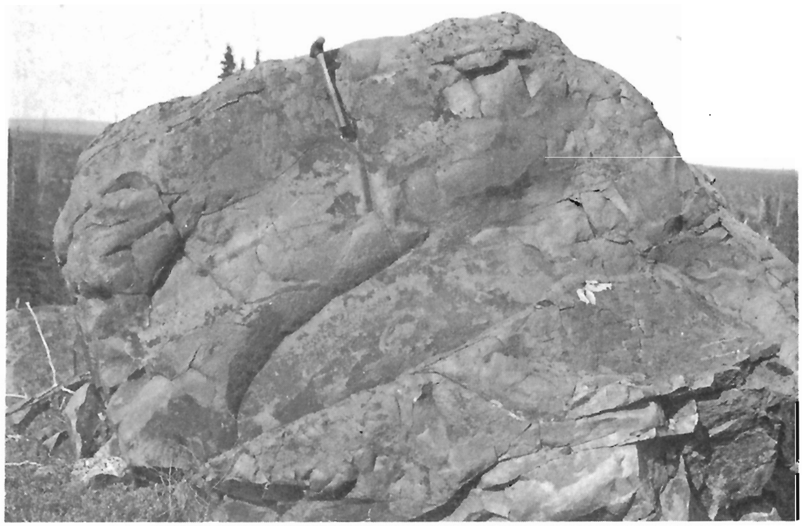


Figure 58. Oxide facies of member VII in outcrop near Ridge Lake. Note the grey colour, thickness of bedding and the overall massive appearance of the rocks.

Minnesotaite appears to be the only silicate in the silicate facies. It occurs as fibrous to felty, and as a rule, poorly defined granules or as fibres or fibrous clusters irregularly disseminated in chert. Siderite, where present, is anhedral and usually medium to coarse grained. It occurs as oval to elongate granules, as small irregularly scattered particles and as cement of siderite, minnesotaite and chert-minnesotaite granules. Magnetite appears most commonly as disseminated euhedral crystals in various parts of the facies.

Oxide Facies

The oxide facies of member VII is the most thickly bedded unit of the Sokoman Formation. Most beds are 1 foot to 3 feet thick, but beds up to 5 feet thick are not unusual. The beds are irregular and typically massive in appearance (Fig. 58). Pitted surfaces formed by weathering of haphazardly scattered patches of carbonate are seen in some of the outcrops in and near the mixed facies areas, and a few widely scattered jasper fragments are present locally near the contact with the underlying jasper-bearing oxide facies of member VI. Small-scale crossbedding outlined by wispy concentration of iron oxides, fine-grained pebbly beds and lenses, and rare thin lenses and stringers composed almost wholly of chert or iron oxides, can also be recognized in a few places on close inspection.

Granules are abundant although most of them cannot be recognized without a microscope. These slabby, irregular to nearly equidimensional particles are composed essentially of chert mixed in various proportions with iron oxides. The iron oxides (mainly magnetite) are disseminated or concentrated at the periphery of granules. The predominant matrix is chert with

Table 7. Chemical Composition of Carbonate Facies

	A Carbonate Facies Member VII	B Carbonate Facies Lake Superior Region
SiO_2	40.94	34.508
Al_2O_3	0.29	1.090
Fe_2O_3	0.78	1.845
FeO	28.75	31.720
MgO	4.41	3.146
CaO	0.26	1.162
Na_2O	0.09	0.040
K_2O	0.00	0.200
MnO	0.79	1.371
TiO_2	0.00	0.072
CO_2	23.18	22.960
P_2O_5	0.01	0.512
H_2O-	0.05	0.876
H_2O+	0.13	
S	0.01	0.072
C	0.01	1.880
Total	99.68	

A. Composite sample of carbonate facies of member VII. J. L. Bouvier, Analyst. Geological Survey of Canada.

B. Average of seven analyses from the Lake Superior region compiled by Gross (1965, p. 85).

small amount of disseminated iron oxides. Elongate particles or irregular felty aggregates of silicates (mainly minnesotaite) are disseminated in granules or in the matrix which surrounds them. Siderite, most common in the mixed facies areas (Fig. 55), occurs as granules and larger oval to irregular patches haphazardly mixed with the chert-iron oxide granules.

The colour of the rocks ranges from various shades of grey to pink. Grey colours predominate except in the areas north and northwest of Stork Lake where the dominant colour is pink. The colour change reflects the increase of hematite northward of Stork Lake.

Mixed Facies

Much of member VII in the areas designated as mixed facies (Fig. 55) consists of complex interfingering of beds, lenses and larger lenticular sedimentary units of contrasting mineralogical composition. Facies changes are commonly erratic. Units composed mainly of chert-siderite, chert-minnesotaite or chert-iron oxides appear in unordered sequences both laterally and vertically, although the oxide facies tend to predominate in the lower parts of the member.

Hybrid rocks consisting of mixtures of oxides, silicates and carbonates occur in various areas of the mixed facies. Under the microscope such rocks are seen as mixtures of granules, pebbles, fragments and lenses of different mineralogical composition. It is not uncommon to see granules and larger particles composed of chert and iron oxides or almost wholly of siderite or minnesotaite in contact with one another.

Depositional Environment

The greater quantity of silicates and siderite in member VII than in the three underlying units of the iron-formation (members IV, V and VI) suggests an increase in the depth of water throughout the Knob Lake basin. The eastern and western flanks of the basin, however, were still within easy reach of waves and currents as evidenced by the thickness and irregularity of bedding of the invariably granular, locally crossbedded and conglomeratic beds in the eastern and western part of the Knob Lake area. The presence of the argillaceous, clastic facies of iron-formation in the south-central and southeastern parts of the Knob Lake area can probably be related to erosion of the Nimish Islands which existed just southeast of the Knob Lake basin.

Environmental interpretation of the facies relationships in the western part of the Knob Lake area is shown in Figure 59. The unusual carbonate facies presents the main problem in interpretation. The mineralogy of the facies (chert-siderite) indicates that Eh and pH of the environment were low so that neither oxides nor silicates could form. The texture of the rocks, on the other hand, suggests deposition in a high energy environment where ordinarily both Eh and pH would be high and where oxides or silicates rather than siderite would be expected to form. To resolve the contradicting evidence, Briggs (1971, pers. comm.) suggested an inflow of low pH, iron-rich riverwater that precipitated the iron as carbonate to form a deltaic deposit which was reworked by waves and currents. Low Eh in the area of carbonate deposition could have been maintained by the oxidation of carbonaceous (organic?)

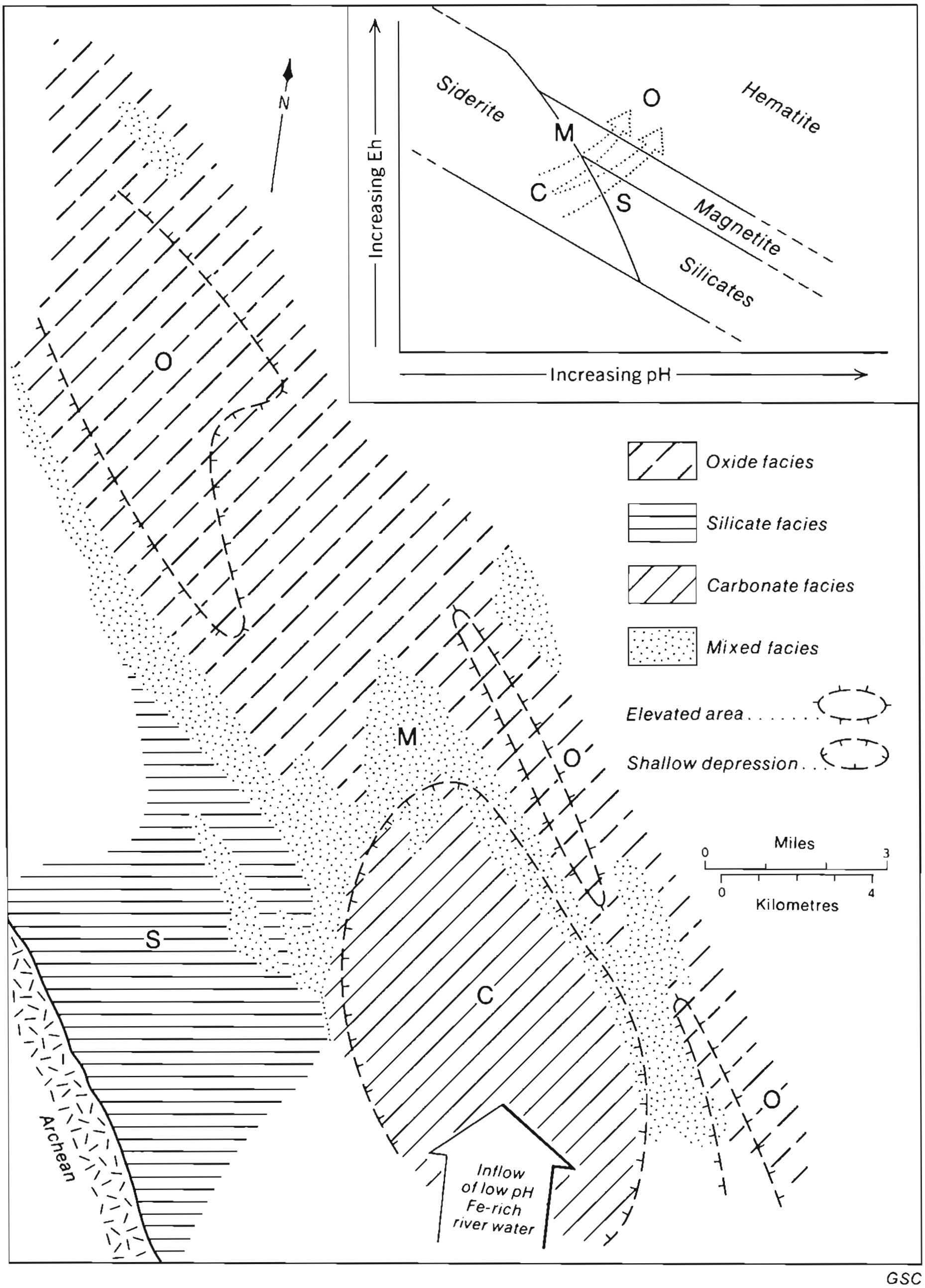


Figure 59. Interpretation of the depositional environment of member VII. Diagram in upper right hand corner shows the relationship of various minerals to Eh and pH. Arrows indicate the mineralogical changes outward from the area of carbonate deposition.

Figure 60. Algal(?) structures as seen in a bedding plane at the top of member VIII west of Stakit Lake.

material which may have been also brought in by the river water. The carbonate delta probably began to form on the south slope of the shallow depression shown in Figure 59. As the accumulating sediment was reworked and redistributed, the delta spread out northwestward along the axis of the small basin. Farther out, silicates and oxides were deposited in response to the increasing pH and Eh of the open sea.

MEMBER VIII

Member VIII is a thin, somewhat irregular unit composed almost wholly of massive-bedded chert, most of it hematitic jasper. The areas south of Iron Arm and southeast of lac Vacher where the cherts of member VIII are closely interbedded with shales and greywackes of member IX, are the only exceptions to the generally simple composition of this part of the Sokoman Formation.

The thickness of member VIII is variable. The member is thinnest near the western margin of the Labrador Trough where it is 4 to 8 feet thick. In other areas the thickness is generally within the range of 10 to 25 feet.

Member VIII is not the most conspicuous part of the Sokoman Formation. However, once one becomes familiar with the pecularities of its colour and general appearance in outcrop, the member becomes one of the most persistently identifiable stratigraphic units. The green and particularly the more common purple-red colours combined with the vitreous-waxy appearance of the cherts and their abnormally low iron content (10 to 15 per cent Fe or less) are the most useful identifying properties of the member. The blotchy distribution of green and purplish colours that Seguin (1961) found so useful in identifying this part of the iron-formation in the eastern part of the ore zone are also diagnostic of the member in most other areas west of Knob Lake. However, near lac de l'Hématite and the areas northwest of it, as well as in most areas east of Knob Lake, the predominant colour is purple-red. In the area surrounding Wishart Lake the rock is almost exclusively green.

West of Stakit Lake and Howells River the top of the member is marked by algal structures (Fig. 60) which are well exposed on the top surfaces of the gently dipping beds. Perrault (1955) was the first to describe the structures and use them as a horizon marker. These structures (Fig. 60) have been identified so far only at the top of member VIII, along the western margin of the Labrador Trough. East of Stakit Lake and Howells River the only structures at the same stratigraphic horizon that may also be of organic origin are a few thin lenses composed of colloform(?) chert aggregates which resemble knotted wood.

Massive beds (a few inches to 1 1/2 feet thick) make up the bulk of member VIII; they are composed mainly of chert with minor amounts of silicates and dust-like hematite. Magnetite is common locally in the upper part of the unit. Other minerals present in small amounts are siderite and ankerite. The silicates that can be identified in thin section are minor minnesotaite, stilpnomelane and a few spherites (15 to 20 microns in size) of greenalite. The silicate which imparts the distinctive green colour to the rocks occurs as dust-like disseminations in chert. X-ray analyses of such chert identify the silicate as greenalite.

Despite the massive appearance of the cherts in outcrop, close examination shows that they are granular to oolitic and locally conglomeratic. Fresh, wetted surfaces of the green cherts when examined under a binocular microscope are seen to consist of closely packed granules which resemble small green peas that have been set in a grey to colourless gelatin. Many of the granules are almost perfect spheres of nearly identical colour, texture and composition. The granules are also similar in size (most are 0.7 to 1.5 millimetres in diameter), although locally a single granule may be several times the size of those surrounding it. Syneresis cracks are common near the centre of some of the larger granules.

Oolitic and conglomeratic cherts are confined mainly to the hematitic purple-red jaspers. The oolites, composed of thin concentric layers of chert and hematite are similar in size, texture and composition to the oolites in the strongly to moderately oolitic zones of member IV. The only obvious difference between oolites of the two members is the much smaller content of iron (not more than 5(?) per cent) in the oolites of member VIII. Most conglomerates are small lenticular layers composed of aphanitic chert pebbles (mainly purple jasper) set in a granular chert matrix. The pebbles are elongate, well-rounded slabs up to 2 inches in size. The texture and composition of these pebbles is similar to the few thin, aphanitic chert layers interbedded with some of the thicker, granular beds.

Depositional Environment

The presence of oolites, pebble conglomerates and of widespread hematitic jaspers indicates a shallow, oxidizing to mildly reducing environment throughout most of the Knob Lake area during deposition of member VIII. The area near the present Wishart Lake where a small subsidiary basin existed during deposition of member VI and VII (see Figs. 52 and 59), remained a small depression during deposition of member VIII. The existence of a somewhat deeper more reducing sedimentary environment in this area during deposition of member VIII is indicated by the scarcity to complete absence of oolites, conglomeratic beds and hematite.

The algal structures indicate clean, shallow, sunlit waters along the western margin of the Labrador Trough. The absence of such structures in other parts of the Knob Lake basin may be due to the increasing depth of water and/or the influx of muddy sediments (shales and greywackes of member IX) which would have inhibited the growth of the algal colonies.

The introduction of clăstic material from the Nimish area was restricted largely to within a few miles of the present Petitsikapau, Dyke and Astray lakes. The influx of clastics increased sharply towards the end of the depositional history of member VIII so that most of the member in the Knob Lake basin became covered by these clastics (shales and greywackes of member IX.

MEMBER IX

The iron-formation comprising member IX consists of interbedded shales, greywackes, cherts and minor siderite-rich rocks. The red, hematite-rich unit mapped in the mines by company geologists as RUIF (Red Upper Iron-Formation) is the altered equivalent of the interbedded shales, greywackes and cherts of member IX. The Red Upper Iron-Formation of Schwellnus (1957) and Gross (1968) correlates with member VIII.

On the basis of the most common iron minerals, the unaltered iron-formation can be assigned to one of two types: silicate-oxide facies or carbonate facies. The silicate-oxide facies is divisible further into a clastic and a non-clastic unit. "Clastic" denotes the part of the silicate-oxide facies which contains feldspar, detrital quartz, fragments of volcanic rocks or any other clasts of foreign derivation. "Non-clastic" indicates the silicate-oxide facies in which no foreign clasts could be definitely identified. Distribution of the facies and thickness of the member are shown in Figure 61.

Clastic Silicate-Oxide Facies

The clastic facies of member IX consists mainly of fine-grained greywackes and shales similar to those in the upper part of the Sokoman Formation in the Nimish area described by Sauve (1953) and Perrault (1955). The correlation of greywackes and shales in the Knob Lake and Nimish areas is easily established by following the outcrops along the ridges which extend from lac Vacher southeast to Dyke, Astray and Petitsikapau lakes.

The greywackes and shales are greenish-, brownish-, and rarely purplish-grey, thin-bedded rocks composed mainly of variable amounts of iron silicates, magnetite, chert and feldspar. The thin bedding, predominant brown weathering and argillaceous rather than cherty appearance are characteristic features of the rocks. The presence of feldspar, detrital quartz and rare fragments of volcanic rocks distinguish the clastic facies from the otherwise similar, stratigraphically equivalent non-clastic facies of member IX in other areas (Fig. 61). Crossbedding is common in some of the thicker

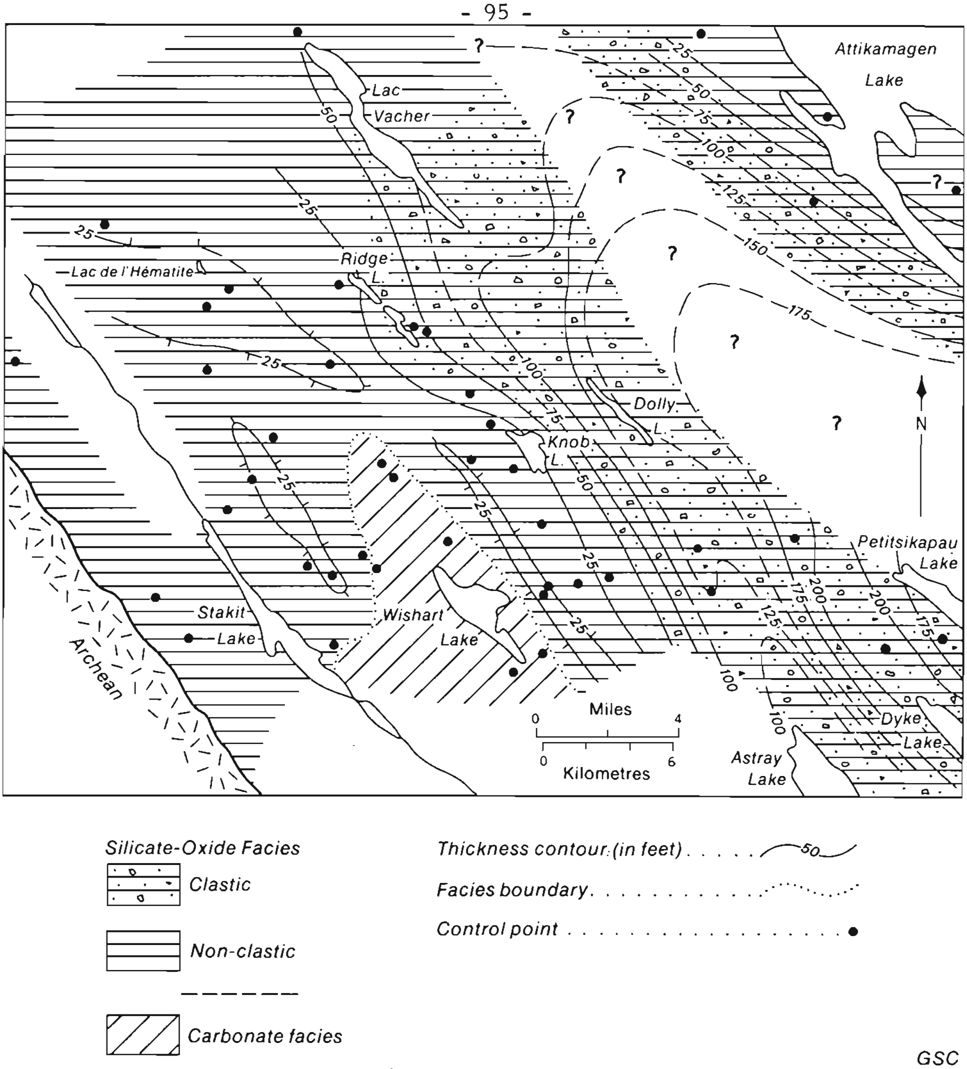


Figure 61. Facies map of member IX.

beds, particularly in the area between Petitsikapau and Astray lakes (Fig. 62) Minor conglomerate beds, many of them with pebbles and lesnse of jasper, are also most numerous in the immediate vicinity of these lakes.

Magnetite, on the average, makes up 20 to 25 per cent of the clastic facies. A small(?) amount of titaniferous magnetite is also present. This magnetite, with grid-like intergrowths of ilmenite, occurs as angular to well-rounded grains, most of which are partly to completely surrounded by overgrowths of ilmenite-free magnetite (Fig. 64). Sauve (1953), who was the first to identify titaniferous magnetite in the Nimish rocks, compared semiquantitative analyses of magnetite concentrates obtained from volcanic rocks and greywackes with analyses of the magnetite extracted from the cherty uncontaminated iron-formation. He found that titanium lines show up strongly in the spectrographic plates of magnetite from volcanic rocks, less strongly in magnetite from greywackes, and not at all in the magnetite from the normal cherty iron-formation. The implication of these results combined with

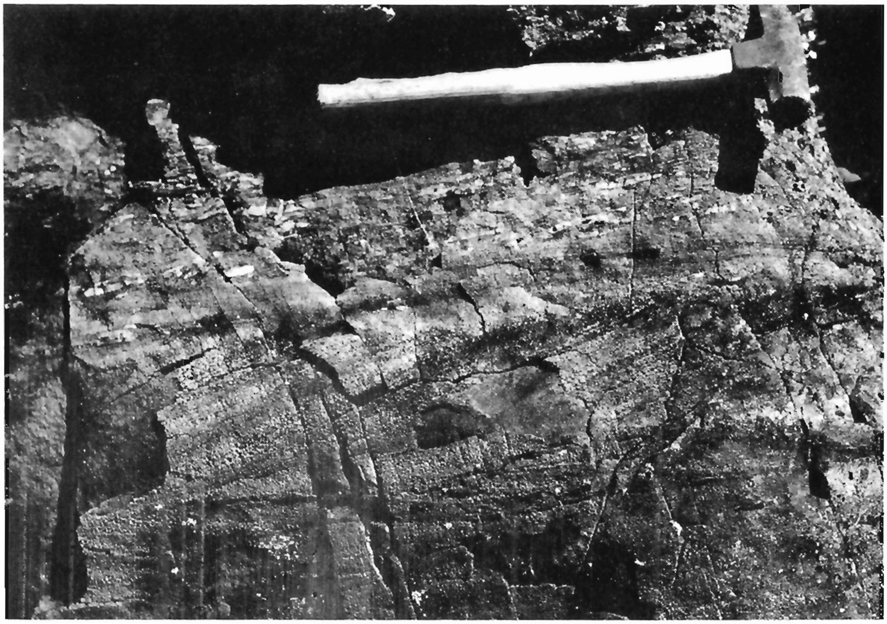


Figure 62. Crossbedding in feldspathic greywackes (clastic facies) of member IX. Dyke-Petitsikapau Lakes area.

Figure 63. Brown weathered magnetic greywackes (non-clastic facies) of member IX northwest of Knob Lake. Note crossbedding in one of the chert beds (top of photograph).

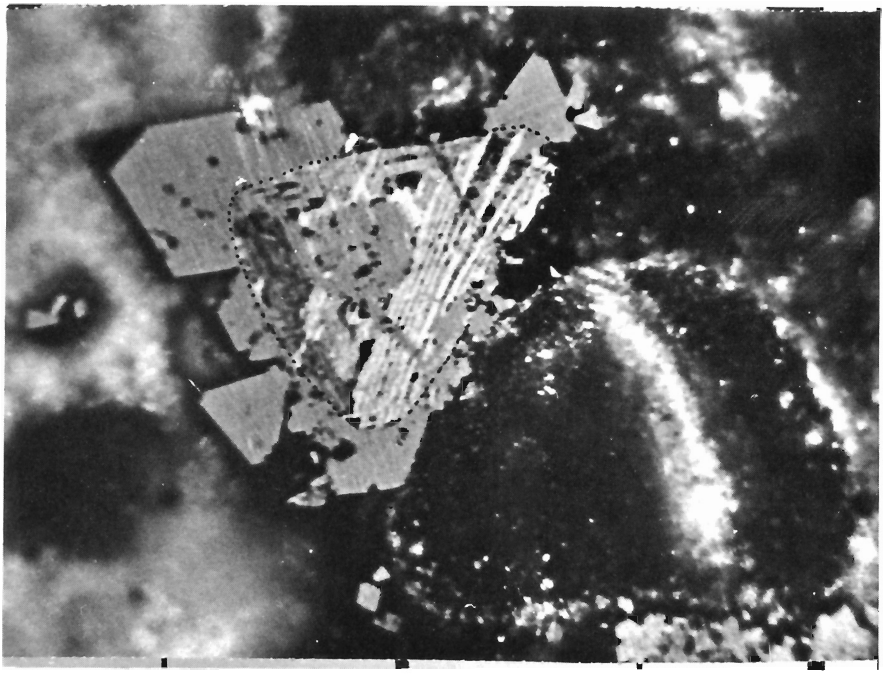


Figure 64. Polished section of greywacke from just north of Astray Lake showing rounded grain of titaniferous magnetite (magnetite-ilmenite intergrowth) surrounded by overgrowth of ilmenite-free magnetite. Note that crystals in the overgrowth have the same crystallographic orientation as the ilmenite-bearing grain. The pellet-like structure in right-hand corner is a jasper granule.

textural evidence (ilmenite-magnetite intergrowths) is that the titaniferous magnetite in the greywackes was derived from the volcanic rocks in the Nimish area. The rounded grains of titaniferous magnetite, such as the ones shown in Figures 32 and 64, are very likely detrital grains derived by erosion of the volcanic rocks on the islands that existed in the Nimish area during Sokoman sedimentation. The overgrowth texture (Fig. 64) suggests that after erosion, transport, and subsequent deposition in iron-rich waters of the Knob Lake basin, the detrital particles of titaniferous magnetite served as seed-grains around which some of the chemically precipitated, ilmenite-free magnetite was deposited.

Minor amounts of the detrital quartz (generally not more than 4 to 6 per cent), which is present in most of the greywackes and shales, occurs as angular to well-rounded grains. In some of the conglomerates and breccias in the Nimish area between the central parts of Dyke Lake and Astray Lake, pebble- to boulder-size fragments of quartzite can also be recognized. These fragments, previously noted by Sauve (1953), are similar to the quartzite of the Wishart Formation which underlies the Sokoman. There is little reason to doubt that the pebbles and boulders of quartzite as well as most of the small clastic grains of quartz in the greywackes were derived, probably by erosion (Sauve, 1953), from the Wishart Formation.

Feldspar is the most common allochthonous detritus in the clastic facies of member IX. The amount of feldspar is highly variable (0 to 50 per cent) but generally increases southeastward towards Dyke Lake and Petitsikapau Lake. The feldspar occurs mainly in the form of angular to rounded, blocky particles 0.1 to 0.4 millimetre in size. With the exception of minor plagioclase, minor myrmekitic (granophyric?) quartz-feldspar intergrowths and a few grains of microcline, the feldspar is an untwinned potassium feldspar (orthoclase? and adularia?). The feldspar is dusted with fine inclusions and has an irregular, mottled appearance in polarized light. An unusual feature of the feldspar in the greywackes when contrasted with the feldspar in the Attikamagen, Wishart and Menihek Formations is its replacement by fine-grained silicates – mainly(?) stilpnomelane. The volcanic rocks in the Nimish area are the most probable source of most of the feldspar in the greywackes. As noted by Sauve (1953), the type of feldspar and the ratio of quartz to feldspar in the greywackes and in the Wishart Formation are so different that the alternate possibility of deriving most of the feldspar from the Wishart Formation must be disregarded.

Fragments of Nimish volcanic rocks are common in the greywackes near Dyke and Astray lakes (Sauve, 1953; Perrault, 1955), but are rare in the stratigraphically equivalent rocks in the Knob Lake area. The volcanic clasts are granule- to pebble-size fragments, still identifiable as volcanic by their vesicular and micro-ophitic textures.

Chert is less common in the clastic facies of member IX than in any other part of the member. It occurs as greenish grey to red interbeds up to 2 or 3 inches thick, and as fragments or lenses of such layers. In thin section chert is also seen as fine-grained granules or thin layers, usually with at least some magnetite or iron silicates, and less commonly as interstitial matrix of the clastic particles.

The silicates, which make up the fibrous-felty "argillaceous" material of the shales and greywackes in the Knob Lake area, are very fine grained, and, with the exception of a few coarser, strongly pleochroic grains of stilpnomelane, are difficult to identify in thin section. X-ray analyses show that the silicates are stilpnomelane with or without admixed minnesotaite, chlorite and mica. In the Dyke-Astray lakes area, the most common silicates are chlorite (chamosite?) and minnesotaite (Sauve, 1953; Perrault, 1955).

Non-Clastic Silicate-Oxide Facies

The non-clastic silicate-oxide facies consists of interbedded magnetic shales, magnetic greywackes and cherts. Siderite-rich rocks which become the dominant lithology of member IX in the area near Wishart Lake (the carbonate facies shown in Fig. 61) first appear in the lower part of the member.

The names "magnetic shales" and "magnetic greywackes", used previously by Dufresne (1952), Perrault (1955) and a number of company geologists in allusion to the argillaceous appearance, the clastic textures and the commonly high magnetite content of the rocks, are useful terms for this distinct type of iron-formation. A previously unrecognized mineralogical characteristic of these rocks in the Knob Lake area is their high content of stilpnomelane. These shales and greywackes are the only stilpnomelane-rich rocks in the Sokoman Formation north to west of Ruth Lake (western part of the ore zone and Stakit Lake area).

The shales are thinly bedded (banded) to laminated, locally fissile rocks of a distinct dark greyish green to greenish brown colour. The typical weathering colour is brown. Mineralogically the shales are a fine-grained mixture of stilpnomelane, magnetite and chert. Minnesotaite is locally abundant, but chlorite is rare. Hematite, with the exception of strongly altered rocks, is absent or present in not more than accessory amounts. Magnetite, which makes up 30 to 40 per cent of the shales, is disseminated or concentrated into thin layers.

The greywackes are mineralogically similar to the shales, but are thicker bedded, coarser grained and more coherent than the shales. The sandy to conglomeratic texture (Fig. 63) stands out clearly in strongly weathered outcrops. Crossbedding is present locally, but is less common and less easily recognized than in the feldspathic greywackes near Petitsikapau, Dyke and Astray lakes.

The cherts interbedded with the shales and greywackes are predominantly dense, granule-bearing, greenish grey rocks composed of chert with subordinate amounts of magnetite and silicates. An unusual feature of some of the thicker, more irregular beds is the well-developed crossbedding (Fig. 63). The quantity of chert interbeds is variable. East of Stakit Lake the combined thickness of the cherts represents anywhere from 10 to 60 per cent of the total thickness of the member. West of Stakit Lake and Howells River the cherts make up over 80 to 90 per cent of the member. In this area the greenish grey chert beds are interspersed with thin, massive layers of magnetite and thin, shaly stilpnomelane-rich layers.

Carbonate Facies

A few thin, discontinuous siderite-rich beds are not uncommon in the lower part of member IX throughout the southwest part of the Knob Lake area. In most places, however, they represent only a very small part of the member. The area indicated as carbonate facies in Figure 61 is an exception. Here, siderite is the dominant iron-bearing mineral of the entire stratigraphic unit.

The carbonate facies is composed almost entirely of alternating chert, chert-siderite, and siderite beds. Interbeds of the previously described magnetic shales and greywackes are common near the boundary of the carbonate facies area.

The chert beds, which comprise approximately a third to a half of the carbonate facies unit, are greenish grey to dark grey and 1 to 4 inches thick. They are composed of numerous spheroidal chert granules set in a somewhat coarser grained chert matrix. The finely disseminated accessory minerals are stilpnomelane, greenalite, and minnesotaite. The dust-like inclusions that outline many of the granules in the dark grey cherts may be carbon. Laminated layers less than half an inch thick and composed of uniform micrograined chert are rare interbeds of the granule-bearing cherts and siderite-rich beds.

The siderite-rich beds, which are commonly thicker than the associated granule-bearing cherts, stand out as brown layers in the otherwise light-coloured outcrops of the carbonate facies. They are composed mainly (50 to 90 per cent) of grey, fine- to coarse-grained siderite. In hand specimen the siderite appears to be a massive intergrowth of anhedral grains. Thin sections show, however, that most of the siderite is in the form of spheroidal to lobate granules, cemented by siderite or a mixture of chert and iron silicates. Stilpnomelane, the most common silicate, makes up 10 to 20 per cent of some of the beds. The stilpnomelane is present both in the sideritic granules and in the matrix which surrounds them, but on the whole appears to be characteristic of the matrix where it occurs as irregular felty patches intergrown with various amounts of chert or siderite.

Depositional Environment

One feature that stands out very clearly from the lithological record of member IX is the influx of clastic material from the Nimish area now occupied largely by Petitsikapau, Dyke and Astray lakes. The presence of feldspar, detrital quartz, volcanic fragments, titaniferous magnetite and the increasing abundance of feldspathic greywackes toward the southeast, all point to the Nimish area as the source. Member IX represents the greatest accumulation of coarse allochthonous detritus in the depositional history of the Sokoman Formation in the Knob Lake area. This agrees well with Sauve's studies (Sauve, 1953) which showed that the Nimish area was most actively eroded at approximately the same time that the upper part of the Sokoman Formation was being deposited. The volcanic centres on the islands were also the most active during that time.

The clastic facies of member IX, which contains the coarsest and the only clearly recognizable clastic material, extended northwestward from the Nimish area along the axial part of the Knob Lake basin. The stilpnomelane-bearing nonclastic facies probably represents that part of the iron-formation which was mixed with fine-grained, mainly clay-size particles transported farther from the Nimish source. The clastic and nonclastic facies of member IX are so closely related in time and space and in composition that it is difficult to escape the conclusion that these units are also genetically related. Although no foreign material can now be identified as such in the nonclastic facies, it seems likely that at least some of the stilpnomelane in these rocks, as in member III, was formed by alteration of the fine detritus derived from the Nimish area. Direct evidence of such alteration is seen in the replacement of feldspar by stilpnomelane in the clastic facies.

The small amount of stilpnomelane and the absence of any recognizable clastic material in the carbonate facies indicate that the area where the carbonate sediment was deposited was largely bypassed by the clastics which spread outward from the Nimish area. The reason for this is not clear. A small ridge may have existed east of the carbonate area and protected it from the influx of detritus. It is also possible that an inflow of river-water, similar to the inflow postulated in the interpretation of member VII (Fig. 59), diverted the muddy currents emanating from the Nimish area past the area of carbonate deposition (Fig. 65).

MEMBER X

Member X consists mainly of bedded cherts with subordinate shales and siderite-rich rocks. It is characterized by the widespread occurrence of black chert, the scarcity of magnetite and the virtual absence of hematite and jasper. The thickness of the member is variable, 20 to 50 feet being the most common range.

The cherts are particularly abundant in the area between Knob Lake and Stork Lake where they account for more than 90 per cent of the member. They are grey to black and less commonly green rocks composed almost entirely of fine-grained chert. Microscopic examination shows that most are composed of abundant microcrystalline chert granules set in an equally fine-grained to somewhat coarser grained chert matrix. The oval to nearly spherical granules are defined by fine-grained greenalite, minnesotaite or by dusty amorphous particles which are probably carbon. The black cherts are similar in many respects to the cherts in the lower part of the Sokoman Formation (Basal Chert and lower part of member I); the green cherts resemble those in members VIII and IX.

Most siderite-rich rocks in member X are within a few miles of the western margin of the Labrador Trough. West of Stakit Lake and Howells River they are interbedded in about equal proportion with black cherts. The sideritic interbeds are a fraction of an inch to several inches thick and composed essentially of siderite. Chert and less commonly minnesotaite and/or stilpnomelane are accessory constituents. The siderite is microcrystalline to medium grained and usually without any pellet-like structures. Faint lamination is evident in most beds, especially on strongly weathered surfaces. Slaty beds are reported by Perrault (1955). Soft-sediment deformation structures, stylolites and chert-siderite breccias are rarely seen in outcrop.

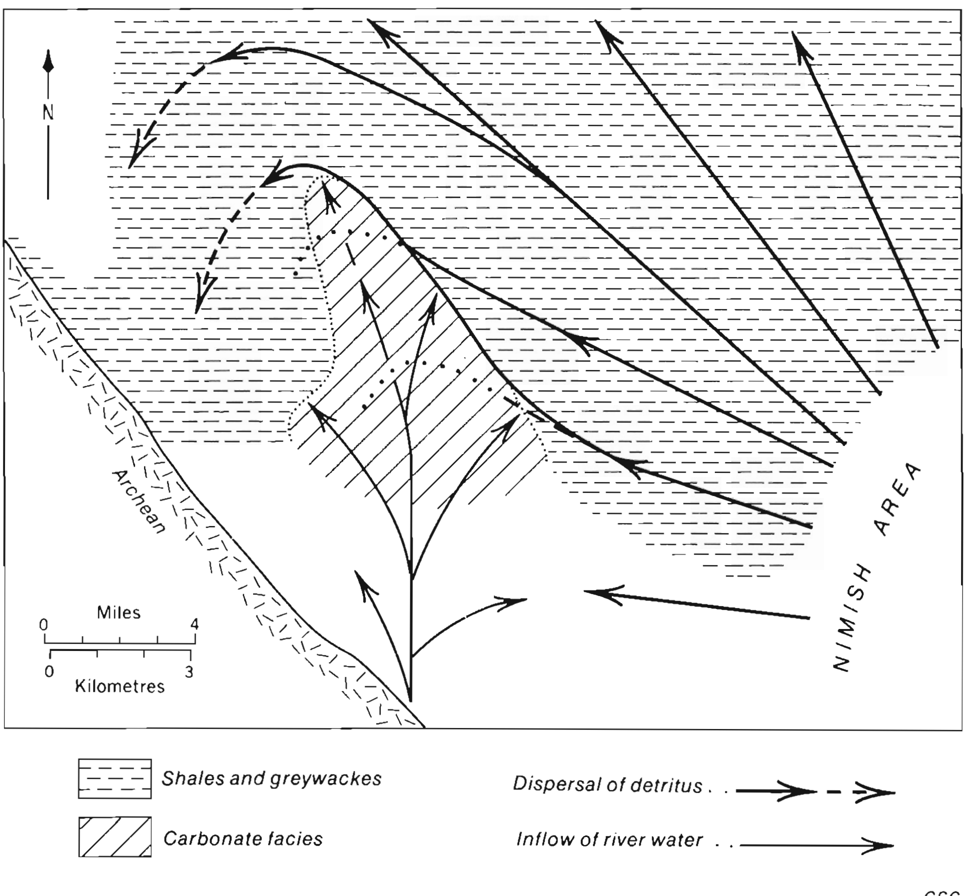


Figure 65. Dispersal of detritus from Nimish area.

Siderite persists as prominent interbeds in member X for 2 to 3 miles east of Stakit Lake and Howells River but in diminishing amounts. The quantity of sideritic beds within the member diminishes from approximately 50 per cent west of Stakit Lake to somewhat less than 10 per cent in the area near Stork Lake. Farther east, particularly in the ore zone, siderite occurs as thin discontinuous stringers and disseminated particles which collectively represent not more than about 5 per cent of the member.

Shales, and less commonly very fine grained greywackes, are abundant on the ridges which extend from lac Vacher to Dyke Lake and Petitsikapau Lake. These rocks resemble the fine-grained clastic facies of member IX. The shales and greywackes of member X can be distinguished from the underlying rocks by their lack of hematite and jasper and by the scarcity of magnetite. Magnetite is rare or absent except in some of the shales near lac Denault and some of the greywackes near Dyke and Petitsikapau lakes which are magnetic. The shales and greywackes of member X are also generally darker and presumably richer in carbon than the clastic rocks of member IX.

Depositional Environment

The increasing depth of the transgressing sea which during Menihek time allowed several hundred feet of carbonaceous sediments of the Menihek Formation to accumulate, also affected the last stages of Sokoman sedimentation. It created a widespread reducing environment in the Knob Lake basin. This is evident from the widespread distribution of carbonaceous chert; the presence of siderite in the once shallow part of the basin (Stakit Lake area); the scarcity of magnetite and the virtual absence of hematite and jasper.

The shallowest parts of the Knob Lake basin during deposition of member X were in the Nimish area near Dyke and Astray lakes. Parts of this area apparently still remained as islands in the advancing sea and supplied detrital sediments (shales, greywackes) to the surrounding areas. As the encroaching sea gradually engulfed the islands the influx of clastics diminished and ceased by the end of the Sokoman time. Volcanic activity in the Nimish area also had subsided by the time the first sediments of the Menihek Formation were deposited.

MINERALOGY OF THE SOKOMAN FORMATION

The following discussion is restricted to the primary (pre-alteration) minerals of the iron-formation. Minerals such as goethite (limonite included), manganese oxides, and secondary hematite (martite, soft red hematite), which formed by supergene alteration of the iron-formation, are not discussed. Small, rare, epigenetic deposits of specularite and massive hematite, similar to those described by Gross (1968, p. 69-73) are also excluded.

CHERT

Chert, the microcrystalline variety of quartz, is the most common and widespread constituent of the Sokoman Formation. In most units (facies, members) of the formation it accounts for more than 40 per cent of all minerals. Units with more than 70 per cent of chert are the Basal Chert member, members VIII and X. The units with abnormally low chert content are the silicate facies in which much of the silica is tied up in silicates, and the clastic facies in which the chert has been diluted by addition of foreign clastic material.

In megascopic appearance the cherts of the Sokoman Formation range from microcrystalline, vitreous rocks to somewhat coarser grained rocks which resemble fine-grained quartzites. Granules are common in all thick- and irregular-bedded cherts, although most of them cannot be recognized without magnification. Thinly layered cherts, on the other hand, are predominantly nongranular.

In thin sections viewed in polarized light, the cherts are seen as a mosaic of interlocked quartz grains of uniform to highly variable grain size. In the finest grained cherts most grains appear to be irregular, patchy, with a wavy to mottled extinction, and in places complexly intergrown. The complexity of intergrowth appears to vary with grain size. Generally the larger the grains, the fewer and less complex the intergrowths.

The texture of chert revealed by electron microscope study is very different from what it appears to be in thin section. Replicas of fresh fracture surfaces of cherts show that the individual grains which make up the chert mosaic are sharply defined polygonal blocks bounded by simple but, as a rule, distinctly curved planes (Fig. 66). This texture appears to be typical of the cherts in the Sokoman Formation. Replicas from nine specimens of various types of chert examined by the writer showed very similar polygonal textures. The main difference in the microtextures was the size of the polygonal blocks.

An unusual texture was seen in one specimen of chert which contained platy silicates (stilpnomelane or chlorite). Fracture surfaces of this chert showed numerous clusters of well-formed quartz crystals (Fig. 67). The perfection of the crystals suggests crystallization in open space, such as a microfracture or a vug. It is also possible that the quartz crystals developed in a plane adjacent to the silicate clusters which offered little resistance to the growth of the crystals.

The polygonal block texture of the cherts in the Sokoman Formation is similar to the texture of the "novaculite type" cherts studied by Folk and Weaver (1952). In discussing the texture, Folk and Weaver explained that the surfaces of the polygonal blocks are curved because:

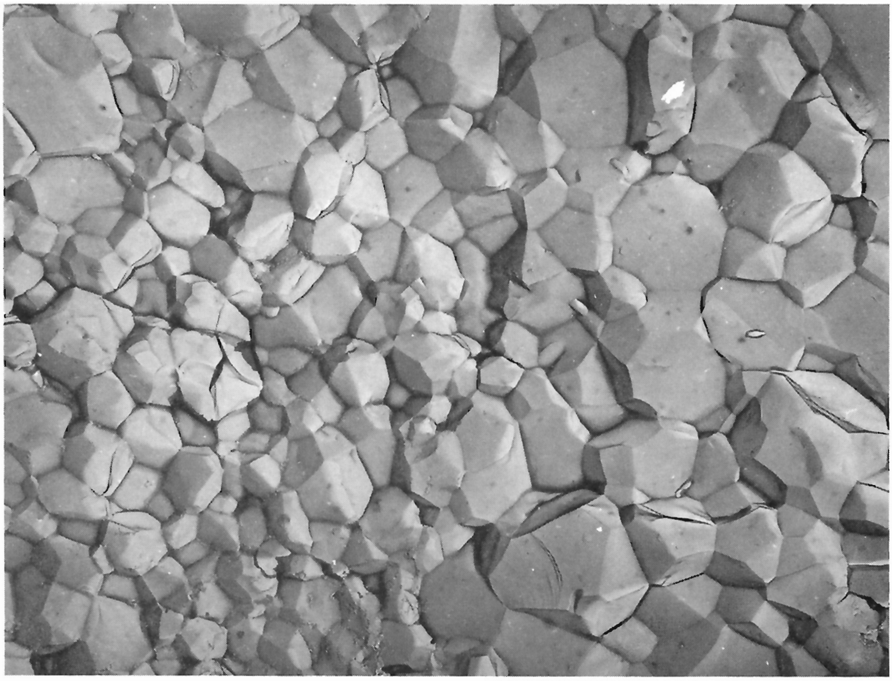


Figure 66. Typical "polygonal block" texture of chert. Photographed with electron microscope. R = replica imperfections. F = fracture surface.

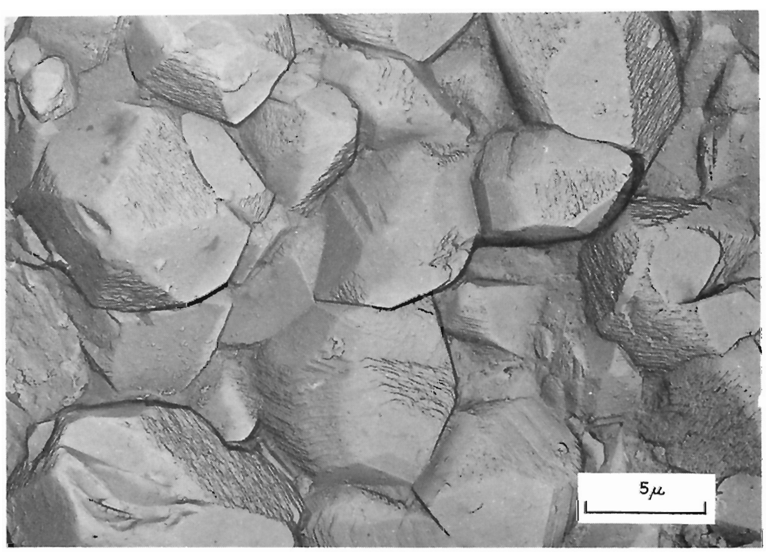


Figure 67. Quartz microcrystals in chert. Photographed with electron microscope.

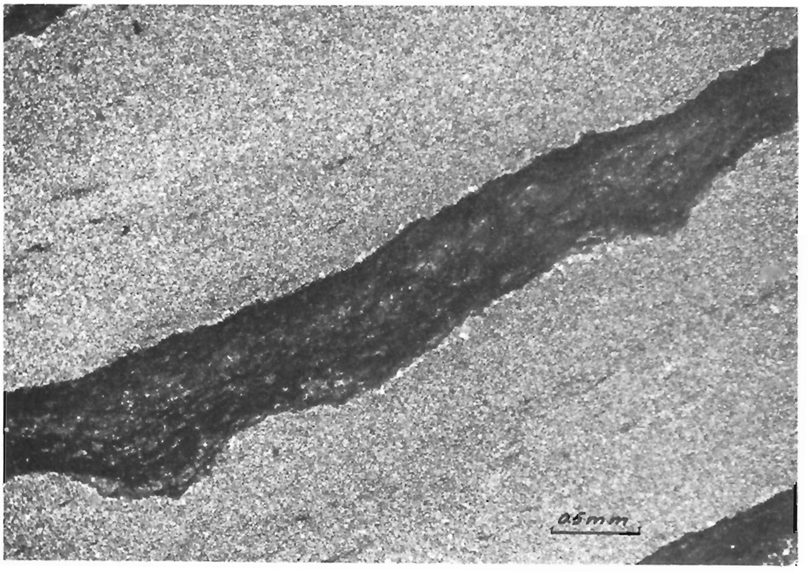


Figure 68. Chert laminae in carbonate facies of member I, Stakit Lake area. The uniformly micrograined texture is typical of thinly layered, nongranular cherts.

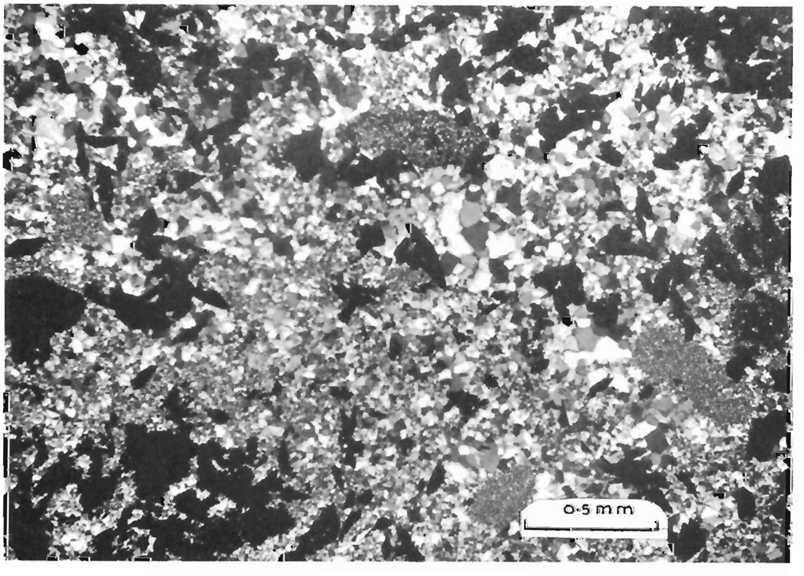


Figure 69. Patchy texture of chert common in granular-conglomeratic beds of member V, Knob Lake Ridge.

> . . . they do not represent crystal faces but merely the interface between two conformable masses – exactly as the interfaces between separate air cells in a soap froth . . .

The anomalous extinction of chert observed in thin section is attributed in part, as suggested by Folk and Weaver, to the interference created by the superimposition of several randomly oriented grains. In thin sections of normal thickness (30 microns) of cherts composed of 5 micron grains, for example, there are bound to be at least six superimposed grains in any line passing at right angle through the plane of the section. The observed optics, therefore, are not that of one but of several grains. The interlocking of grains seen in thin section is probably due to the same effect. The cherts which appear to be composed of interlocking grains in thin section show no evidence of intergrowth when examined by electron microscope. Except where inclusions are present, the contacts between adjoining grains are perfectly smooth (Fig. 66). The distortion created by superimposition of grains can be partly verified by studying thin sections of variable thickness. By shifting the field of view from thicker to thinner portion of a section, the individual grains of the chert mosaic become noticeably better defined. The effect of grain superimposition also explains why the intergrowths in chert appear to decrease with increasing grain size. As the individual grains become larger, there are fewer grains to interfere with the true rendition of the texture and optical property of the grains.

One of the characteristics of the chert in the Sokoman Formation is its great range in grain size. The common range is from a few microns to 150 microns, but grains of up to 300 microns are not unusual. The finest grained and the most persistently uniform cherts are those in the thinly layered, nongranular parts of the iron-formation. The chert in the carbonate facies of member I is the type example (Fig. 68). The predominant range of grain size of the chert is a fraction of a micron to 9 microns with the average being close to 5 microns. Coarser grained patches can be seen locally along bedding planes or in places where the rocks become even slightly granular. Such grain variations, however, are quantitatively insignificant. On the whole, the chert is very uniform in grain size and texture, both in hand specimen and under the microscope. At the other extreme, the cherts of coarsest grain and of most variable texture and grain size are those in the oolitic, granular and conglomeratic units of the iron-formation, such as members IV, V, VI and VII. In the cherts of these units it is not uncommon to see the entire range of grain size, from a few microns to several hundred microns, in the same thin section. The overall texture is a patchy network characterized by very irregular distribution of chert of different grain size (Fig. 69). The variation in grain size, however, is not as unsystematic as it might appear. In most granular and oolitic cherts, as noted by many previous investigators, the chert in the granules and oolites is finer grained than the chert in the matrix (Fig. 70). There are numerous exceptions. In some cherts, the entire granules or various parts of these structures may be of approximately the same grain size or even of considerably coarser grain size than the matrix. In general, however, the coarser grained chert occurs in the matrix rather than in the granules or oolites.

The coarser grain size of granular and oolitic cherts results in part, as suggested by Gross (1961), from the greater quantity of interstitial water

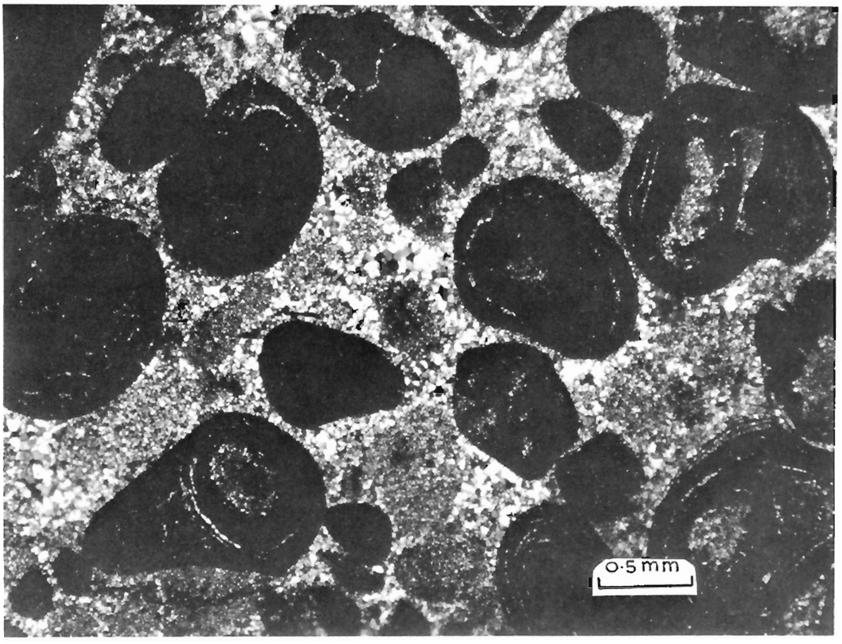


Figure 70. Chert-hematite granules and oolites in distinctly coarser grained chert matrix. Member IV, Stakit Lake area.

which promoted diffusion and grain growth during recrystallization of the sediments. Some coarser grained chert was undoubtedly also deposited as open-space filling-cement in the interstices between the granules and oolites in these originally sandlike, porous rocks, similar to the way sparry calcite cement is deposited in intraclastic limestones.

The grain size of chert in some Precambrian iron-formations correlates well with the intensity of metamorphism (James, 1955). In the sediments of the Knob Lake area, however, the textures of chert are not related to the intensity of metamorphism or the intensity of structural deformation, but to primary sedimentary textures of the rocks. As far as the grain size of chert is concerned, the situation is analogous to that of unmetamorphosed carbonate sediments. Generally, the limestones with the greatest amount of sparry, coarser grained calcite are the intraclastic (granular, oolitic, fragmental) limestones which formed in high energy environments of deposition (Beales, 1965). Beales (1965) shows that there are many exceptions to this general rule, but concedes that "this is a useful generalization for a majority of granular limestone types" (Beales, 1965, p. 55). This generalization applies equally well to the cherts in the iron-formation of the Knob Lake area.

MAGNETITE

Magnetite is the most common and widespread iron oxide of the unaltered iron-formation. The quantity of magnetite is difficult to estimate in the field because of erratic and locally intense martitization. Drilling of unaltered iron-formation and examination of polished sections, however, shows clearly that magnetite is the dominant iron oxide. Accurate figures are not available, but a safe estimate is that magnetite represents at least 65 per cent of the primary iron oxides. The remainder is hematite.

Most of the magnetite occurs in the middle part of the iron-formation, particularly in members IV, V and the oxide facies of members VI and VII, all of which are noted for the abundance of chert and iron oxides and the scarcity of silicates and siderite. With the exception of member VII and the transition zone of member IV, the carbonate mineral characteristic of these units is ankerite not siderite. Magnetite also occurs in various silicate facies, especially in those which contain little or no siderite (member III in the western part of the area, silicate facies of member VI, shales and greywackes of member IX). The units of iron-formation which contain only trace amounts or no magnetite are the carbonate facies of members I, II, VII, IX and X, the Basal Chert member and the carbonaceous and pyritic shales (silicate-sulphide facies) of member I.

The predominant mineral assemblages of magnetite-bearing rocks are chert-magnetite, chert-magnetite-hematite and chert-magnetite-silicates. Ankerite is abundant in some magnetite-rich rocks. Siderite, on the other hand, is one of the less common associates of magnetite. In general, those units of the Sokoman Formation in which siderite is a major component contain little or no magnetite unless silicates (particularly minnesotaite) are also present in abundance. There are three notable exceptions. The thinly layered chert-siderite-magnetite interbeds in the oxide facies of member I is one, and the texturally and mineralogically similar rocks in the transition zone of member IV is another. Lenses and beds with abundant siderite and magnetite also occur locally in the mixed facies of member VII. All three of the above exceptions contain abundant siderite and magnetite in the same beds, but virtually no silicates. The first two are also unusual in that they contain, in addition to siderite and magnetite, small but persistent amounts of hematite.

The change in distribution of magnetite in most units of the iron-formation is accompanied by the change in the sedimentary texture of the rocks. As a rule, magnetite shows a decided preference for those units of the iron-formation which are characterized by irregular bedding and the abundance of granules. Member III is an example. As the amount of magnetite in the unit increases and the silicates and siderite become less numerous, the iron-formation changes from laminated, evenly banded and nongranular to granular as well as more thickly and irregularly bedded. It should be noted, however, that although magnetite-rich parts of the iron-formation are predominantly granular and irregularly bedded, much of the magnetite in some of them (minnesotaite-magnetite facies of member III, conglomeratic-banded facies of member V, shales and greywackes of member IX) is in thin, massive layers rather than in granule form.

Under the microscope the great majority of magnetite particles, regardless of the type of iron-formation examined, are seen as euhedral to subhedral octahedra. The crystals are disseminated or concentrated into dense layers, stringers or irregular to rounded aggregates. The stringers and layers of magnetite are generally parallel to the bedding, but crosscutting relationships are also seen. This is particularly true of single crystals and small crystal aggregates, many of which transect bedding planes or the boundaries of oolites and granules.

Figure 71. Polished section showing magnetite replacing carbonate.

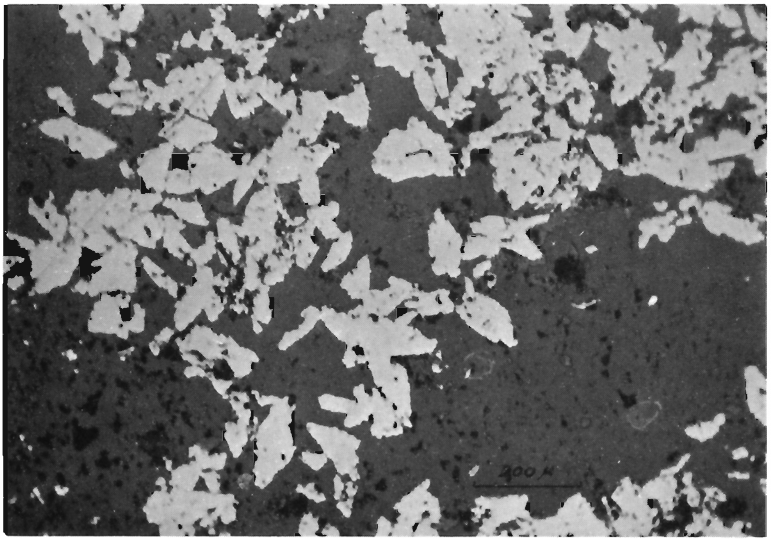


Figure 72. Polished section showing rhomboid to spear-shaped crystals of magnetite. Specimen from ankeritic part of member V east of Wishart Lake.

The rhomboid to spear-shaped crystals and aggregates of magnetite (Figs. 71, 72) are common in some of the oxide facies in which medium- to coarse-grained ankerite or siderite[1] is present. They are particularly numerous in the ankeritic parts of members V and VI. This type of magnetite, as suggested by its shape (Fig. 72), the replacement textures (Fig. 71), and close association with carbonate-bearing rocks, was undoubtedly formed by replacement of the well-crystallized carbonates. The replacement of fine-grained carbonates by magnetite is much less common. Such magnetite is anhedral and has a patchy texture. Three points should be emphasized: (1) the process by which the secondary magnetite forms does not appear to be a simple decomposition or alteration (oxidation) of carbonate to magnetite but a volume-for-volume replacement; (2) the replacement is not restricted to a particular carbonate mineral, although it appears to be most common in ankerite; and (3) the amount of magnetite which has formed at the expense of carbonate represents, very probably, not more than a few per cent of the total magnetite in the iron-formation.

The least common type of magnetite occurs as crosscutting veinlets. The veinlets, a fraction of an inch wide and usually not more than an inch long, are most common in the oxide facies where massive, nongranular layers of chert or iron-oxides are present. Most of these veinlets cannot be recognized without a microscope. They are seen best in polished specimens of the thin-bedded part of member V. Most of the veinlets, as suggested by Dufresne (1952), are probably shrinkage fractures that have been filled with magnetite. The mineralogical peculiarity of the veinlets is that most of them contain no mineral other than magnetite (chert and less commonly ankerite is present in some) despite the fact that the veinlets pass through layers which are composed of chert, hematite, magnetite or of any combination of these minerals. There also is no evidence that any iron or silica has been depleted or added near the contacts of the veinlets. The veinlets certainly do not appear to have originated by local diffusion of material from the beds adjoining the veinlets. They were probably formed by precipitation of iron from the water that was being expelled during compaction and consolidation of the iron-formation.

The size of magnetite crystals ranges from a few microns to 3 or 4 millimetres, 30 to 200 microns being the most common range. Crystals larger than about 600 microns are extremely rare. The finest grained magnetite is in the thinly layered nongranular facies and the coarsest grained in the granular and irregularly bedded facies of the iron-formation. In the laminated, nongranular carbonate-silicate facies of member III, for example, magnetite is invariably less than 20 microns in size. In the irregularly bedded and granular silicate facies of members III and VI and the shales and greywackes of member IX most of it is under 70 microns, whereas in the strongly granular and irregularly bedded units, such as members IV, V, VI and VII, it is generally coarser than 50 microns.

---

[1] Carbonates in polished sections were identified by X-ray analysis of small amounts of carbonate powder extracted with a needle under the microscope.

Origin

A minor amount of magnetite in the Sokoman Formation has formed by replacement of ankerite and siderite and is now present in the form of pseudomorphs after these minerals. The remainder of magnetite is well crystallized, but its textures offer no clue as to what the mineral may have been originally.

The majority of present-day geologists believe that most magnetite in Precambrian iron-formations is primary-diagenetic, that is, formed by recrystallization of an original magnetite precipitate or by crystallization of a pre-existing metastable ferric or ferro-ferric oxide in a diagenetic environment. The strong stratigraphic control of magnetite distribution in the Sokoman Formation, the changes in abundance of magnetite with the change in primary sedimentary textures of the iron-formation, and the appearance of magnetite in mineral assemblages and in sequences predicted by chemical sedimentary models, argue strongly in favour of primary origin. The texture of magnetite, however, is clearly not primary. It is the result of recrystallization of the original precipitate during diagenesis and metamorphism.

The possibility, suggested by some geologists (most recently by LaBerge, 1964), that most magnetite is a metamorphic product formed by alteration (oxidation) of siderite or greenalite seems unlikely. Some magnetite may be metamorphic, but, as pointed out by James, "it is questionable whether this mineral is ever produced in abundance by metamorphism of iron-formation" (James, 1955, p. 1476). Gruner (1946) and White (1954) also noted that the quantity of magnetite does not increase with increasing metamorphism of the Biwabic Formation by the Duluth Gabbro. The studies of Gunderson and Schwartz (1962) and the most recent one by French (1968) confirm their observations. French, who paid close attention to the behaviour of carbonates in the progressive metamorphism of the Biwabic Formation, finds no evidence of "progressive oxidation of either siderite or ankerite to produce magnetite" (French, 1968, p. 61). The Sokoman Formation in the Labrador Trough passes through different zones of metamorphism, but so far, to the writer's knowledge, no one has demonstrated that the amount of magnetite in the iron-formation increases with increasing intensity of metamorphism. The alteration of greenalite seems the least likely source of magnetite in the Sokoman Formation. Greenalite is present locally, but nowhere is there any evidence which would suggest that this mineral decomposes to form magnetite.

HEMATITE

Hematite is texturally very different from magnetite. It also has a more restricted distribution than magnetite and is much less commonly associated with silicates and siderite.

The oxide facies of members I, IV, V, VI and VIII contain nearly all of the hematite in the Sokoman Formation. With the exception of member I, all of these units are noted for their irregularity of bedding, the abundance of granules, the presence of numerous intraformational conglomerates, and locally of very strongly oolitic rocks. The jaspilites of member I are nongranular, non-oolitic and thinly banded to laminated. They are, however, locally interbedded with conglomerates, which like hematite, do not appear in any other facies of the member. Texturally uniform (massive) bands and laminae of hematite are also present in member V where they occur as single, virtually monomineralic layers or as alternating sharply defined layers, each with a different proportion of hematite, magnetite and chert. Although these so-called "metallic bands" are predominantly nongranular and non-oolitic, they are interbedded with beds which are strongly granular, locally crossbedded and commonly conglomeratic (see Figs. 45 to 47).

Mineralogically, the great majority of hematite-rich rocks are noted for the scarcity or absence of silicates and siderite. Ankerite is the only locally abundant carbonate mineral. Conversely, silicate-rich and siderite-rich rocks with more than accessory amounts of hematite are unusual. The typical assemblage is chert-hematite-magnetite and less commonly chert-hematite.

Unlike magnetite, most hematite is very fine grained and not well crystallized. It occurs mainly as subhedral to anhedral elongate, rounded or irregularly shaped grains, and less commonly as euhedral microspecularite plates. The hematite is very uniformly to irregularly disseminated or concentrated in diffuse to sharply defined layers which outline bedding planes, granules and oolites. Well-formed oolites occur only in hematite-bearing rocks. Such large, multishelled oolites as those in the upper part of member IV (Figs. 40, 43) are not known in any other mineralogical facies of the Sokoman Formation.

The size of single hematite particles ranges from submicroscopic to a maximum of about 250 microns, with 90 per cent or more of the total being under 25 microns. Much of the coarsest hematite is found in the beds which contain medium- to coarse-grained ankerite. In such beds the hematite is in the form of very irregular patches or subhedral to euhedral, spear-shaped crystals. Some of these crystals and irregular aggregates, as suggested by Dufresne (1952), were undoubtedly formed by replacement of carbonate; other crystals with a typical crystal form of hematite (flat rhombohedra and plates), may have formed by recrystallization of an originally fine-grained hematite at the time of recrystallization of the carbonate rock.

Origin

Most of the hematite in the Sokoman Formation is almost certainly of primary origin. It occurs mainly in those units of the iron-formation whose sedimentary textures clearly indicate deposition in shallow, frequently turbulent and well-oxygenated water. This is precisely the environment where one would expect the precipitation of ferric oxides to take place. The original precipitate may have been some hydrous oxide – $Fe_2O_3 \cdot nH_2O$, according to Garrels (1959) – which through dehydration and recrystallization was transformed into the hematite seen in the iron-formation at the present time.

Some hematite has formed by replacement of ankerite. However, there is no indication that this process was quantitatively important. The amount of hematite which may be of replacement origin probably does not exceed more than a few per cent of the total in the iron-formation.

GREENALITE

Greenalite, a hydrous iron silicate, was first described and named by Leith (1903) and more accurately defined by Jolliffe (1935) and Gruner (1936). Gruner considers it to be the ferrous analogue of antigorite. Deer et al. (1962) relate greenalite structurally to chamosite and classify it as septechlorite. Flaschen and Osborn (1957) have synthesized greenalite and found it stable at temperatures as high as 470°C.

Greenalite is a common silicate in the Biwabic and Gunflint iron-formations of the Lake Superior region and in the Roper River ironstones of Australia (Cochrane and Edwards, 1960). Greenalite is also locally abundant in the Sokoman Formation. Perrault (1955) mentions beds with up to 80 per cent greenalite in the Dyke Lake area. Elsewhere it is scarce or absent.

In the Knob Lake area, there are two distinct types of greenalite which for descriptive purposes are designated as "fine-grained greenalite" and "coarse-grained greenalite".

Fine-grained Greenalite

The fine-grained greenalite is similar to the type material from the Lake Superior region. It has a distinct green colour and is micro- to cryptocrystalline. In large masses it is weakly but distinctly anisotropic; where altered or finely disseminated it appears to be isotropic.

In the Knob Lake area, greenalite is most common in the green cherts of members VIII, IX and X, but even here it rarely makes up more than a few per cent of the beds. Cherts with more than 5 per cent of greenalite are uncommon. The greenalite occurs mainly as very fine, uniform disseminations in chert granules. Rare spherites of greenalite are seen here and there in some of the granules as well as in some of the chert which surrounds them. Minnesotaite is generally present in the same rocks as greenalite, but the two minerals are not closely associated. The minnesotaite occurs as fibres or fibrous clusters haphazardly disseminated in chert with or without greenalite. Less commonly, it is seen as fibres which protrude outward from the surfaces of some greenalite spherites and granules.

In the minnesotaite-magnetite facies of member III, greenalite occurs as irregularly shaped aggregates in chert and chert-minnesotaite granules and as dusty disseminations in chert. In the silicate facies of member I, greenalite appears in the form of massive, irregular laminae less than 0.5 millimetre thick, which alternate with thicker siderite-minnesotaite layers; and also as granules with various proportions of chert and minnesotaite. Some greenalite is probably present also in the green chert layers interbedded with massive layers of magnetite in the upper part of the silicate facies, but it is so finely disseminated and in such small quantities that it cannot be identified.

Coarse-grained Greenalite

The diabase dykes which cut across the Sokoman Formation and other rocks in the area are poorly exposed and not well known. Outcrops of the iron-formation near the contacts of the dykes are equally scarce so that little

Table 8. X-ray Powder Data for Greenalite (Fe-Ka radiation, Mn filter).

Mesabi Range Lake Superior District				Sokoman Formation Knob Lake Area			
Greenalite (Blake, 1964)		Greenalite (Youell and Steadman, 1961)		Greenalite		Mineral "X"	
dA	I	dA	I	dA	I	dA	I
7.2	100	7.21	70	7.18	100	7.14	100
3.6	50	3.60	40	3.60	60	3.57	50
2.79	20	2.77	5	2.78	20	2.87	10
2.74	10						
2.58	70	2.59	100	2.60	90	2.66	80
2.46	10						
2.34	5						
2.20	30	2.20	90	2.21	70	2.23	60
2.13	5						
2.07	5						
1.82	10	1.819	50	1.822	35	1.83	40
1.73	5						
1.60	30	1.602	60	1.612	60	1.66	30
1.56	20	1.565	60	1.573	40	1.62	20
1.54	10						
1.51	5	1.514	10	1.514	10	1.51	20
		1.454	10	1.471	5	1.41	5
1.38	5						
1.37	10			1.369	5		

Table 9. Chemical Analyses of Greenalite

	A	B	C
	Mesabi, Minnesota	Roper River, Australia	Hope Lake, Knob Lake Area
SiO_2	32.02	32.27	35.87
TiO_2	-	0.05	0.25
Al_2O_3	1.01	3.03	0.51
Fe_2O_3	22.95	14.83	14.55
FeO	29.15	36.00	35.18
MnO	-	0.87	1.76
MgO	5.34	2.11	3.65
CaO	-	tr.	0.03
Na_2O	-	0.14	1.4
K_2O	-	0.10	0.00
H_2O+	9.52	8.72	6.66
H_2O-		2.37	0.04
F	-	-	0.06
Total	99.99	100.49	99.96

A. Leith's analysis (Leith, 1903) adjusted by Gruner (1936).

B. As reported by Cochrane and Edwards (1960).

C. Sample Z-91. H.B. Wiik, analyst. Sample submitted for analysis was at best 94-94% pure. Recognizable impurities were: 5-6% mineral "X" and trace amounts of unidentified opaque material. H.B. Wiik also detected small amount of sulphur in the analyzed material.

is known about the extent to which these iron-rich sediments were metamorphosed. A few isolated outcrops near the east contact of a dyke on the ridge north of Hope Lake are the only exposures seen by the writer in which the iron-formation showed obvious signs of metamorphism. The rocks a few feet from the dyke are fine- to medium-grained, thinly but indistinctly bedded and consist mainly of green iron silicates. The rocks are clearly coarser grained, more massive in appearance and of different composition than the laminated to banded, microcrystalline chert-siderite and chert-siderite-minnesotaite rocks exposed farther from the dyke.

The dominant mineral in the metamorphosed rocks, unlikely as it may seem, is greenalite. The X-ray pattern of the mineral is almost identical with the type material from the Lake Superior district (Table 8), and its chemical composition closely matches the two other published analyses of greenalite (Table 9).

The specimen (Z-91) from which the greenalite concentrate was prepared for analysis consists of approximately 80 per cent of greenalite, 12 per cent of an unknown light greenish grey mineral (mineral "X"), 7 per cent of ankerite (confirmed by X-ray), a few grains of quartz and trace amounts of an unidentified black, dusty material which occurs as small diffuse patches in greenalite. The typical texture of the rock is shown in Figure 73.

The finest grained greenalite is green, nonpleochroic and nearly isotropic. The largest tabular porphyroblasts (200 to 800 microns long), however, are pleochroic from pale yellow to green (similar pleochroism reported by Cochrane and Edwards, 1960) and show first order birefringence colours. The greenalite is optically negative, has a small 2V = (5° - 10° ?) and good basal cleavage (Fig. 74).

Mineral "X" occurs as very irregular to elongate fine-grained patches surrounded by greenalite or as discrete aggregates of stubby, tabular grains partly enveloped by rims of ankerite. In thin section, the mineral is colourless, or nearly so, with a slight yellowish tinge. It is optically negative with 2V close to zero and is not pleochroic.

The optical properties and X-ray pattern (Table 10) of mineral "X" could not be matched with any known mineral. However, with the exception of its colour and pleochroism, the mineral is very similar to greenalite. It is probably a magnesian or a manganoan variety of greenalite.

Origin

Greenalite is one of the few silicates in Precambrian iron-formations which is accepted by nearly all geologists as a primary sedimentary mineral. The fine-grained greenalite in the Sokoman Formation is probably of similar origin.

The coarse-grained greenalite is undoubtedly metamorphic. The greenalite rocks near the dyke are on strike of unaltered iron-formation composed mainly of chert and siderite, and it is probable that most, if not all, of the greenalite has formed by reaction between these two minerals. A simple equation which would express the reaction is:

$$\underset{\text{siderite}}{\underline{3\,FeCO_3}} + \underset{\text{chert}}{\underline{2\,SiO_2}} + 2\,H_2O \rightarrow \underset{\text{greenalite}}{\underline{Fe_3Si_2O_5(OH)_4}} + 3CO_2$$

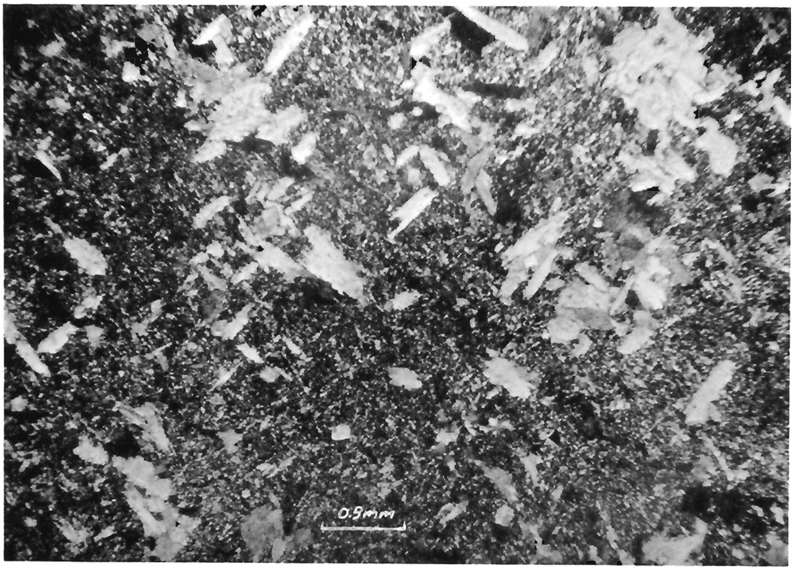


Figure 73. Photomicrograph showing typical porphyroblastic texture of greenalite rock. The large porphyroblasts and most of the finer grained matrix are greenalite (polars crossed).

Figure 74. Cleavage in the coarse-grained greenalite. Photographed with electron microscope.

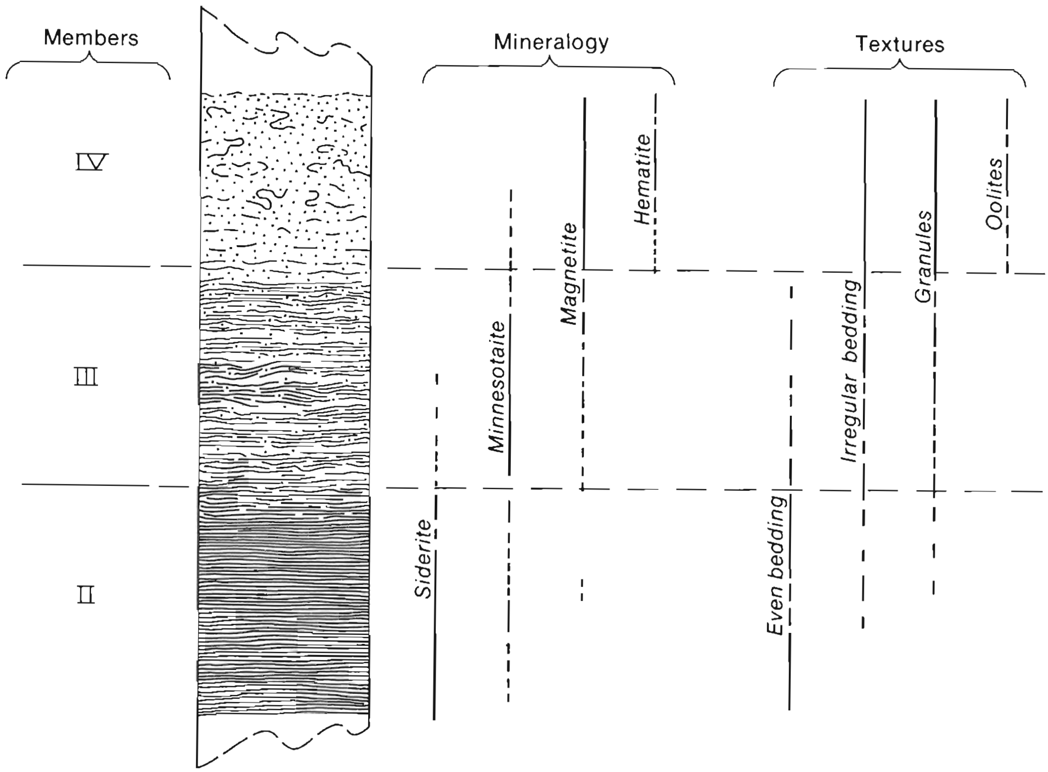


Figure 75. Relationship of minnesotaite to other minerals and to sedimentary textures. The stratigraphic section shown is just southeast of GSC Wishart Lake. (Lat. 54°42'54", Long. 66°48'37".)

The ankerite and mineral "X" (colourless greenalite) are probably the byproducts of the reaction, with the necessary MgO, CaO and MnO supplied by the impure siderite. The composition of siderite in the iron-formation near Hope Lake is unknown. It is likely, however, that the siderite, like other siderites in Precambrian iron-formations, contains several per cent of MgO and smaller amounts of MnO and CaO (see Table 13).

MINNESOTAITE

Minnesotaite is the most common silicate of the Sokoman Formation. The largest concentrations of minnesotaite are in the silicate facies of members I, III (minnesotaite facies), VI and VII in which it accounts for 40 to 70 per cent of all minerals. Minnesotaite is also present in other silicate, silicate-carbonate and silicate-oxide facies of the iron-formation but in subordinate amounts.

The typical minnesotaite-rich rock (members I, III, VI) is thinly though irregularly bedded (Figs. 33, 34). Granules are common in the thicker more irregular beds, but crossbedding and conglomeratic beds are rare. Oolites are unknown.

The most common associates of minnesotaite are chert, siderite and magnetite. Hematite is rarely present in the same rocks with minnesotaite.

Generally, rocks which contain more than a few per cent of hematite have little or no minnesotaite. The converse is also true. Mixed rocks, composed of haphazardly mixed granules, pebbles or fragments of different mineralogical composition, are the only exceptions. Even in such hybrid rocks, however, the association between minnesotaite and hematite is not as close as it might appear to be. The transition zone of member IV, where it overlies the minnesotaite-rich rocks of member III, is the best example. In many beds of the transition zone minnesotaite and hematite are equally abundant but not intimately mixed with one another. Minnesotaite is confined to the central parts of the granules and hematite to the periphery of the granules and the chert which surrounds them. In some of the granular-conglomeratic beds of member V, granules composed mainly or wholly of minnesotaite are in contact with pebbles or granules of jasper or with granules composed of various proportions of chert, magnetite and hematite. Minnesotaite is present also in the chert interstitial to the granules of different composition, but again only where hematite is scarce or absent.

The best illustration of the relationship of minnesotaite to other iron minerals and to the sedimentary textures of the rocks is found in stratigraphic sections of members II to IV, in the ore zone (Fig. 75).

In the sequence of mineralogical changes minnesotaite occupies a position intermediate between siderite and magnetite. The sequence siderite-minnesotaite-magnetite is typical of the Sokoman Formation. As Figure 75 shows, the mineralogical changes closely follow the changes in sedimentary textures which reflect the changes in the sedimentary environment from low energy, reducing to high energy, oxidizing. A similar relationship between siderite, minnesotaite, magnetite and sedimentary textures is seen in the facies changes in member III (see Fig. 29).

Minnesotaite-rich rocks are easily identified in the field by the characteristic yellow-orange colour imparted to the outcrops by the alteration of minnesotaite. On fresh surfaces minnesotaite is light greenish grey, olive green to nearly white, and finely fibrous to massive in appearance. In thin section it is colourless to very light green, in places with a slight yellowish cast. With the exception of the darkest varieties, the mineral is not pleochroic. It occurs as dense, felty-massive to fibrous-acicular concentrations in bands, granules and irregular aggregates; as clusters of decussate or radiating fibres (needles); as disseminated acicular crystals; and rarely as cross-fibres in monomineralic or chert-minnesotaite veinlets. The length of individual crystals ranges from a few microns to approximately 150 microns. Larger crystals are rare. The finest grained minnesotaite occurs in massive, thin layers and in central parts of granules; the coarsest occurs near the periphery of granules or anywhere coarser grained chert is also present.

The only known chemical analyses of minnesotaite are listed in Table 10. The first three are from cherty Precambrian iron-formations, and the last from a lead-zinc replacement deposit in Paleozoic limestone.

Origin

Microscopic examination of minnesotaite-bearing rocks shows clearly that much of the minnesotaite is well crystallized and that at least some of the crystallization took place after the sediment was formed. To this extent minnesotaite is not a primary mineral. However, just because the texture of

Table 10. Chemical Analyses of Minnesotaite

	Mesabi Range Gruner (1946)	Cuyuna Range Blake (1964)	Knob Lake Area Sokoman Formation**	Bluebell Mine Perrault and Hebert (1968)
SiO_2	51.29	51.79	50.14	51.47
Al_2O_3	0.61	1.46	0.37	1.57
TiO_2	0.04	0.08	0.02	
P_2O_5		0.08	0.01	
Fe_2O_3	2.00	0.61	2.00	3.25
FeO	33.66	35.65	36.68	30.50
MnO	0.12	0.78	0.27	1.83
MgO	6.26	3.21	4.39	5.10
CaO	0.00	0.10	0.10	0.00
Na_2O	0.08	0.06	0.05	
K_2O	0.03	0.44	0.15	
H_2O+	5.54	5.03	5.24	5.88
H_2O-	0.24	0.03	0.44	0.16
F		0.02		
CO_2		0.14		
Total	99.87	99.48	99.91	99.74

**Minnesotaite from a cross-cutting veinlet in the lower part of Sokoman Formation (member III?). Submitted for analysis by G.A. Gross, Geol. Surv. Can. Analysts: J.A. Maxwell; K_2O and Na_2O by J.L. Bouvier; CaO by W.F. White, Geol. Surv. Can.

minnesotaite is not primary it does not mean that the original substance had to be something other than minnesotaite. Minnesotaite could have been deposited initially as a very fine grained, possibly gel-like precipitate as proposed by Gruner (1946), and simply recrystallized after sedimentation. The rare veinlets of minnesotaite could have been formed by solutions expelled during compaction and crystallization of the sediment.

The possibility that minnesotaite has formed at the expense of some other pre-existing silicate cannot be discounted although convincing evidence of such alteration is lacking. In view of minnesotaite's mineral associations and of the close accord between mineralogical changes and the sedimentary textures which reflect primary differences in the depositional environment, a primary origin of minnesotaite is the simplest and the most likely alternative.

STILPNOMELANE

Stilpnomelane is one of the more common silicates of the Sokoman Formation in the Knob Lake area, but it is not well known. Most of it occurs in areas of little economic interest and it has therefore received little attention.

Identification

Stilpnomelane is a difficult mineral to identify and can be overlooked even in thin section. Most needle-like to blade-like crystals larger than about 50 microns present no problem. The high birefringence and strong pleochroism (pale yellow to brownish, smoky green of ferrostilpnomelane and yellow to dark brown, almost black, of ferristilpnomelane) are sufficient for positive identification. Such textbook examples of stilpnomelane, however, are uncommon. Most stilpnomelane in the Sokoman Formation is extremely fine grained (under 15 microns) so that the birefringence is moderate to low and pleochroism weak to imperceptible. Such stilpnomelane can be easily mistaken for chlorite, minnesotaite or even greenalite, particularly when it is imbedded in fine-grained siderite, impregnated with opaque inclusions or strongly altered. In many instances identification is impossible without X-ray analysis.

Stilpnomelane gives a poor X-ray pattern but its strongest reflection at approximately 12A (about 9.2 degrees 2θ) is diagnostic. It clearly distinguishes stilpnomelane from all other minerals in the iron-formation, even the often closely associated chlorites and minnesotaite. Examples of the diffraction patterns in the low 2θ region critical for identification of stilpnomelane are shown in Figure 76.

In the field, the brown weathering and argillaceous appearance are useful guides in recognition of the stilpnomelane-rich rocks in the Knob Lake area (see Figs. 30, 62 and 63). The criteria, however, are by no means infallible and should be used with caution. Weathered rocks composed mainly of siderite, for example, are invariably brown. Mixed with fine-grained silicates, such rocks may also be argillaceous in appearance.

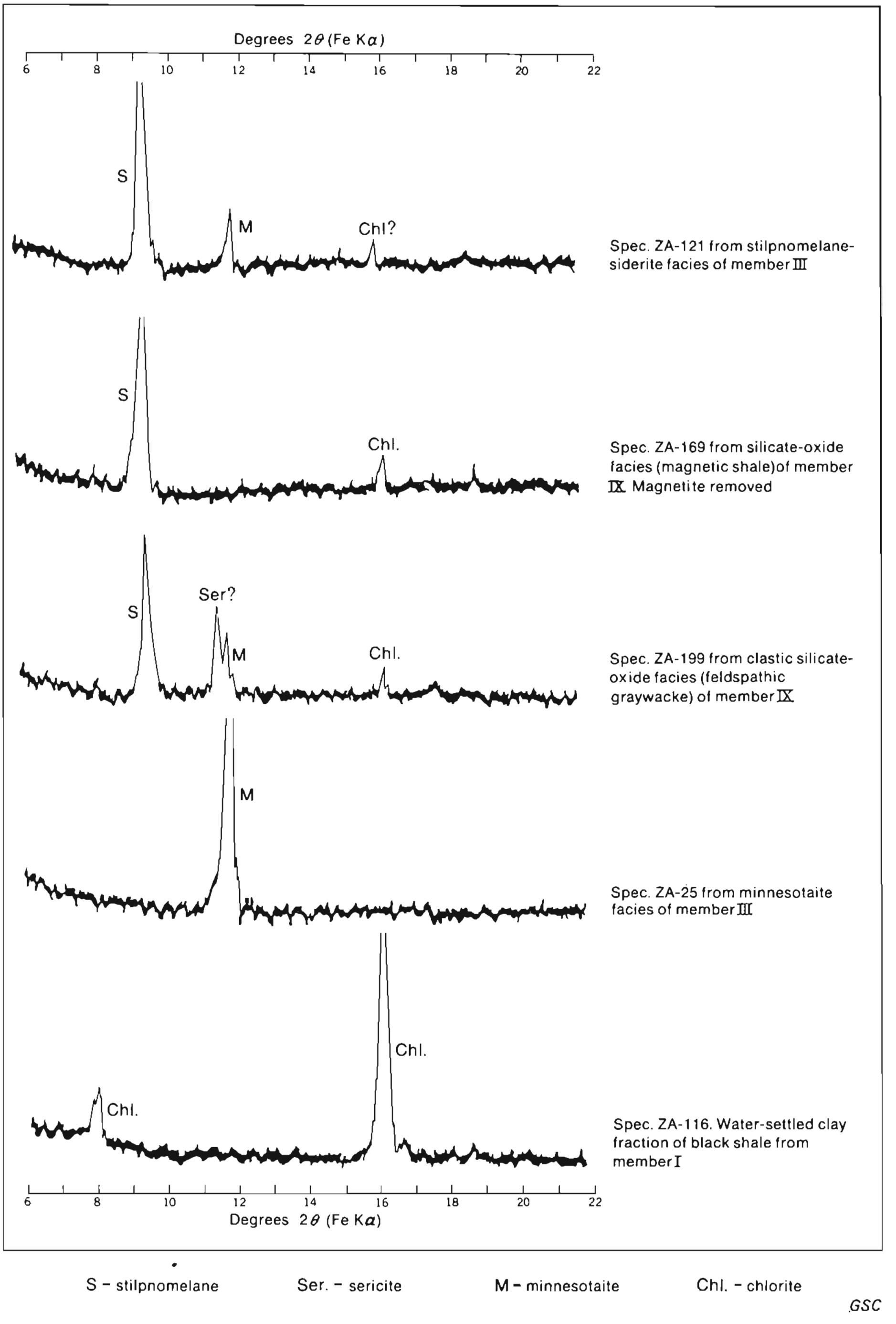


Figure 76. 2θ range of diffractometer patterns critical in identification of stilpnomelane.

Distribution, Texture and Mineral Associations

Stilpnomelane is most common in the northwestward-trending zone which extends outward from the Nimish area near Petitsikapau, Dyke and Astray lakes, and passes between Knob Lake and Iron Arm, the area outlined by the stilpnomelane-siderite facies of member III (Fig. 29). With the exception of member IX, all units of the iron-formation outside this area contain little or no stilpnomelane.

Member IX and the stilpnomelane-siderite facies of member III are the main stilpnomelane-bearing units in the area. Stilpnomelane is probably also common (?) in the brown clastic and argillaceous beds in members V, VII, VIII, X and in some of the green (brown when weathered) silicate layers in member II, in the same general area as the stilpnomelane-siderite facies of member III.

In the thinly layered stilpnomelane-siderite facies of member III, most stilpnomelane is micro- to cryptocrystalline and has a felty to finely fibrous texture. It is closely associated with chert, siderite and in places with minnesotaite. Crosscutting veinlets, a millimetre or two in width, composed of stilpnomelane or of stilpnomelane and chert are rarely seen in some of the thin-bedded rocks. In fresh rocks the stilpnomelane is green (ferrostilpnomelane) but on weathered surfaces it is invariably brown (ferristilpnomelane).

In member IX, the texture, mineral association and composition (ferrous-ferric ratio) are more varied. In the carbonate facies of the member (Fig. 61), stilpnomelane is seen as green fibrous to felty patches irregularly intergrown with chert or siderite, and less commonly as randomly scattered, large (up to 0.3 mm) acicular crystals (prismatic sections of thin plates) in chert or carbonate. Elsewhere in the member (silicate-oxide facies, Fig. 61), stilpnomelane is generally brown, although green and brownish green varieties are also present. Magnetite is the most persistent associate of stilpnomelane in the silicate-oxide facies. Other minerals present in highly variable amounts are chert, minnesotaite and chlorite. The shales and greywackes of the clastic facies contain also feldspar, detrital quartz, less commonly sericite (?) and rarely ilmenite. The stilpnomelane occurs as dense, felty to fibrous, massive or granular layers, generally speckled by disseminated magnetite, and a part of the fine-grained "argillaceous" matrix in the clastic rocks. Radial and decussate aggregates of scaly to acicular crystals up to 50 microns in size and larger aggregates are rare.

In most clastic, stilpnomelane-bearing rocks examined by the writer, the feldspars are at least partially replaced by fine-grained silicates. In some rocks, 20 to 90 per cent of the individual feldspar grains as large as 150 microns are replaced. The fibrous silicates replacing feldspar are usually oriented at right angles to the grain boundaries. The fibres project inward from the periphery of grains so that the central parts of the grains are the last to be replaced. Detrital quartz, if present, is also fringed by silicates, but to a much lesser extent than feldspar.

In some shales and greywackes, the silicates which replaced feldspar appear to be chiefly stilpnomelane; in others the silicates could not be identified. They may be stilpnomelane, chlorite, sericite(?) and possibly minnesotaite. X-ray analyses of some greywackes show that several of the silicates are present in the same rock, but it is often impossible to differentiate them in thin section.

Origin

Stilpnomelane is a common silicate of many iron-rich rocks, particularly those which have been subjected to low-grade regional metamorphism. It also occurs in veins which cut stilpnomelane-bearing rocks and in contact metamorphic zones near dykes and sills (James, 1966). Although stilpnomelane is a common metamorphic mineral, not all stilpnomelane in metamorphosed sediments is necessarily of metamorphic origin. Gruner (1946) and White (1954) believe that the stilpnomelane in the Biwabic Formation is a primary-diagenetic mineral which crystallized from a primary gel of the appropriate composition. Perrault (1955) has suggested that some of the fine-grained stilpnomelane in the Dyke Lake area may also be primary. The recent discovery of stilpnomelane in unmetamorphosed cherty iron-formation of Carboniferous age (Schultz, 1966), shows that stilpnomelane can form at low temperature and pressure. Trendall (1966, p. 1457) estimated that the temperature to which the stilpnomelane-bearing Brockman Iron-Formation was subjected after sedimentation probably did not exceed 160°C. Since stilpnomelane was one of the first-formed components of the iron-formation, it is probably diagenetic rather than metamorphic (Trendall, 1966).

Most of the stilpnomelane in the Knob Lake area is confined to the units of the iron-formation which have been contaminated by volcanic and detrital material derived from the Nimish volcanic area just southeast of the Knob Lake basin. As discussed under "Member III", the stilpnomelane is believed to have formed by the reaction of this material with the iron- and silica-saturated water of the Knob Lake basin. Most of the reaction, as suggested by the peripheral replacement of feldspar, has probably occurred after deposition, conceivably by interaction of the sediment with its interstitial water. However, the reconstitution of the finest tuffaceous and clastic material could very well have begun during sedimentation. The formation of the stilpnomelane veinlets and the growth of larger crystals could be late diagenetic or metamorphic phenomenon.

The main and probably the only role of volcanic material in the formation of stilpnomelane was its contribution of alumina – one of the essential components of stilpnomelane. Iron and silica, the other two necessary constituents were readily available throughout the deposition of the Sokoman Formation, but alumina was not. Precambrian iron-formations are noted for their low content of alumina, the average being 1.6 per cent (Lepp and Goldich, 1966). Most units of the Sokoman Formation (see Gross, 1968, p. 21, 28) contain less than 1 per cent of alumina, whereas stilpnomelane requires 4 to 7 per cent. The main reason why stilpnomelane is absent from such non-clastic (uncontaminated) silicate facies as those of members VI, VII and the minnesotaite and minnesotaite-magnetite facies of member III, could be that there was not sufficient alumina for stilpnomelane to form.

Although the availability of iron, silica and alumina is important, it is not the only prerequisite for the formation of stilpnomelane. The black carbonaceous shales of member I (clastic silicate-sulphide facies), for example, contain all the necessary chemical components, but very little or no stilpnomelane (Fig. 76, spec., ZA-116). This is also true of the carbon-rich shales closely associated with the Precambrian iron-formations of the Lake Superior region and of those in Africa and Australia. This strongly suggests that the oxidation-reduction capacity of the environment is also important in determining whether stilpnomelane will or will not form. Apparently stilpnomelane cannot form in very strongly reducing environments where carbon is stable, even when all the essential chemical constituents of stilpnomelane are readily available.

CROCIDOLITE

The alkali amphiboles riebeckite, magnesioriebeckite and crocidolite are neither as common nor as well known as most other constituents of Precambrian iron-formations, but they are by no means rare. The extensive crocidolite-riebeckite deposits of South Africa and Western Australia are some of the largest accumulations of alkali amphiboles in the world. Other Precambrian iron-formations with locally abundant amphiboles of this type are the Krivoi Rog iron belt in the Ukrainian Shield (Semenenko et al., 1956), the Imataka Formation in the Guayana Shield of Venezuela (Ruckmick, 1963), the Negaunee Formation of the Lake Superior region (Tsu-Ming Han, 1971, pers. comm.), and the Sokoman Formation of the central Labrador Trough.

The crocidolite deposits of the Labrador Trough are similar in many respects to the much larger deposits in South Africa and Western Australia. The Hook Lake and Trough Lake deposits (Fig. 78) have been described by Kirkland (1950) and Neal (1949) but the others are virtually unknown. The presence of crocidolite near Knob Lake, near Attikamagen Lake and on Dolly Ridge has not been previously reported.

Composition and Nomenclature

Riebeckite, magnesioriebeckite, crossite and glaucophane are closely related soda amphiboles (Fig. 78). There is a complete solid solution of riebeckite with magnesioriebeckite and of magnesioriebeckite through crossite with glaucophane (Deer et al., 1962). Amphiboles of ferroglaucophane composition are unknown.

Crocidolite is the fibrous analogue of riebeckite-magnesioriebeckite (Fig. 78). Rhodusite is the less commonly and somewhat inconsistently used name for amphiboles which are close to the field of magnesioriebeckite composition.

The chemical analysis of crocidolite from the lac Denault area, the first analysis of crocidolite from the Labrador Trough, is given in Table 11. The material analyzed was hand-picked under the microscope from a thin asbestiform layer obtained from the jaspilitic part of member IV at the southern extremity of the outcrop area near lac Denault.

The crocidolite from the lac Denault locality, as shown by the following formula calculation and the plot in Figure 78, is a magnesioriebeckite with nearly half of the magnesium replaced by ferrous iron. The amount of ferric iron is in excess of that required by the ideal riebeckite-magnesioriebeckite structure (i.e., 2.33 instead of 2.00), whereas the sum of ferrous iron, magnesium and manganese is 0.34 less than the stoichiometric norm (2.66 instead of 3.00). It is possible that the excess of ferric iron reported in the chemical analysis is due to oxidation of the original ferrous iron during sample preparation (in the process of grinding and pulverizing the sample) or by supergene alteration. If so, the composition of the lac Denault crocidolite would be close to the midpoint between riebeckite and magnesioriebeckite and its formula would approach $Na_4Fe_3^{+2}Mg_3Fe_4^{+3}Si_{16}O_{44}(OH)_4$.

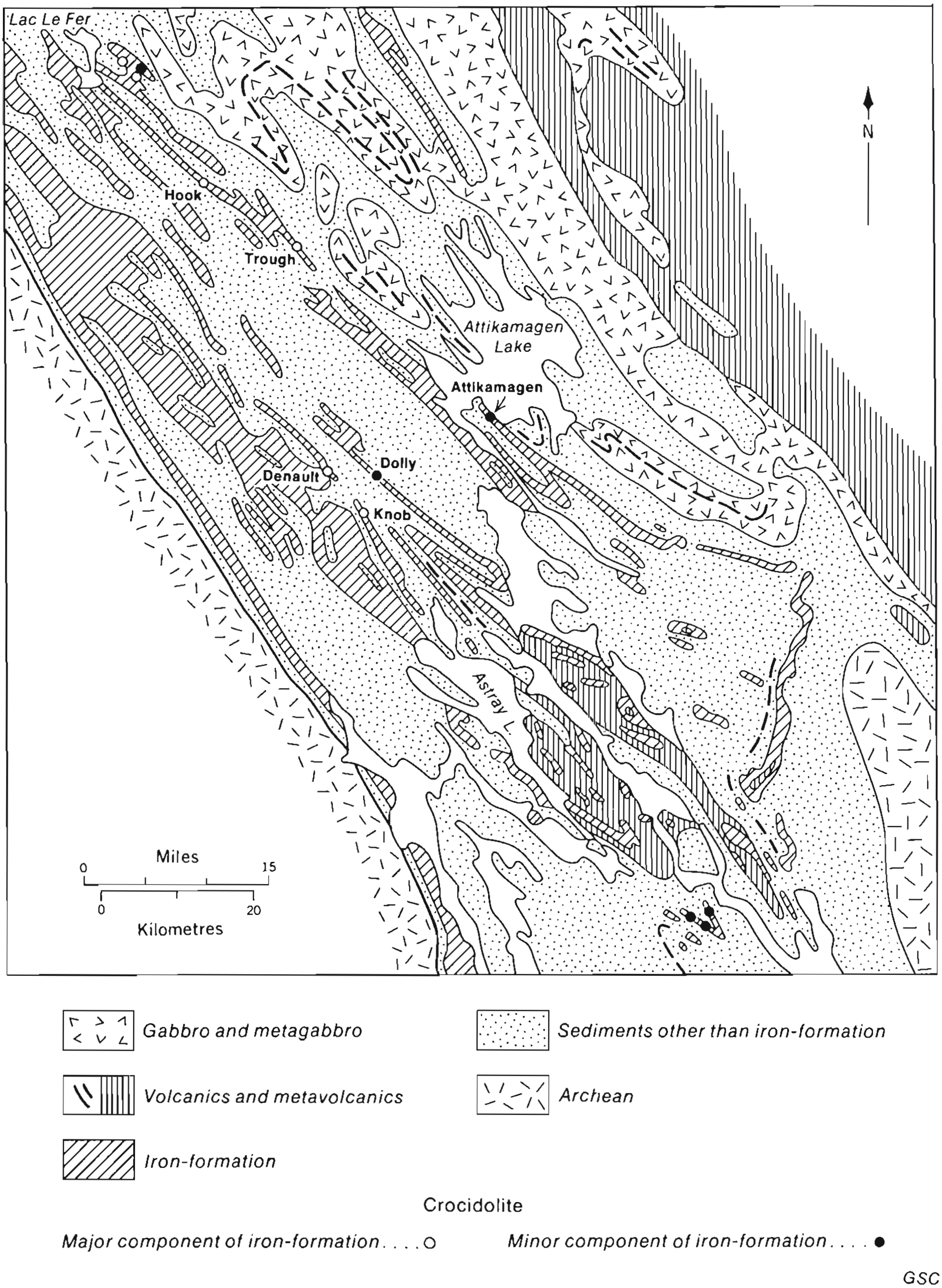


Figure 77. Crocidolite occurrences in the central Labrador Trough. Geology after Gross (1968) and Frarey (1961).

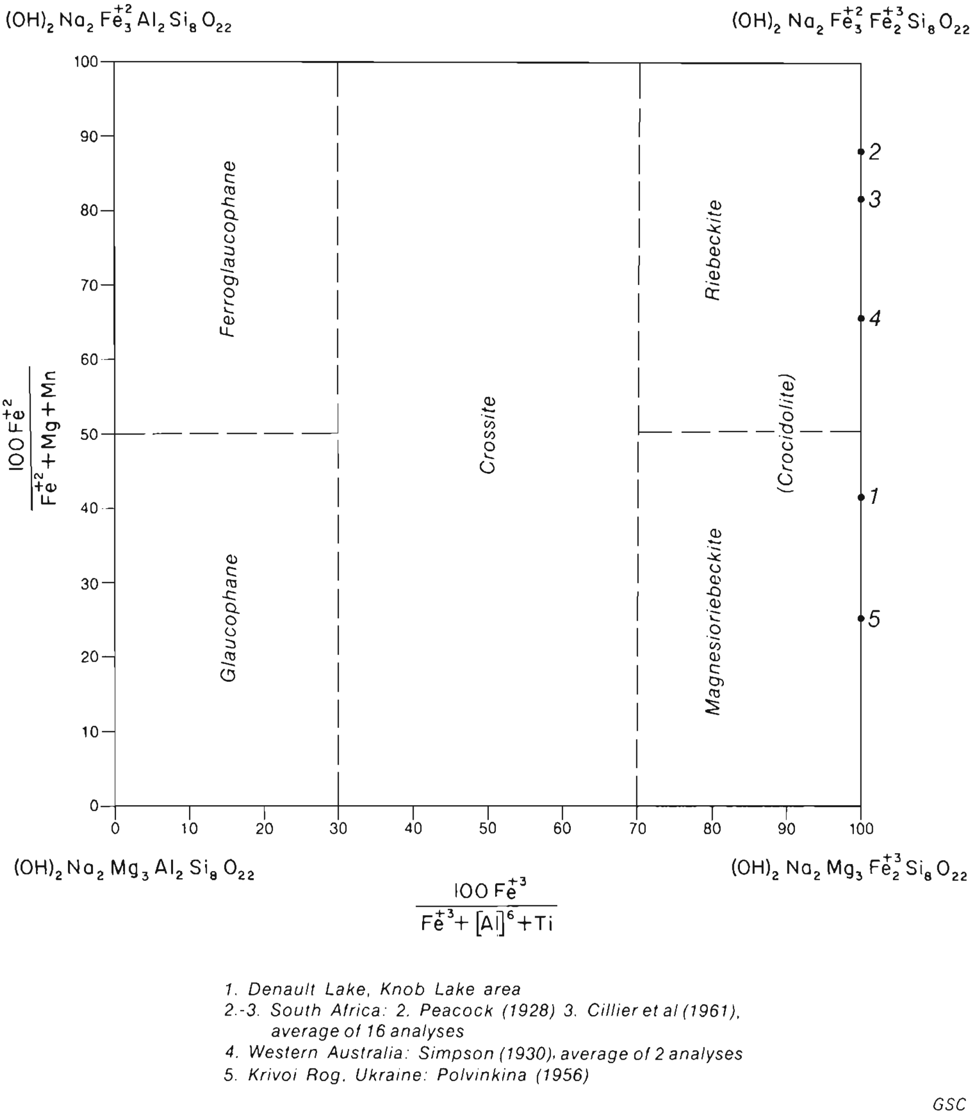


Figure 78. Composition of crocidolites in Precambrian iron-formations. Classification of amphiboles after Miyashiro (1957), and Deer et al. (1962).

Stratigraphic and Structural Setting

At the Denault, Knob, Attikamagen, Dolly, Hook and Trough localities (Fig. 77) crocidolite occurs only in two of the stratigraphic units of the Sokoman Formation - members IV and V. The other deposits of crocidolite are also confined to the iron-formation, but their precise stratigraphic position is unknown.

Table 11. Chemical Analysis of crocidolite and crocidolite-bearing Iron-Formation from lac Denault area

	A Crocidolite	B Crocidolite-bearing Iron-formation
SiO_2	54.14	55.37
TiO_2	0.07	0.07
Al_2O_3	0.26	0.14
Fe_2O_3	21.07	33.56
FeO	9.27	7.14
MnO	0.06	0.06
MgO	7.39	2.49
CaO	0.11	0.37
Na_2O	5.1	1.12
K_2O	0.00	0.05
H_2O+	2.02	0.51
H_2O-	0.02	0.00
F	0.09	0.04
Total	99.60	100.92

A. Asbestiform crocidolite from member IV.

B. Composite sample of chip samples collected from lower 30 feet of member V.

Analyst: H. B. Wiik.

Table 12. Calculation of the crocidolite formula

	Wt. % of Oxides	Molecular Proportion of Oxides	Atomic Proportion of Oxygen	No. of Anions Based on 24(O, OH)		No. of Ions in Formula		
SiO_2	54.14	0.9010	1.8020	15.904	Si	7.95		8.00
Al_2O_3	0.26	0.0026	0.0078	0.069	Al	0.05		
TiO_2	0.07	0.0007	0.0018	0.016	Ti	0.01		
Fe_2O_3	21.07	0.1319	0.3957	3.492	Fe^{+3}	2.33		5.00
FeO	9.27	0.1292	0.1292	1.141	Fe^{+2}	1.14		
MnO	0.06	0.0008	0.0008	0.007	Mn	0.01	1.51	
MgO	7.39	0.1833	0.1833	1.618	Mg	1.62		
CaO	0.11	0.0020	0.0020	0.018	Ca	0.02	0.11	1.58
Na_2O	5.1	0.0823	0.0823	0.726	Na	1.45		
H_2O+	2.02	0.1121	0.1121	0.989	OH	1.98		
H_2O-	0.02	-	-	-	-			2.02
F	0.09	0.0047	0.0047	0.042	F	0.04		

The crocidolite-bearing member IV near lac Denault, Hook Lake, Trough Lake and Knob Lake consists of thin (less than 1 inch) cherty and iron oxide-rich layers which are interspersed in places with crocidolite layers. Pink to red jasper is common except in the lower part of the member. The upper, main part of member IV at these localities is texturally unusual. The oolites and granules, which are so common in this part of the member in other parts of the Knob Lake area, are scarce or absent. Member IV in the lac Denault area is also unusual in that the upper part of the member is very thin or missing.

The small amounts of crocidolite identified in thin sections of specimens from the Dolly Ridge and Attikamagen Lake localities occur as disseminated fibres in the more common granular (Dolly Ridge) and oolitic (Attikamagen Lake) beds of member IV.

The crocidolite-bearing member V of the different localities resembles the conglomeratic-banded facies of member V in many other parts of the Knob Lake area where no crocidolite is present. It is a texturally variable unit composed of irregularly bedded, granular and locally conglomeratic chert-iron oxide beds with variable amounts of crocidolite, and of very thin and massive layers composed mostly of iron oxides, chert or jasper. The thin chert and iron oxide layers are particularly common near Hook Lake and Trough Lake. At the lac Denault and Knob Lake localities, member V consists chiefly of 3- to 10-inch-thick beds. In outcrop, these beds appear to be massive, but are actually granular and in places finely conglomeratic. The crocidolite occurs mainly as irregular disseminations in the granular and conglomeratic beds. Thin, nearly monomineralic layers of asbestiform crocidolite are present locally in the thinly bedded (banded) parts of member V near Hook Lake and Trough Lake, but none could be found at the other localities.

The structural setting of the widely scattered crocidolite deposits is varied. South of Knob Lake, the essentially monoclinal strata dip 35 to 40 degrees to the northeast. A few drag folds (most of them in the thinly layered members III and IV), a few joints and one small fault in the lower part of member V, are the only obvious signs of deformation. Near lac Denault, the very gently undulating beds in the crocidolite-bearing iron-formation dip northeastward at 20 to 35 degrees and are cut by several transverse faults of small apparent displacement. Some differential movement has also occurred locally within the thicker massive beds as evidenced by the preferred orientation of the long fibres of crocidolite. At Hook Lake, the crocidolite-bearing iron-formation is in the eastern limb of a large anticline. The strongly sheared and jointed strata dip 25 to 35 degrees to the northeast. Intense shearing and local, small-scale crumpling of the thin beds suggest the presence of at least two faults which parallel the northwesterly strike of the strata in the upper part of the crocidolite zone. Small slips are numerous in various parts of the zone, most of them located along thin layers of crocidolite which have been smeared out into unctuous, slickensided material. The iron-formation just east of Trough Lake is strongly and, in places, intricately folded. The folds in the crocidolite-bearing iron-formation range from small crumples to folds with amplitude of several feet. These folds appear to be subsidiary structures (drag folds) of a much larger northwest-trending anticline. Joints, many of them roughly parallel to the axial planes of folds, are abundant and locally closely spaced. No faults or prominent shear zones are present, but most of the asbestiform layers show some evidence of deformation brought about by differential movement along the contacts of the layers.

A comparison of the structure of various crocidolite localities shows clearly that the abundance and distribution of crocidolite are not related to intensity of deformation nor to any particular type of structure – folds, faults or joints. The control appears to be strictly stratigraphic.

Textures and Mineral Associations

The textures of the crocidolite in the Central Labrador Trough resemble the textures of the South African and Western Australian crocidolites. Nearly every texture described by Hall (1930), Miles (1942) and Du Toit (1945) is duplicated at one place or another by the crocidolite in the Sokoman Formation.

Mass-fibre crocidolite (potential crocidolite, mass-fibre crocidolite, riebeckite, mass-fibre riebeckite)[1]. This is the most common type of crocidolite in the Sokoman Formation. The beds of iron-formation with finely disseminated to dense aggregates of crocidolite are massive in outcrop. The fibrous texture of crocidolite is not apparent except on fracture surfaces of beds in which some shearing has taken place. The pale blue colour of freshly broken or cut specimens is characteristic of all rocks with more than 10 to 15 per cent of crocidolite. The colour unmistakably distinguishes crocidolite-bearing beds from all other silicate rocks in the area.

The amount of mass-fibre crocidolite at any one of the localities is highly variable from bed to bed and from place to place within a single bed. Most of it occurs in the irregular, massive-looking granular beds of member V. Dense, nongranular layers composed almost entirely of massive crocidolite are uncommon. In thin section (Fig. 79), the crocidolite is seen as slender fibres haphazardly intergrown with chert or concentrated into felty to massive, nearly opaque aggregates and layers. The fibres range from nearly submicroscopic particles to hair-like, slightly curved fibres a few microns in diameter and several hundred microns in length. In granular beds, the crocidolite is present in both the granules and the matrix which surrounds them, although the amount of crocidolite in each may be very different.

Needle crocidolite (acicular crocidolite, riebeckite). Needle-like crystals of crocidolite, sufficiently large to show (in thin section) the cleavage and diamond-shaped outlines typical of amphiboles, are rarely seen in some of the mass-fibre beds. The needle-like crystals are commonly 10 to 40 microns in diameter and up to several millimetres in length. Some of them have brush-like or feathery terminations. Such crystals are identical to those described by Miles (1942, p. 16).

Asbestiform crocidolite (crocidolite proper, crocidolite). Asbestiform crocidolite is found only in the thinly bedded (banded) parts of the iron-formation. Most of the crocidolite in the middle and upper part of member IV is of this type. Asbestiform crocidolite was not seen in granular, oolitic or conglomeratic beds or along contacts between such beds. Abrupt, well-defined contacts, such as those between the thin massive layers of iron oxides (metallic bands) and chert in the banded parts of the iron-formation, appear to be a prerequisite for the development of asbestiform crocidolite.

---

[1] In parentheses are other names which have been used by African and Australian geologists.

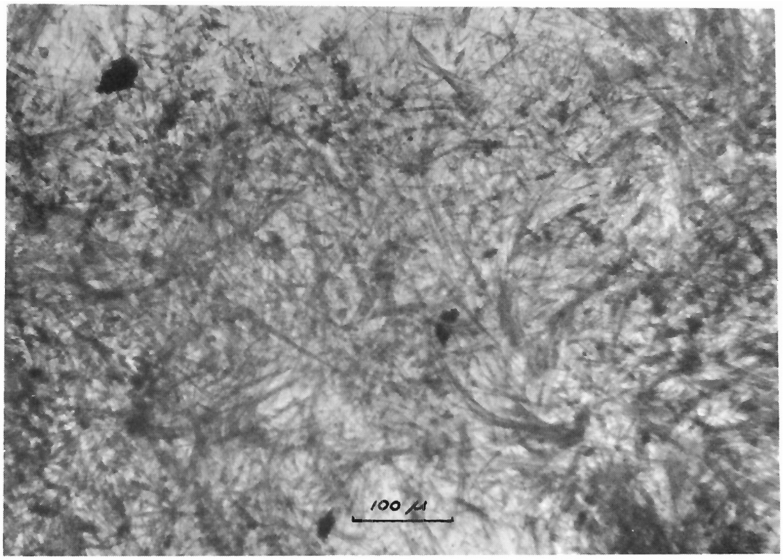


Figure 79. Mass-fibre crocidolite as seen in thin section. Member V, lac Denault area.

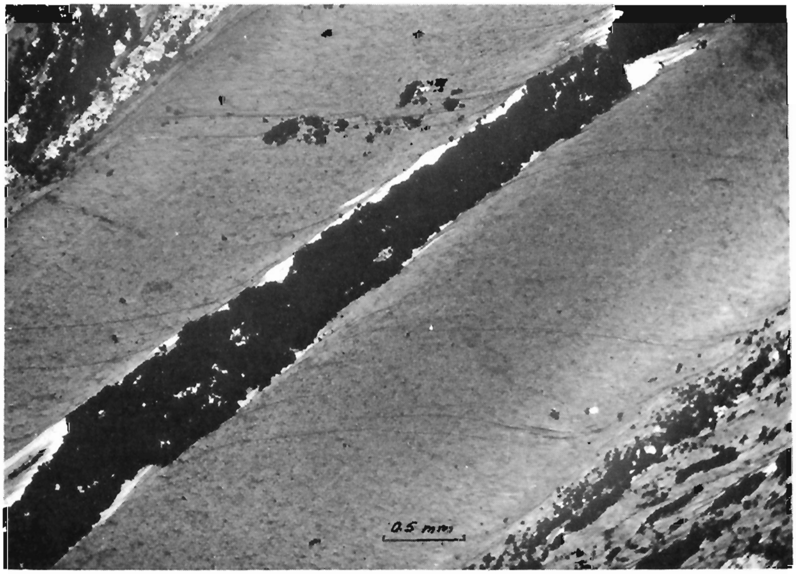


Figure 80. Photomicrograph of asbestiform crocidolite. Member IV, Hook Lake area.

Asbestiform crocidolite is most abundant in the Trough Lake area where asbestiform layers up to half an inch thick are common. Asbestiform layers, or seams as they are sometimes called, are also common near Hook Lake, although at this locality they are rarely thicker than one quarter inch. South of Knob Lake a very small amount of asbestiform crocidolite is present in the banded part of member IV. The rare seams of crocidolite are less than one-eighth-inch thick. Near lac Denault asbestiform crocidolite occurs in layers up to one-third-inch thick at the southern extremity of the outcrop area.

Individual crocidolite seams at any one locality are of variable thickness and, as a rule, cannot be traced for more than a few feet along strike. They pinch and swell, bifurcate or very gradually merge along strike to form composite seams. The contacts of the seams with adjoining chert or iron oxide beds are sharp but commonly irregular. Many of the seams are asymmetrical with one contact being even and the other corrugated. Similar irregularities of asbestiform seams are also common in Africa (Hall, 1930) and Australia (Miles, 1942).

The densely packed, parallel fibres of crocidolite making up the asbestiform seams are usually oriented at 50 to nearly 80 degrees to the bedding. A perfect cross-fibre (90 degree) orientation is typical of the essentially undeformed iron-formations in South Africa and Western Australia, but unknown in the Sokoman Formation where all seams are at least slightly deformed. In cross-section normal to the bedding, most fibres show shapes similar to those illustrated by Figure 80, the form depending on whether differential movement (shearing) has occurred along one or both contacts of the seams. In general, the fibres in the thickest, and apparently the most competent layers, show the greatest amount of curvature (drag) near the contacts; whereas the fibres in very thin layers show the greatest tendency to be tilted en masse towards the plane of bedding (and shear). Seams with fibres oriented parallel to or at very small angles to the bedding (slip-fibre crocidolite) are most common in the Hook Lake area where the greatest amount of shearing occurred. Most slip-fibre crocidolite appears to have formed by shearing of very thin asbestiform seams and some by shearing of thin layers of mass-fibre crocidolite. Crosscutting veinlets, a millimetre or less in width, are not uncommon in the Hook Lake and Trough Lake areas, although on the whole they are quantitatively insignificant. Such veinlets collectively account for probably not more than 1 per cent of all crocidolite in the Sokoman Formation.

Chert and magnetite are the closest associates of crocidolite. This is especially true of the asbestiform variety. Secondary iron oxides, such as limonite, soft red hematite and martite, are confined to weathered and altered rocks. Primary hematite is rarely a major component of some of the crocidolite-bearing beds in member IV. Thin beds and lenses of hematitic jasper are also found in member V. The rare carbonate is ankerite. The only silicate, other than crocidolite, is minnesotaite. The minnesotaite is rarely present in more than accessory amounts. It occurs mainly as fibrous clusters in chert-magnetite granules which contain little or no crocidolite.

Origin

The amphiboles of the riebeckite-magnesioriebeckite series from under a wide range of pressure and temperature conditions in environments that range from magmatic to sedimentary. Riebeckite is confined largely to igneous rocks, principally granites and syenites, but is also known as a metamorphic and metasomatic mineral. Authigenic riebeckite has been recently reported by Allen (1967). Magnesioriebeckite occurs mainly in metamorphic rocks, less commonly in metasomatic deposits and very rarely in igneous rocks. Diagenetic magnesioriebeckite has been described by Beyseyev (1966). The fine-grained riebeckite-magnesioriebeckite in the Green River Formation is believed to be authigenic (Milton and Eugster, 1959).

Crocidolite apparently forms mainly by hydrothermal action and by diagenesis or low-grade metamorphism of soda-rich ferruginous sediments.

The Precambrian Krivoi Rog iron belt in the Ukrainian Shield contains some of the best examples of hydrothermal metasomatism which resulted in the development of soda amphiboles. The northern and northwestern, strongly metamorphosed parts of the belt were locally affected by intense soda metasomatism closely associated with aplitic-pegmatitic granites, granite pegmatites and albitites (Semenenko et al., 1956). One of the main iron silicates of the metamorphosed and metasomatized iron-formation is magnesioriebeckite, most of which has formed by replacement of cummingtonite (Semenenko et al., 1956; Polovinkina, 1953). Some of the closely associated riebeckite is also of metasomatic origin. Semenenko and his co-workers (1956, p. 115, 504) show that in some instances it is the final product of cummingtonite alteration. The sequence of changes they observed is as follows: cummingtonite → Na-amphibole → magnesioriebeckite → riebeckite. Soda metasomatism also affects augite which is transformed into aegirine-augite and finally into aegirine (Semenenko et al., p. 115). The unusual accessory minerals which appear in some of the metasomatized rocks are albite, tourmaline and apatite. The crocidolite, according to Polovinkina (1953), is a late-stage hydrothermal-metasomatic mineral. It occurs as crosscutting veinlets, as fine-grained replacements of the iron-formation and as fibrous encrustations around the pre-existing coarse-grained, nonfibrous riebeckite and magnesioriebeckite.

The crocidolite deposits in South Australia described by King (1961) and those in Rhodesia and Bolivia discussed by Drysdall and Newton (1960) also appear to be of hydrothermal origin. The South Australian deposits (King, 1961) are in dolomites and marbles of Cambrian and Precambrian age. They are localized in zones of intense fracturing and are closely associated with strange, igneous-looking rocks composed mainly of albite, tourmaline and biotite. The crocidolite deposit in Rhodesia is similar (Drysdall and Newton, 1960). It is within a restricted area of a strongly metasomatized dolomitic sediment. The crocidolite occurs in an anastomosing network of veins which show no tendency to parallel bedding or one another. The dolomitic host rock is noted for the presence of albite porphyroblasts and tourmaline, the latter comprising up to 50 per cent of the rock. Mica, apatite and monazite are also present. In Bolivia (Drysdall and Newton, 1960; King, 1961) crocidolite is found in a stockwork of veins within sandstone, slate and dolomite. The associated minerals are various borosilicates including tourmaline and danburite.

The development of crocidolite in the iron-formations of Africa and Western Australia is attributed by most geologists to endogenetic processes. The manner in which crocidolite developed, the method by which the sodium was entrapped in the sediments, and the extent to which the sodium and other elements were concentrated in the process of crocidolitization are uncertain. It is generally agreed, however, that all the components necessary for the formation of crocidolite were derived from within the iron-formations themselves, rather than from an igneous source.

The crocidolite in the Sokoman Formation is probably also of endogenetic origin. The crocidolite-bearing units of the formation show none of the characteristics of known hydrothermal crocidolite deposits. The distribution of crocidolite is stratigraphically rather than structurally controlled, and does not appear to be related to igneous intrusives or to centres of igneous activity. Moreover, there is no evidence of other hydrothermal activity in the area.

SIDERITE

Siderite is the most common carbonate in the Sokoman Formation. It is particularly abundant in the lower part of the formation (members I and II) and locally also in the upper part (carbonate facies of members VII, IX and X).

The typical siderite-rich iron-formation, the carbonate and carbonate-silicate facies of members I, II, III and X is thinly bedded and very fine grained; granules and spherites are scarce or absent. The spherites are usually 20 to 30 microns in diameter, poorly defined and seemingly structureless. Some have minute opaque cores. The spherites resemble similar aggregates in other Precambrian siderite rocks and the sphaerosiderites of younger ironstones. The granules are small (50 to 300 microns in size) and poorly formed. Many are mottled and some are at least partly composed of siderite spherites. The origin of these granules is not clear. The only conclusion that can be reached with any degree of certainty is that they are syngenetic. In finely laminated beds, wispy layers of carbonaceous material sweep around the granules indicating that the granules existed as such prior to compaction.

The siderite in the carbonate facies of members VII and X, unlike the carbonate in the thin-bedded facies, is typically granular and medium to coarse grained. The granules are commonly 0.1 to 3.0 millimetres in size, and oval, lobate or slabby in shape. The size of the anhedral siderite grains, which make up the granules and some of the matrix, ranges from a fraction of a millimetre to several millimetres.

Chert, silicates and carbonaceous rocks are the closest associates of siderite. Magnetite is rarely a major component of siderite-rich rocks. Hematite is generally absent. The sideritic beds in the jaspilite of member I and in the transition zone of member IV, are the only notable exceptions. These beds contain small but persistent amounts of hematite, which from all indications appears to be primary.

Composition

The siderite in Precambrian iron-formation generally contains one to several per cent of magnesium, manganese and a smaller quantity of calcium substituting for ferrous iron. A chemical analysis (J. L. Bouvier, analyst) of siderite concentrate obtained from a single specimen of the carbonate facies of member VII (ZA-64, p. 143) is as follows: 8.63% SiO_2, 1.68% Fe_2O_3, 43.99% FeO, 1.45% MnO, 6.86% MgO, 0.54% CaO, 35.88% CO_2. Microscopic examination of the concentrate showed in addition to siderite and chert a small amount of opaque material, most of which was probably hematite. Trace amounts of silicates also may have been present, but their identification is not certain. X-ray analyses showed siderite to be the only carbonate in the concentrate. Composition of the siderite calculated from the above chemical analysis is shown in column 1 of Table 13. In this calculation, all of the CaO, MgO and MnO were assigned to the carbonate. Similar calculation based on the chemical analysis of a composite sample (analysis A, Table 7) gave nearly identical results (column 2, Table 13).

The composition of siderite in other units of the Sokoman Formation is not known, but it is probably variable. Perrault (1955) has found that the Fe:Mn ratio of siderite in the lac Otelnuk area ranges from 19.5 to 6.8. The variability of siderite composition is also evident from the results published by Huber (1959) and James (1954, 1966; see Table 13).

Some consistent variation is noted by James (1954). He indicates (James, 1954, p. 255) that siderite in some carbonate rocks containing free carbon is high in manganese. The same may be true of the siderite in the lower part of the Sokoman Formation. The carbonaceous siderite-rich rocks of members I and II are in places strongly stained with secondary manganese oxides. This suggests that the siderite from which these oxides were probably derived is abnormally manganiferous.

Origin

Precambrian iron-formations composed largely of chert and siderite are generally accepted as primary sediments which formed in reducing, low energy environments. The mineralogical and lithological associations, as well as the sedimentary textures of most siderite-rich rocks in the Sokoman Formation, suggest a similar origin.

The medium- to coarse-grained siderite in the carbonate facies of members VII and IX contrasts sharply with the microcrystalline siderite in the carbonate and carbonate-silicate facies of members I and II in the Wishart Lake area. The difference in grain size is obviously not related to the depth of burial, intensity of deformation or metamorphism. The difference is strictly stratigraphic and closely related to the primary sedimentary textures of the respective rock-types. It is attributed to the greater quantity of water which promoted recrystallization and grain growth during lithification of the strongly granular and originally more porous carbonate sediments in members VII and IX.

Table 13. Composition of Siderite in Precambrian iron-formation

	Knob Lake Area				Iron River, Michigan						Gogebic, Michigan		Gunflint, Minnesota		Marquette, Michigan	
	1		2		3		4		5		6		7		8	
	% wt.	% mol.	% wt.	% mol.	% wt.	% mol.	% wt.	% mol.	% wt.	% mol.	% wt.	% mol.	% wt.	% mol.	% wt.	% mol.
$FeCO_3$	80.0	75.3	80.9	76.2	78.5	76.4	80.7	77.9	83.8	81.2	80.9	76.8	80.2	75.8	89.0	85.7
$MnCO_3$	2.7	2.5	2.2	2.1	14.0	13.7	9.1	8.9	7.4	7.2	4.1	3.9	0.5	0.5		
$MgCO_3$	16.2	21.0	16.1	20.8	6.8	9.1	9.0	11.9	8.0	10.7	12.9	17.0	12.6	16.4	9.4	12.5
$CaCO_3$	1.1	1.2	0.8	0.9	0.7	0.8	1.2	1.3	0.8	0.9	2.1	2.3	6.7	7.3	1.6	1.8
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

1. Based on analysis of siderite concentrate from specimen Z-64.
2. Based on analysis of composite sample, analysis B, Table 7.

3-8. Taken from James (1966, p. W4).

ANKERITE (FERROAN DOLOMITE)

Perrault (1955) was the first to emphasize the fact that siderite is not the only iron carbonate in the Sokoman Formation. Although on the whole far subordinate to siderite, ankerite is at least locally a major component of the formation. Its composition, distribution and mineral associations, however, are still not well known.

Composition

Perrault (1955) estimates the composition of ankerite in the Sokoman Formation as $Ca\,(Mg_{.60}\,Fe_{.36}\,Mn_{.04})\,(CO_3)_2$. His estimate is based on the optical properties of the carbonate and on the assumption that the carbonate has the same Fe:Mn ratio as the average siderite in the Otelnuk Lake area.

Two partial analyses of samples obtained from two five-foot intervals of crushed drillcore from the upper part of member V in the Knob Lake area are as follows:

	A	B
FeO	8.76	7.59
MnO	0.55	0.41
MgO	2.67	1.94
CaO	6.56	5.17
CO_2	9.96	7.92

Since microscope examination and X-ray analyses of the samples showed no other minerals except chert (quartz), hematite, magnetite and a single phase dolomite-type carbonate, the composition of the carbonate could be readily calculated by assigning all of the CaO, MgO and MnO to the carbonate and combining the remainder of CO_2 proportionally with FeO. The calculated composition of the carbonate is as follows:

	A		B	
	% wt.	%mol.	%wt.	%mol.
$CaCO_3$	52.6	51.8	51.2	51.3
$MgCO_3$	25.0	29.2	20.2	24.0
$FeCO_3$	18.4	15.6	24.9	21.5
$MnCO_3$	4.0	3.4	3.7	3.2
Total	100.0	100.0	100.0	100.0

Distribution, Texture and Mineral Associations

Ankerite is found here and there in various units of the Sokoman Formation, but generally in not more than accessory amounts. It appears to be most abundant in member V, particularly in the so-called "massive-bedded facies" (Fig. 81). Thirty-two carbonate-bearing samples from different parts of member V were checked by X-ray analysis. Only two of these contained siderite in addition to ankerite. The remainder contained only ankerite.

Ankeritic beds occur in various parts of the massive-bedded facies, but are especially common in the lower and upper part of the unit. The

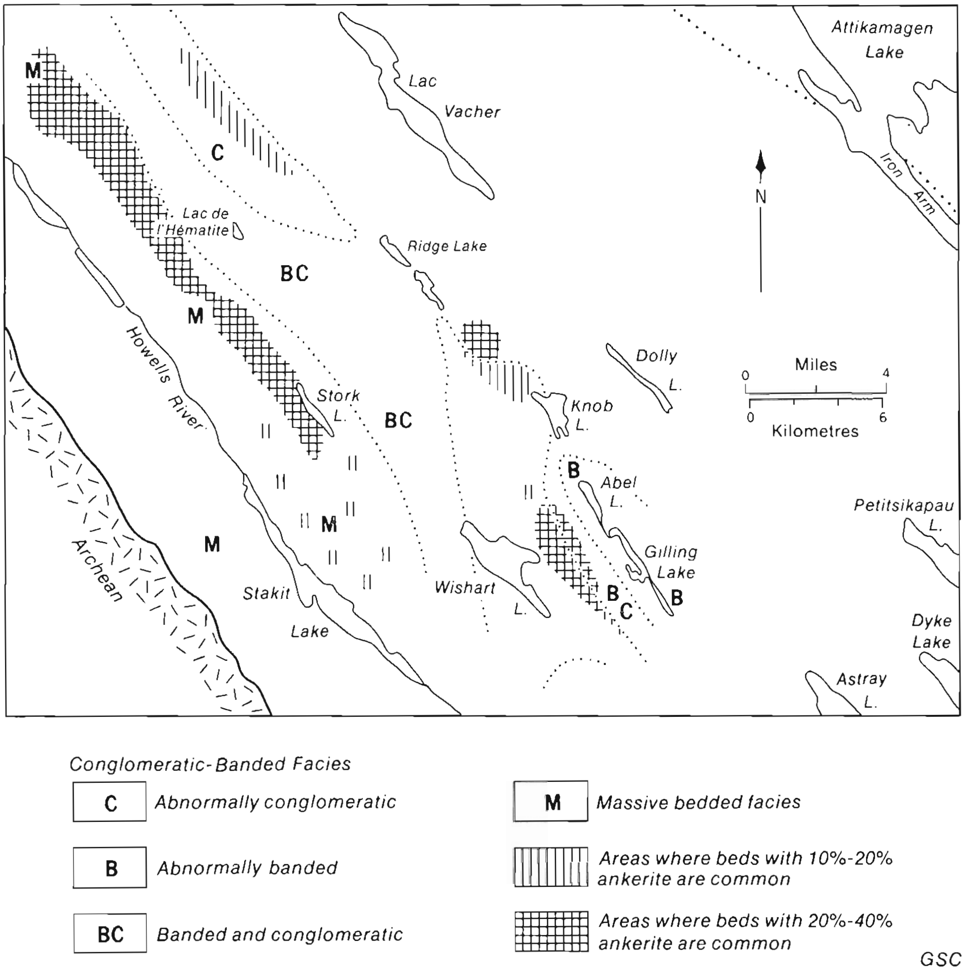


Figure 81. Ankerite-rich areas in member V.

largest amount of ankerite is in the area which extends from just south of Stork Lake northwestward to Irony Mountain, where beds with more than 20 per cent of carbonate are numerous. The carbonate is irregularly distributed. Much of it occurs as haphazardly scattered patches and oval aggregates, and as very irregular beds and lenses. Most of it is granular and, as noted by Perrault (1955), distinctly crystalline. Individual crystals are up to several millimetres in size.

Ankerite-rich beds are also present in a number of places in the conglomeratic-banded facies (Fig. 81). The carbonate occurs as fine disseminations, as irregular to nearly spherical aggregates and as well-defined layers and lenses. In the area east of Wishart Lake (Fig. 81), some of the beds composed chiefly of ankerite are up to 2 inches thick. This area is also one of the few places where large, slabby ankerite pebbles can be seen in some of the conglomeratic beds.

The disseminated ankerite occurs as subhedral to euhedral rhombs, commonly 30 to 300 microns in size. In beds and lenses composed predominantly of ankerite, the carbonate is anhedral and microcrystalline to coarse

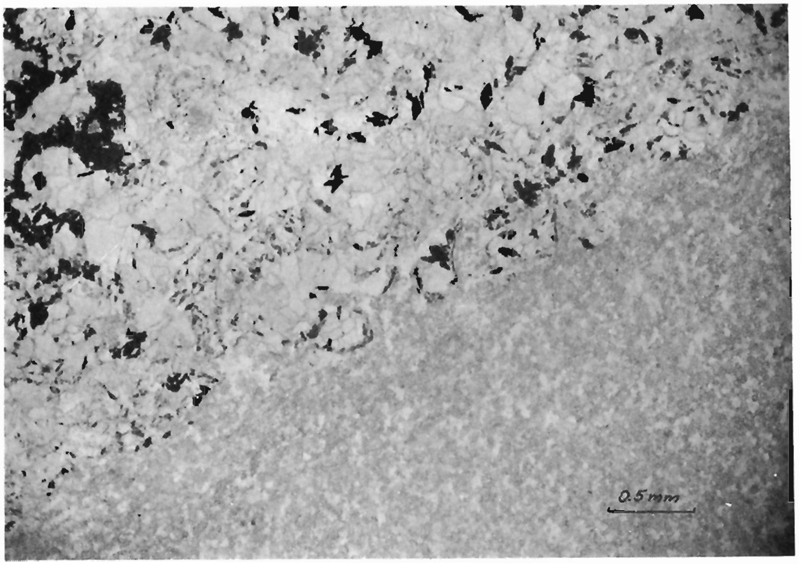


Figure 82. Photomicrograph showing granular, coarse-grained ankerite in contact with nongranular, fine-grained ankerite. The opaques are hematite.

grained. The coarser-grained variety is typically granular (Fig. 82). Fine-grained ankerite, so rare in the massive-bedded facies, is not uncommon in the conglomeratic-banded rocks. It occurs as thin beds and lenses which are in sharp contact with layers of chert, jasper or iron oxides; some are in contact with relatively coarse-grained ankerite (Fig. 82). The fine-grained carbonate, unlike its coarser counterpart, is comparatively uniform in texture. It is generally even grained and nongranular.

The colour of ankerite is white, grey, pale buff, pink or purplish pink but buff and pink colours predominate. On weathering the carbonate assumes various shades of brown.

The ankerite is associated with a great variety of minerals, but chert, hematite and magnetite are the most common. Ankerite is the characteristic carbonate of most hematite-bearing oxide facies of the iron-formation, although siderite is the predominant carbonate of some (member I, transition zone of member IV).

Origin

The texture of most ankerite in the Sokoman Formation is certainly not primary. Likewise, there can be little doubt that at least some recrystallization and redistribution of the carbonate has taken place after the enclosing sediments were deposited. Whether any of the ankerite was originally a primary precipitate is uncertain. However, the distribution and mineral association of ankerite make sense if it is assumed that the ankerite was originally deposited as calcium carbonate.

The precipitation of calcium carbonate is strongly affected by changes in pH. High pH favours the precipitation of calcite, whereas a decrease in

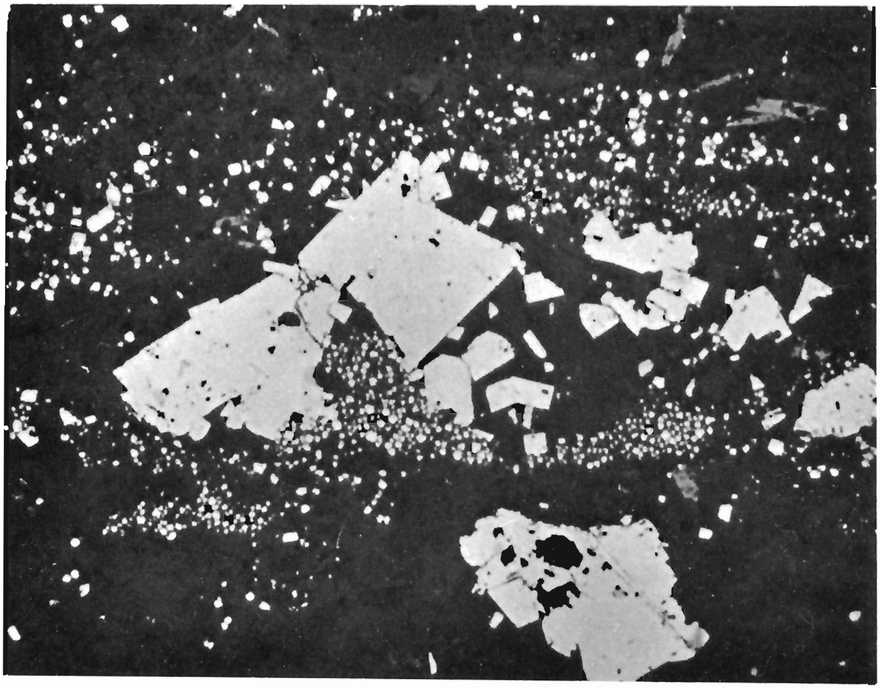


Figure 83. Typical texture of pyrite. Note the two distinct sizes of pyrite crystals.

pH lessens the chances that the mineral will form. The stability of calcite is independent of Eh so that it can form and coexist with a variety of Eh-dependent minerals ranging from pyrite to hematite. At low Eh, the associated minerals could be pyrite, silicates or siderite, and at high Eh, magnetite and hematite.

The most favourable sites for calcium carbonate deposition would be the shallow parts of sedimentary basins where higher temperatures and intense water turbulence would cause the loss of CO_2. Shallower waters are also likely to be well oxygenated, particularly where the wave and current activity is intense, so that iron oxides could precipitate in the company of calcite. It is not surprising, therefore, to find that ankerite (assuming that it was originally calcite) is preferentially associated with iron oxides. It is also reasonable that in member V it is most common in the massive-bedded facies which formed in the shallow, high energy offshore ridge on the western flank of the Knob Lake basin.

PYRITE

Most of the pyrite in the Sokoman Formation is confined to the black, carbonaceous and tuffaceous shales of member I. The pyrite is generally very fine grained (Fig. 83) and difficult to recognize except in polished sections. It appears to be most abundant in the area between Knob Lake and Iron Arm within the lower 30 to 50 feet of member I where beds containing 10 to 20 per cent pyrite are not uncommon. A few thin layers may contain as much as 50 per cent of the mineral (Fig. 20, p. 66).

The shales contain two distinct types of pyrite (Fig. 83): microcrystalline pyrite with most crystals ranging from 3 to 8 microns, and comparatively coarse-grained pyrite with crystals commonly 50 microns to several hundred microns in size.

The microcrystalline pyrite accounts for more than 90 per cent of the pyrite in member I. It occurs as disseminated crystals or as dense aggregates of crystals in the form of streaks, lenses and continuous layers which invariably parallel the bedding of the shales. The coarse-grained pyrite is much more erratically distributed. It is haphazardly disseminated or concentrated into irregular aggregates and crosscutting veinlets (Fig. 20). Some of it appears to fill short crosscutting fractures that may be shrinkage cracks (Fig. 20).

The coarse pyrite is commonly, although by no means always, present in the same beds as the microcrystalline variety. The presence of the large crystals, however, does not affect the distribution or texture of the fine-grained pyrite even where the two types of pyrite are in contact with one another (Fig. 83). In places where the coarse pyrite envelopes the small crystals, it does not assimilate them but retains them as distinct inclusions (Fig. 83). The texture and distribution of the surrounding pyrite is also unchanged (Fig. 83).

The two pyrites appear to have formed at different times. The microcrystalline type is probably primary-diagenetic pyrite which precipitated from the sea water during or shortly after deposition of the shale. The coarse-grained pyrite appears to have crystallized at some later stage in the history of the rock. However, there is no evidence to suggest that it was formed by recrystallization of the pre-existing fine-grained sulphide. The coarse pyrite was probably deposited by some of the last interstitial solutions expelled during consolidation of the shale.

ORIGIN OF THE SOKOMAN FORMATION

CHEMICAL CONTROLS OF THE DEPOSITIONAL ENVIRONMENT

The sequence of mineralogical changes which typifies the iron-formations of the Lake Superior region is sulphide-carbonate-silicate-oxide, the last three being the dominant facies of the region (James, 1954, 1966). The same sequence of changes is also characteristic of the Sokoman Formation in the central Labrador Trough, although the sulphide facies is poorly represented.

Member I is the only unit of the Sokoman Formation which exhibits the complete array of laterally gradational facies ranging from black, carbonaceous, pyritic shales deposited in the deepest part of the Knob Lake basin, to predominantly red, chert-magnetite-hematite jaspilites formed in the shallowest area of the basin (Fig. 27). Member II, although its mineralogy is not well known and apparently more complicated, shows on a regional scale a lateral change from carbonate facies, which correspond to the deeper parts of the Knob Lake basin, to iron-formation composed mainly of chert and iron oxides located in the shallower areas of the depression. The mineralogical differences in member III are not as extreme as in member I, but on the whole just as systematic, ranging from carbonate-silicate to silicate-oxide (Fig. 29).

The change from carbonate to silicate to oxide is even more apparent in vertical sequences in the lower part of the iron-formation (members I to IV). The most obvious and in places remarkably systematic variations occur within member I in the central Stakit Lake area. The mineralogy of the iron-formation ranges upward from chert-carbonate in the lower part of the member to chert-carbonate-silicate, chert-silicate-carbonate, chert-silicate-oxide and finally to chert-iron oxide at the very top of the member. Similar changes can be seen in members II through IV in the ore zone (see Fig. 75). Lateral and vertical mineralogical changes are not as simple nor as easy to follow in the upper part of the Sokoman Formation, but in general follow the same sequence: carbonate-silicate-oxide.

The outstanding thing about the facies relationships is that the changes in mineralogy are closely related to the changes in sedimentary textures of the rocks, both of which reflect the change from deep to shallow environment of deposition. The lateral changes depended largely on the topography of the basin and can be related to distances of the depositional sites from shore; vertical variations depended mainly on the changes in the overall depth of water in the sedimentary basin, due to the variations in sea level (i.e., transgression or regression of the sea).

The correlation between sedimentary textures and mineralogy in the Sokoman Formation is by no means ideal. Member VII, where little correlation between textures and mineralogy is apparent, is the prime example. In other units of the formation, however, inconsistencies in correlation are minor. On the whole, particularly when one considers large stratigraphic units on regional scale, the correlation between textures and mineralogy is excellent. Thus, most oxide facies of the formation are typically granular, irregularly bedded, comparatively thick bedded, and in places conglomeratic, crossbedded and oolitic. Most carbonate facies, on the other hand, contain none of these shallow water features except locally some small and poorly

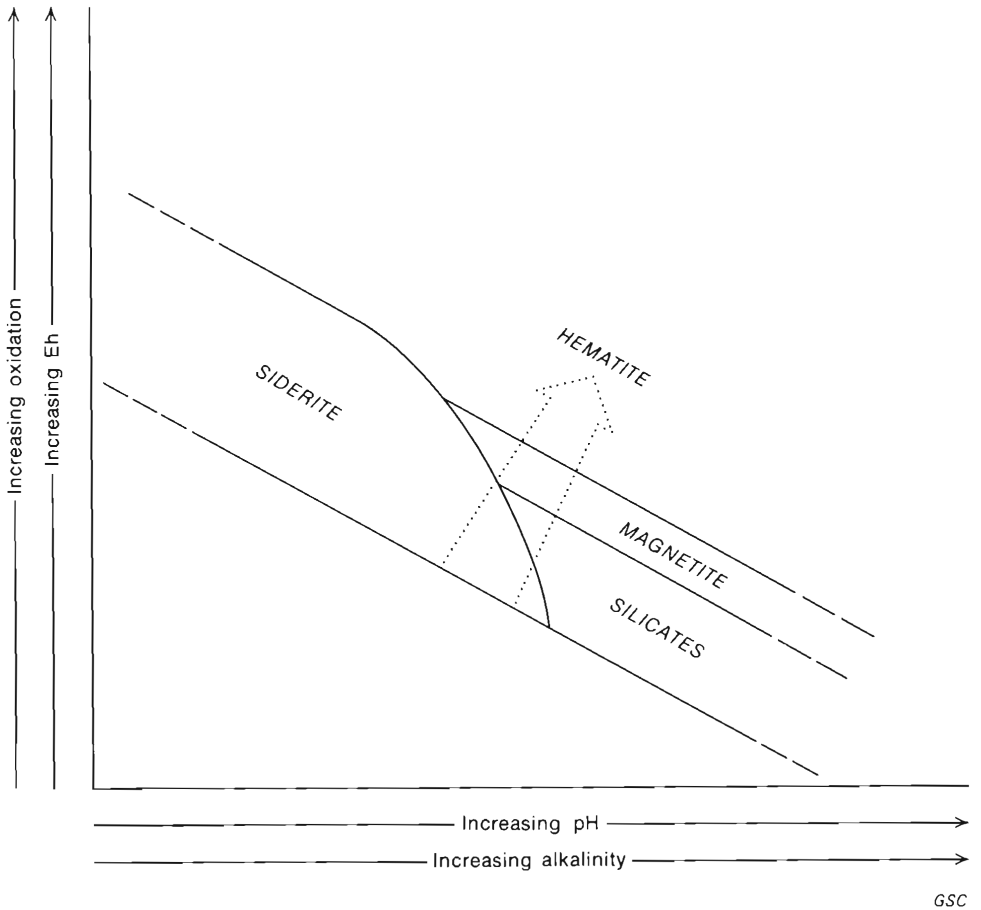


Figure 84. Stability fields of siderite, iron silicates to Eh and pH. The arrow indicates the prevalent sequence of mineralogical changes in the Sokoman and other Superior-type iron-formations.

formed granules. The silicate facies, which occupy an intermediate position in the sequence of changes from carbonate to oxide, are also texturally intermediate between the two.

The most common sequence of facies changes in the Sokoman Formation and in other Superior-type iron-formations is carbonate-silicate-oxide. The specific minerals which best fit this sequence are siderite, the most common carbonate; minnesotaite, the most common silicate; and magnetite and hematite, the only primary oxides. The order of succession is siderite-minnesotaite-magnetite-hematite. These minerals in most cases adhere closely to their position in the sequence so that mixtures of non-sequential minerals, such as of hematite and siderite or of minnesotaite and hematite, are rare.

Figure 84 shows the relation of various minerals to the oxidation-reduction potential (Eh) and to the hydrogen ion concentration (pH), the two main factors which determine the type of minerals formed. The sequence of facies observed in the Superior-type iron-formations could have resulted through the changes in Eh, pH, or both. The most likely path followed in the change from deep to shallow environment during deposition of the iron-formations is shown in Figure 84. Although Eh and pH are independent variables, in natural sedimentary environments the change in Eh is commonly accompanied by a change in pH. Ideally, as the Eh increases in the change from deep and stagnant to shallow and well-oxygenated areas of deposition, the pH also increases due to the escape of carbon dioxide into the atmosphere as the result of the increasingly warmer and more turbulent waters in the shallower areas.

Although Eh is perhaps the most obvious control of the mineralogical changes since it reflects the oxidation state of the iron in the resulting minerals, pH is also important. As Figure 84 shows, the silicate field can be bypassed entirely if the change from reducing to oxidizing condition takes place at sufficiently low pH. This has only rarely occurred during deposition of the Superior-type iron-formations as evidenced by the fact that sedimentary units composed largely or wholly of chert, siderite and iron oxides with few or no silicates are scarce in the Sokoman Formation as well as in the iron-formations of the Lake Superior region. Sedimentary units which contain both siderite and hematite are extremely rare. Only two examples can be cited from the Knob Lake area: the thinly bedded chert-siderite-magnetite-hematite interbeds in the oxide facies of member I is one, and the texturally and mineralogically similar rocks in the transition zone of member IV is another. In each case, the presence of siderite and iron oxides in the same rocks and the absence of silicates can be readily explained by a change toward a more acid environment of deposition.

PHYSICAL CONTROLS OF THE DEPOSITIONAL ENVIRONMENT

One of the outstanding features of the Superior-type iron-formations is that they are not simple chemical precipitates but chemical sediments which have been affected by waves and currents during sedimentation. The Superior-type iron-formations are in part intraformationally reworked sediments, analogous in this respect, as noted by LaBerge (1967) and Dimroth (1968), to intraclastic limestones (calcarenites).

The granules, oolites, irregular bedding and the less common cross-bedding, which characterize the Superior-type iron-formations (Gross, 1965), are all indicative of wave and current activity penecontemporaneous with sedimentation. The intraformational conglomerates which are found only in the iron-formations of the Superior-type and which are particularly common in the Sokoman Formation (see Figs. 45 to 47) are the most obvious proof of intraformational reworking of the chemical sediments. Many of the local differences in the thickness of beds and larger stratigraphic units (members V and VII in particular) as well as many of the mineralogically hybrid rocks, such as those in the transition zone between members III and IV or those in the mixed facies of member VII, are also best explained by intraformational reworking of sediment.

Intraformational erosion, in most places, does not seem to have been very deep or extensive. In most units of the Sokoman Formation, the amount of sediment removed probably did not exceed more than a few inches to several feet, with the possible maximum of 40 to 60 feet in such variable units as

V and VII. The distance of transport of the larger particles also does not appear to have been great. In examining the granular and conglomeratic units of the iron-formation one gets the impression that much of the movement of the reworked sediment consisted of shifting it from place to place within a small area rather than of moving the sediment large distances away from its original place of deposition. Most pebbles and fragments appear to have been locally derived (see Figs. 26, 46, 47, 53 and 54) – certainly from the same stratigraphic units (members) – and transported no more than a few feet to several hundred feet. Small granules and still finer particles were undoubtedly transported farther. The trend surface analyses of the distribution of total Fe in members III and V (Figs. 36, 51) suggest that some of the finest clay-size material was transported at least several miles. The predominant movement of the fine material was evidently eastward – downslope from the shallow high energy areas, north of the present Stakit Lake, to the somewhat deeper and less turbulent parts of the basin.

The physical energy of the depositional environment played an important role not only in determining the textures of the iron-formations, but also in determining to some degree the mineralogy of the sediments. First of all, the intensity of turbulence, as explained previously, would affect the Eh and pH of the environment by determining in part the oxygen and carbon dioxide content, and so indirectly exert control over the type of minerals that would form. The result is a good correlation between sedimentary textures and mineralogy of the iron-formation. Secondly, any local erosion and redistribution of the sediment would modify the physical configuration of the depositional environment and consequently affect the pattern of circulation, the intensity of wave and current action and the chemical gradients within the area of deposition and, therefore, the mineralogical character of the subsequently deposited materials. The erratic facies changes in member VII as well as some of the local mineralogical variability in other units of the iron-formation can be attributed to the irregularities in the depositional interface created by local erosion and redistribution of the accumulated sediment. Thirdly, the reworking of the sediment and the transport of the eroded material from one environment of deposition to another would result, if the distance of transport or the depth of erosion and the degree of mixing is sufficiently great, in mineralogically hybrid rocks. The rocks which contain granules, pebbles and fragments of different composition within the same beds undoubtedly formed in this manner.

The effects of wave and current activity on the chemical precipitates of iron and silica during deposition become most obvious when the wave- and current-worked units of the Superior-type iron-formations are compared with the thick and extensive iron-formations of South Africa and Western Australia which show little or no evidence of turbulence contemporaneous with sedimentation. The Brockman iron-formation in the Hamersley Ranges of Western Australia (Trendall, 1965, 1968), for example, is 2,200 feet thick and occupies an area of approximately 50,000 to 30,000 square miles. It contains no granules, oolites, conglomerates or other features that might be attributed to high physical energy of the depositional environment. The formation is thinly bedded to laminated throughout and characterized by "spectacular lateral continuity on a minute scale" (Trendall, 1965, p. 64). Thin beds within the iron-formation can be traced for miles with little apparent change. The lateral continuity and uniformity of the iron-formation appear to be so ideal that it has led Trendall (1965) to conclude that the sediments were probably deposited in water which was of even depth and "amazingly still" (Trendall, 1965, p. 64), and Miles (1963, p. 119) to observe that:

> lithological uniformity is a normal and characteristic feature of Precambrian jaspilites and banded iron-formation, suggestive of long continued and uniform conditions of deposition.

No such statements can be made about the more variable Superior-type iron-formations, certainly not about the Sokoman.

EXTERNAL CONTROLS OF DEPOSITIONAL ENVIRONMENT

The environment in which the Sokoman Formation in the Knob Lake area was deposited was also affected by the following:

(1) Tectonic modifications of the Labrador Trough which led to the development of the Knob Lake basin.
(2) Fluctuations of sea level which in part controlled the oxygen availability and the levels of physical energy during sedimentation.
(3) Volcanic activity, uplift and erosion in the Nimish area which supplied clastic material into the basin of deposition.

Tectonic Activity

In current geosynclinal theory the miogeosyncline is not a single linear belt of subsidence but a series of basins which collectively define the zone of transition between the craton and the eugeosyncline (Krumbein and Sloss, 1963, p. 415). The Knob Lake basin, revealed by the pattern of mineralogical and textural changes in the Sokoman Formation, was probably one of such shallow interconnected basins that existed in the miogeosynclinal part of the Labrador Trough during Sokoman sedimentation.

The first suggestion of a basin in the Knob Lake area comes from the facies relationships of the Denault and Attikamagen Formations. The dolomite suggests the shallower sites of deposition, and the shale east of Knob Lake the deeper areas of deposition. This basin was apparently shallow and largely filled with sediment by the end of Wishart time.

The basin in which the Sokoman Formation was deposited formed after the deposition of the Wishart Quartzite. As noted in the discussion of the quartzite, the contact between the Wishart and Sokoman Formations corresponds to an abrupt change from coarse clastic to predominantly chemical sedimentation. Comparing the lower part of the Sokoman with the upper part of the Wishart, it is also evident that a significant change in the configuration of the sea floor had occurred by the time that deposition of member I was completed. The upper 10 to 30 feet of the Wishart in the Knob Lake area consists almost entirely of medium- to coarse-grained, irregularly bedded quartzites which were clearly deposited in a shallow, high-energy environment. The thin but widespread conglomerates at the very top of the quartzite west of Knob Lake indicate that this area was somewhat more elevated than the areas in the east (Fig. 85, profile A). Otherwise, there appears to be no

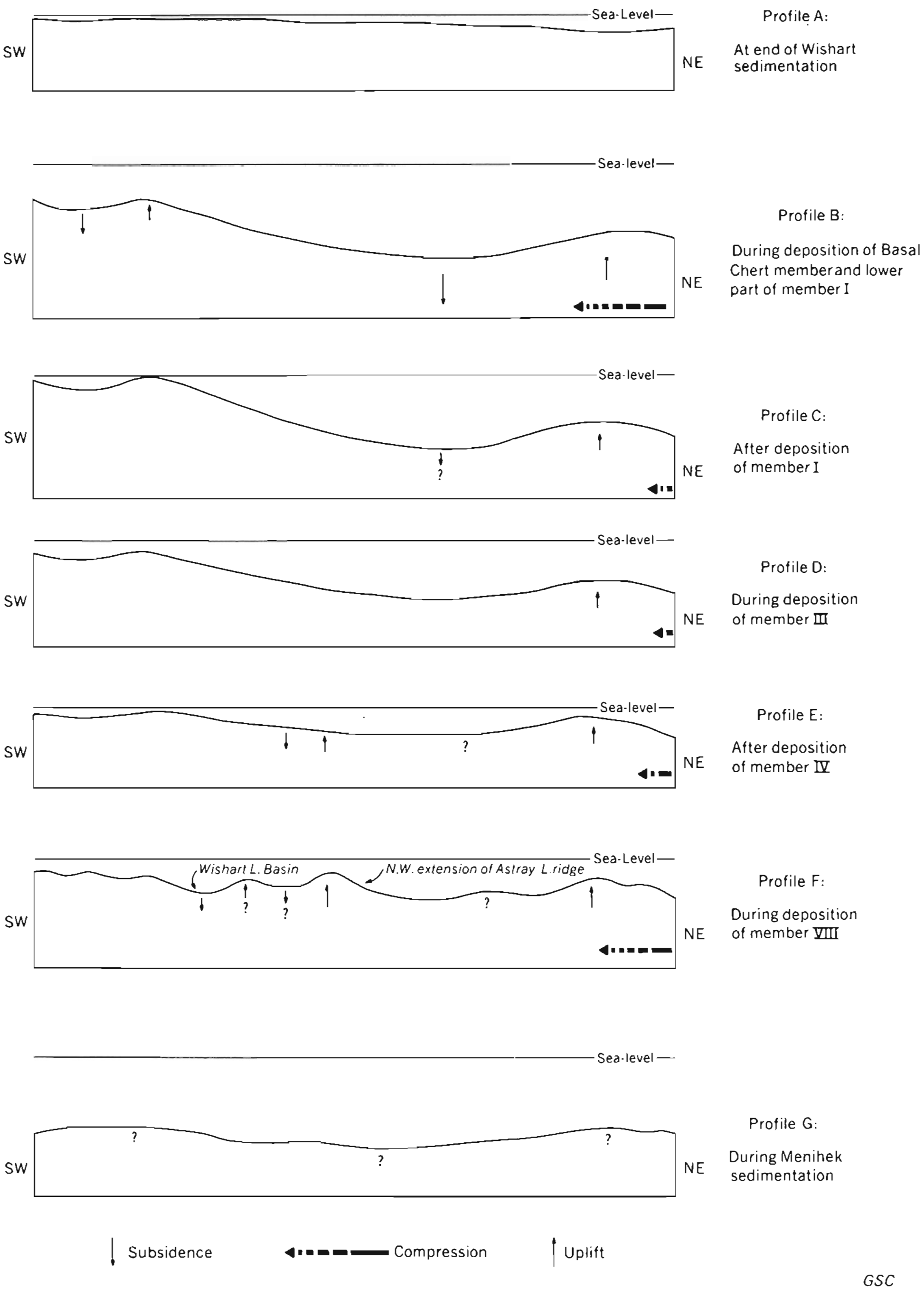


Figure 85. Diagrammatic profiles of the Knob Lake Basin at different stages of sedimentation.

significant difference in the environment in which the upper part of the formation was deposited. There is no obvious indication that the Knob Lake basin existed at this time.

The textures, mineralogy and facies changes of the iron-formation comprising member I contrast sharply with the upper part of the Wishart Quartzite. Not only does the lower part of the iron-formation indicate deposition in deeper and drastically less turbulent water, but its well-differentiated facies suggest the existence of a basin with a greater difference in topographic relief over the same area where the comparatively uniform quartzite was deposited previously. The change in the topography of the depositional area can be best explained by structural readjustments within the Labrador Trough, or more specifically, by gentle buckling of the miogeosynclinal part of the trough by compressive tectonic forces directed from the northeast (Fig. 85). The cause of the structural readjustments may be the early manifestation of the tectonic forces which eventually culminated in the Hudsonian orogeny. This is suggested by the similar northwesterly orientation of the Knob Lake basin, its subsidiary depressions and ridges, and of the faults and folds attributed to the Hudsonian orogeny. The outlines of the Knob Lake basin show up most clearly from the facies relationships in the lower part of the Sokoman Formation. Mineralogical and textural changes in members I, III and IV outline a northwesterly elongated basin with a prominent northwesterly-trending offshore ridge, and a similarly oriented embayment west of it (Figs. 27, 29, 39). The similarity of results obtained by independent analyses of the three members indicates that after the initial deformation which formed the Knob Lake basin the area did not undergo any major structural modifications during deposition. The gradual and generally systematic lateral changes in thickness, texture and mineralogy in the lower part of the Sokoman are consistent with a tectonically stable environment. The only change that can be attributed to structural readjustment is the uplift of the eastern flank of the basin to the extent that during the last stages in the sedimentary history of member IV, the eastern flank was at approximately the same depth below sea level as the offshore ridge in the west (Fig. 85, profile E).

During deposition of member V, the pattern of sedimentation begins to change. Some of the local and erratic changes in thickness and mineralogy of members V through IX can be attributed to local intraformational erosion and redistribution of the accumulating sediments and to periodic influx of clastic material from the Nimish area. Changes resulting from structural modification of the basin appear to have been also important. The small depression in the area of Wishart Lake appears for the first time during deposition of member VI. The sudden appearance of this depression and its persistence through deposition of members VI to VIII (and IX?) despite the fact that it was continually filled with sediment, suggests that this small subsidiary basin was of tectonic origin. The Astray Lake Ridge (the area just north of present Astray Lake) where very little sediment accumulated during deposition of members V through VII was probably a northward extension of the Nimish landmass. There is no indication that this ridge (Fig. 85), profile F) existed during the early sedimentary history of the iron-formation. Its appearance as an elevated area during deposition of members V and VI is difficult to explain without structural readjustment.

The increasing depth of water of the transgressing seas during the late stages of Sokoman sedimentation effectively conceals the topography of the Knob Lake basin. It may be that the basin was largely filled with sediment by this time, so that the thick cover of carbonaceous shales of the Menihek Formation was deposited over not more than a gently undulating surface (Fig. 85, profile G).

Fluctuations of Sea Level

When only the major changes from bottom to top of the iron-formation are considered, one can see a very definite pattern. The composition of the formation changes from silicate-carbonate to oxide and back to silicate-carbonate (Table 14). This generalized sequence fits most of the iron-formation in the Labrador Trough from Wabush Lake in the south (Gastil and Knowles, 1960) to north of Payne Bay (Gross, 1968). The changes are obviously regional in scope and probably reflect the changes in environment due to one complete regressive-transgressive cycle of the sea (Table 14).

Smaller repetitive changes in composition of the iron-formation can also be recognized (Schwellnus, 1957). The changes highlighted by four prominent jasper horizons (the oxide facies of member I and members IV, VI and VIII), probably represent small, periodic changes in sea level.

The causes of the major and minor fluctuations of sea level are not known. However, regardless of how they are interpreted, the changes played an important role in determining the mineralogical and textural character of the iron-formation by controlling in part the oxygen availability and the levels of physical energy (wave and current activity) during sedimentation.

Volcanism, Uplift and Erosion in the Nimish Area

The sequence of events in the Nimish area agrees well with the history that can be reconstructed from the stratigraphic record left by the clastic facies of the iron-formation in the Knob Lake area. Volcanic activity began early in the history of the Nimish area (Fig. 9), but was most intense during deposition of the Sokoman Formation. The first major outburst of volcanism combined with probable uplift and erosion began with deposition of

Table 14. Fluctuation of Sea Level

Subdivisions of Sokoman I. F.	Dominant facies	Environ-ment	Order of repetition	Changes in sea level
UPPER Members: VII, VIII, IX, X	Silicate-carbonate	Deep reducing	A	Deep ↖ Transgression
MIDDLE Members: IV, V, VI	Oxide	Shallow oxidizing	B	Shallow
LOWER Members: I, II, III	Silicate-carbonate	Deep reducing	A	Regression ↗ Deep

member I but subsided during deposition of members II, III and IV. Volcanic activity, uplift and erosion recurred during deposition of members V to VIII and reached a climax during deposition of member IX.

The general activity in the Nimish area appears to be closely related to the structural modifications which occurred in the Knob Lake area. The first intense volcanic activity coincided closely with the initial development of the Knob Lake basin and was renewed during later structural modifications of the depression. It seems likely that the tectonic compressive forces which affected the Knob Lake area were also instrumental in reopening old and forming new fissures which served as conduits for the upwelling magma. The compressive tectonic forces directed from outside the Nimish area may have also facilitated and possibly even initiated the ascent of the magma by compressing the magma reservoir.

The volcanism, uplift and erosion in the Nimish area had one obvious affect on Sokoman sedimentation. It contributed clastic material which contaminated the accumulating chemical precipitates of iron and silica for tens of miles outward from the Nimish area. The introduction of clastic material into the Knob Lake basin was confined largely to a northwesterly-trending zone between the present Knob Lake and Iron Arm. Most of the iron-formation (members II to VIII and member X) east to north of Iron Arm and west to northwest of Knob Lake escaped contamination. In fact, if only these parts of the iron-formation were exposed, it would be impossible to tell that any volcanism or erosion was taking place only 20 to 25 miles away. Only during the most intense periods of volcanism and erosion, concurrent with deposition of members I and IX in the Knob Lake area, does the contamination extend over larger areas. But even then the contamination in areas east of Iron Arm and west of Knob Lake is accomplished mainly by very fine, clay-size material which offers no direct evidence of having been derived from the Nimish source.

The coincidence of volcanism, uplift and erosion in the Nimish area with deposition of the Sokoman Formation invites another question. How important were the events in the Nimish area in contributing iron and silica to the iron-formation? Some iron and silica was undoubtedly introduced in the form of clastic particles derived by erosion of the iron-formation in the Nimish area. Some iron, as suggested by Sauve (1953), may also have been derived by weathering of the volcanic rocks in that area. The amount of iron and silica contributed by either process is unknown, but it was probably small and confined mainly to the south-central part of the Knob Lake basin.

The possibility of deriving iron and silica from volcanic exhalations also exists, but so far has not been substantiated. Neither Sauve (1953), Perrault (1955), nor the present writer has found any evidence which would indicate that volcanic emanations contributed iron or silica directly to the areas in which the Sokoman Formation was deposited. Volcanism in the Nimish area appears to have acted solely as a dilutant and contaminant of the chemical precipitates of iron and silica rather than a contributor of these elements.

REFERENCES

Allen, G.C.
1967: Riebeckite occurrence in southern Virginia; Va. Miner., v. 13, p. 10.

Baragar, W.R.A.
1960: Petrology of basaltic rocks in part of the Labrador Trough; Geol. Soc. Amer. Bull., v. 71, p. 1583-1644.

1963: Wakuach Lake map-area; Geol. Surv. Can., Paper 62-38, 4 p.

1967: Wakuach Lake map-area; Geol. Surv. Can., Mem. 344, 174 p.

Barghoorn, E.S., and Tyler, A.S.
1965: Micro-organisms from the Gunflint Chert; Science, v. 147, p. 563-577.

Beales, F.W.
1965: Diagenesis in pelleted limestones; in Pray, L.C., and Murray, R.C., eds., Dolomitization and limestone diagenesis; Soc. Econ. Paleontol. Mineral., Spec. Publ. no. 13, p. 49-70.

Beyseyev, O.B.
1966: K mineralogii i genezisu rodusita iz Dzhezkazganskoy i Yuzhno-Minusinskoy vpadin (On mineralogy and genesis of rhodusite from Dzezkazgan and south Minusinsk depressions); Doklady Dkademii Nauk SSSR, v. 167, p. 416-419.

Blake, R.L.
1964: Some iron phyllosilicates of the Cuyuna and Mesabi Districts in Minnesota; U.S. Dept. Interior, Bur. Mines, Rept. Invest. 6394, 33 p.

Cilliers, J.J. Le R., Freeman, A.G., Hodgson, A., and Taylor, H.F.W.
1961: Crocidolite from the Koegas-Westerberg area, South Africa; Econ. Geol., v. 56, p. 1421-1437.

Clarke, F.W.
1924: The data of geochemistry, 5th ed.; U.S. Geol. Surv., Bull. 770, 841 p.

Cochrane, G.W., and Edwards, A.B.
1960: The Roper River sedimentary oolitic formations; Austral., Commonwealth Sci. Ind. Res. Organ., Mineragraph. Invest. Tech. Paper 1, 28 p.

Davies, K.G.
1963: Northern Quebec and Labrador journals and correspondence 1819-35; London, The Hudson's Bay Record Society, 415 p.

Deer, W.A., Howie, R.A., and Zussman, J.
1962: Rock-forming minerals, v. 2, Sheet silicates; London, Longmans, Green, 270 p.

Dimroth, E.
1968: Sedimentary textures, diagenesis and sedimentary environment of certain Precambrian ironstones; Neues Jahrb. Geologie u. Palaontologie Abh., v. 130, p. 247-274.

Donaldson, J.A.
1966: Marion Lake map-area, New Quebec-Newfoundland; Geol. Surv. Can., Mem. 338, 85 p.

Drysdall, A.R., and Newton, A.R.
1960: Blue asbestos from Lusaka, Northern Rhodesia, and its bearing on the genesis and classification of this type of asbestos; Am. Mineral., v. 45, p. 53.

Dufresne, C.
1952: A study of the Kaniapiskau System in the Burnt Creek-Goodwood area, New Quebec and Labrador, Newfoundland; unpubl. Ph.D. thesis, McGill Univ., Montreal.

DuToit, A.L.
1945: The origin of the amphibole asbestos deposits of South Africa; Geol. Soc. S. Afr. Trans., v. 48, p. 161-206.

Flaschen, S.S., and Osborn, E.F.
1957: Studies of the system iron oxide-silica-water at low oxygen partial pressures; Econ. Geol., v. 52, p. 923-943.

Folk, R.L., and Weaver, C.E.
1952: A study of the texture and composition of chert; Am. J. Sci., v. 250, p. 498-510.

Frarey, M.J.
1952: Wilbob Lake, Quebec and Newfoundland; Geol. Surv. Can., Paper 55-42, 8 p.

1961: Menihek Lakes; Geol. Surv. Can., Map 1087A.

Frarey, M.J., and Duffell, S.
1964: Revised stratigraphic nomenclature for the central part of the Labrador Trough; Geol. Surv. Can., Paper 64-25, 13 p.

French, B.M.
1968: Progressive contact metamorphism of the Biwabik iron-formation; Mesabi Range, Minnesota; Minn. Geol. Surv. Bull. 45, 103 p.

Garrels, R. M.
1959: Reactions at low temperatures and pressures; in Abelson, P.H., ed., Researches in geochemistry; New York, John Wiley and Sons, p. 25-37.

Garrels, R.M., and Christ, C. L.
1965: Solutions, minerals and equilibria; New York, Harper and Row, 450 p.

Gastil, G., and Knowles, D. M.
1960: Geology of the Wabush Lake area, Southwestern Labrador and Eastern Quebec; Geol. Soc. Amer. Bull., v. 71, p. 1243-1254.

Gill, J. E.
1927: Origin of the Gunflint iron-bearing formation; Econ. Geol., v. 27, p. 687-728.

Goodwin, A. M.
1956: Facies relations in the Gunflint Iron-Formation; Econ. Geol., v. 51, p. 565-595.

Gross, G. A.
1951: A comparative study of three slate formations in the Ferriman Series in the Labrador Trough; unpubl. M.Sc. thesis, Queen's Univ., Kingston, Ont.

1960: Iron-formations and the Labrador Geosyncline; Geol. Surv. Can., Paper 60-30, 7 p.

1961: Metamorphism of iron-formations and its bearing on their beneficiation; Can. Inst. Mining Met. Trans., v. 54, p. 24-31.

1965: Geology of iron deposits in Canada, v. I, General geology and evaluation of iron deposits; Geol. Surv. Can., Econ. Geol. Rept. 22, 174 p.

1968: Geology of iron deposits in Canada, v. III, Iron ranges of the Labrador Geosyncline; Geol. Surv. Can., Econ. Geol. Rept. 22, 179 p.

Gruner, J. W.
1936: The structure and chemical composition of greenalite; Am. Mineral., v. 21, p. 449-455.

1946: The mineralogy and geology of the taconites and iron ores of the Mesabi range, Minnesota; Iron Range Resources and Rehabilitations; St. Paul, Minn., 127 p.

Gunderson, J. N., and Schwartz, G. M.
1962: The geology of the metamorphosed Biwabik iron formation, Eastern Mesabi district, Minnesota; Minn. Geol. Surv. Bull. 43, 139 p.

Gustafson, J. K., and Moss, A. E.
1953: The role of geologists in the development of the Labrador-Quebec iron ore districts; Am. Inst. Mining Eng. Trans., v. 196, p. 593-602.

Hall, A. L.
1930: Asbestos in the Union of South Africa; S. Afr. Geol. Surv. Mem. 12, 152 p.

Harrison, J. M.
1952: The Quebec-Labrador iron belt, Quebec and Newfoundland; Geol. Surv. Can., Paper 52-20, 21 p.

Hotchkiss, W. O.
1919: Geology of the Gogebic range and its relation to recent mining developments; Eng. Mining J., v. 108, no. 11, p. 443-452.

Howell, J. E.
1954: Silicification in the Knob Lake Group of the Labrador iron belt; unpubl. Ph.D. thesis, Univ. of Wisconsin.

Huber, N. K.
1959: Some aspects of the origin of the Ironwood iron-formation of Michigan and Wisconsin; Econ. Geol., v. 54, p. 82-118.

Illing, L. V.
1954: Bahaman calcareous sands; Am. Assoc. Petrol. Geol. Bull., v. 38, p. 1-95.

Iron Ore Company of Canada Staff
1955: The operations of the Iron Ore Company of Canada Limited; Can. Mining J., v. 76, p. 38-57.

James, H. L.
1954: Sedimentary facies of iron formation; Econ. Geol., v. 49, p. 235-293.

1955: Zones of regional metamorphism in the Precambrian of northern Michigan; Geol. Soc. Amer. Bull., v. 66, p. 1455-1488.

1966: Chemistry of the iron-rich sedimentary rocks; U.S. Geol. Surv. Prof. Paper 440-W, 60 p.

Jolliffe, F.
1935: A study of greenalite; Am. Mineral., v. 20, p. 405-425.

King, D.
1961: The occurrence and comparative mineralogy of South Australian magnesian crocidolites; Royal Soc. S. Austral., Trans., v. 84, p. 119-128.

Kirkland, R. W.
1950: A study of the Kaniapiskau system northwest of the Attikamagen Lake; unpubl. Ph.D. thesis, McGill Univ., Montreal.

Krumbein, W.C., and Garrels, R.M.
1952: Origin and classification of chemical sediments in terms of pH and oxidation-reduction potentials; J. Geol., v. 60, p. 1-33.

Krumbein, W.C., and Sloss, L.L.
1963: Stratigraphy and sedimentation (2nd ed.); San Francisco, Freeman and Company, 660 p.

LaBerge, G.L.
1964: Development of magnetite in iron-formations of the Lake Superior region; Econ. Geol., v. 59, p. 1313-1342.

1966a: Altered pyroclastic rocks in iron-formation in the Hamersley Range, Western Australia; Econ. Geol., v. 61, p. 147-161.

1966b: Altered pyroclastic rocks in South African iron-formation; Econ. Geol., v. 61, p. 572-581.

1967: Evidence on the physical environment of iron-formation deposition; Paper presented at the 13th Annual Institute on Lake Superior Geology Meeting, East Lansing, Michigan.

Leith, C.K.
1903: The Mesabi iron-bearing district of Minnesota; U.S. Geol. Surv. Mon. 43, 316 p.

Lepp, H., and Goldich, S.S.
1964: Origin of Precambrian iron-formations; Econ. Geol., v. 59, p. 1025-1060.

Lowden, J.A. (comp.)
1960: Age determinations by the Geological Survey of Canada, Report I - isotopic ages; Geol. Surv. Can., Paper 60-17, 51 p.

1961: Age determinations by the Geological Survey of Canada, Report 2 - isotopic ages; Geol. Surv. Can., Paper 61-17, 127 p.

Low, A.P.
1896: Report on explorations in the Labrador peninsula along Eastmain, Koksoak, Hamilton, Manikuagan and portions of other rivers; Geol. Surv. Can., Ann. Rept., 1895, Rept. L, 386 p.

Mengel, J.T.
1965: Precambrian taconite iron-formation: a special type of sandstone (abst.); Geol. Soc. Amer. Program, 1965 Annual Meetings, p. 106-107.

Miles, K.R.
1942: The blue asbestos-bearing banded iron-formations of the Hamersley ranges, Western Australia; West. Austral. Geol. Surv., Bull. 100, p. 5-37.

Miles, K. R.
1963: Discussion of paper by Catley, D. E., Some aspects of the genesis of the Iron Duke iron ore body and associated rocks; Australasian Inst. Mining Met. Proc., no. 208, p. 119.

Milton, C., and Eugster, H.P.
1959: Mineral assemblages of the Green River formation in Abelson, P.H., ed., Researches in geochemistry; New York, John Wiley and Sons, p. 118-150.

Miyashiro, A.
1957: Chemistry, optics and genesis of the alkali-amphiboles; J. Fac. Sci. Univ. Tokyo, sec. II, v. 11, p. 57.

Moore, E.S., and Maynard, J.E.
1929: Solution, transportation, and precipitation of iron and silica; Econ. Geol., v. 24, p. 272-303, 365-402, 506-527.

Nanz, R.H.
1953: Chemical composition of pre-Cambrian slates with notes on the geochemical evolution of lutites; J. Geol., v. 61, p. 182-242.

Neal, H. E.
1949: The geology of the Hook Lake area, New Quebec, with special reference to the iron-formation; unpubl. M.Sc. thesis, Univ. of Toronto.

Newell, N.D., Purdy, E.G., and Imbrie, J.
1960: Bahamian oolitic sand; J. Geol., v. 68, p. 481-497.

Peacock, M. A.
1928: The nature and origin of the amphibole asbestos of South Africa; Am. Mineral., v. 13, p. 241-285.

Perrault, G.
1955: Geology of the western margin of the Labrador Trough; unpubl. Ph.D. thesis, Univ. of Toronto.

Perrault, G., and Hebert, P.
1968: Minnesotaite from the Bluebell Mine, Riondel, B.C. (abst.); Can. Mineral., v. 9, p. 579.

Pettijohn, F.J.
1957: Sedimentary rocks (2nd ed.); New York, Harper and Row, 717 p.

Polovinkina, Yu. Ir.,
1953: Cummingtonit i schelotchnye amphiboly Krivogo Roga (Cummingtonite and alkali amphiboles from Krivoy Rog); Mineralogichesky Sbornik, no. 7, p. 167-186.

Ruckmick, J.C.
1963: The iron ores of Cerro Bolivar, Venezuela; Econ. Geol., v. 58, p. 218-236.

Rusnak, G.A.
1960: Some observations of recent oolites; J. Sediment. Petrology, v. 30, p. 471-480.

Sauve, P.
1953: Clastic sedimentation during a period of volcanic activity, Astray Lake, Labrador; unpubl. M.Sc. thesis, Queen's Univ., Kingston, Ont.

Schmidt, R.G.
1963: Geology and ore deposits of the Cuyuna North Range; Minnesota; U.S. Geol. Surv. Prof. Paper 407, 96 p.

Schultz, R.W.
1966: Lower Carboniferous cherty ironstones at Tynagh, Ireland; Econ. Geol., v. 61, p. 311-342.

Schwellnus, J.E.G.
1957: Ore controls in deposits of the Knob Lake area, Labrador Trough; unpubl. Ph.D. thesis, Queen's Univ., Kingston, Ont.

Seguin, M.
1961: Geology of Knob Lake Ridge; unpubl. M.Sc. thesis, McGill Univ., Montreal.

Semenenko, N.P., Polovko, G.V., Ladieva, V.D., and Makuchina, A.A.
1956: Petrographia zhelezistychkremnistych formatsiy Ukrainskoy SSR (Petrography of the ferruginous-siliceous formations of the Ukrainian SSR); Kiev, Akademia Nauk Ukr. SSR, 535 p.

Simpson, E.S.
1929: Contributions to the mineralogy of Western Australia, ser. V, Riebeckite, Hamersley range, N.W. Div.; Proc. Roy. Soc. West. Austral., v. 16, p. 38.

Stevenson, I.M.
1962: Lac Basil, Quebec; Geol. Surv. Can., Paper 62-37, 4 p.

Stinchcomb, B.L., Levin, H.L., and Echols, D.J.
1965: Precambrian graphitic compressions of possible biologic origin from Canada; Science, v. 148, p. 75-76.

Stubbins, J.B., Blais, R.A., and Zajac, I.S.
1961: Origin of the soft iron ores of the Knob Lake range; Can. Inst. Mining Met. Trans., v. 64, p. 37-52.

Trendall, A.F.
1965: Progress report on the Brockman Iron Formation; 1964 Annual Rept., West Austral. Geol. Surv., p. 55-65.

Trendall, A.F. (cont.)
1966: Altered pyroclastic rocks in iron-formation in the Hamersley range, Western Australia; Econ. Geol., v. 61, p. 1451-1458.

1968: Three great basins of Precambrian banded iron formation deposition: A systematic comparison; Geol. Soc. Amer. Bull., v. 79, p. 1527-1544.

Turner, F.J., and Verhoogen, J.
1960: Igneous and metamorphic petrology; New York, McGraw-Hill, 602 p.

Tyler, S.A., Barghoorn, E.S., and Barret, L.P.
1957: Anthracitic coal from Precambrian upper Huronian black shale of the Iron River district, Northern Michigan; Geol. Soc. Amer. Bull., v. 68, p. 1293-1304.

Van Hise, C.R., and Leith, C.K.
1911: Geology of the Lake Superior region; U.S. Geol. Surv., Monograph 28, 641 p.

White, D.A.
1954: The stratigraphy and structure of the Mesabi range, Minnesota; Minn. Geol. Surv. Bull. 38, 92 p.

Williams, in Davies, K.G.
1963: Northern Quebec and Labrador journals and correspondence 1819-35; London, The Hudson's Bay Record Society, 415 p.

Youell, R.F., and Steadman, R.
1961: X-ray powder data for greenalite, in Brown, G., The X-ray identification and crystal structures of clay minerals; London Mineralogical Society, p. 125.

BULLETINS

Geological Survey of Canada

Bulletins present the results of detailed scientific studies on geological or related subjects. Some recent titles are listed below (Information Canada No. in brackets):

200 Part I Biostratigraphy of some Early Middle Silurian Ostracoda, eastern Canada; Part II Additional Silurian Arthropoda from arctic and eastern Canada, *by* M. J. Copeland, $1.50 (M42-200)
201 Archaeocyatha from the Mackenzie and Cassiar Mountains, Northwest Territories, *by* R. C. Handfield, $2.00 (M42-201)
202 Faunas of the Ordovician Red River Formation, Manitoba, *by* D. C. McGregor, *et al.*, $2.00 (M42-202)
203 Geology of lower Paleozoic formations, Hazen Plateau and southern Grant Land Mountains, Ellesmere Island, *by* H. P. Trettin, $3.00 (M42-203)
204 Brachiopods of the Detroit River Group (Devonian) from southwestern Ontario and adjacent areas of Michigan and Ohio, *by* J. A. Fagerstrom, $2.00 (M42-204)
205 Comparative study of the Castle River and other folds in the Eastern Cordillera of Canada, *by* D. K. Norris, $2.00 (M42-205)
206 Geomorphology and multiple glaciation in the Banff area, *by* N. W. Rutter, $2.00 (M42-206)
207 Geology of the resurgent cryptoexplosion crater at Mistastin Lake, Labrador, *by* K. L. Currie, $2.00 (M42-207)
208 The geology and origin of the Faro, Vangorda and Swim concordant zinc-lead deposits, Central Yukon Territory, *by* D. J. Templeman-Kluit, $3.00 (M42-208)
209 Redescription of *Marrella splendens* (Trilobitoidea) from the Burgess Shale, Middle Cambrian, British Columbia, *by* H. B. Whittington, $3.00 (M42-209)
210 Ordovician trilobites from the central volcanic mobile belt at New World Island, northeastern Newfoundland, *by* W. T. Dean, $2.00 (M42-210)
211 A Middle Ordovician fauna from Braeside, Ottawa Valley, Ontario, *by* H. Miriam Steele and G. Winston Sinclair, $2.00 (M42-211)
212 Lower Cambrian trilobites from the Sekwi Formation type section, Mackenzie Mountains, Northwestern Canada, *by* W. H. Fritz, $4.00 (M42-212)
213 Sequence of glacial lakes in north-central Alberta, *by* D. A. St-Onge, $2.00 (M42-213)
214 Classification and description of copper deposits, Coppermine River area, District of Mackenzie, *by* E. D. Kindle, $4.00 (M42-214)
215 Brachiopods of the Arisaig Group (Silurian-Lower Devonian) of Nova Scotia, *by* Charles W. Harper, Jr., $5.00 (M42-215)
216 Baffin Island sandurs: a study of Arctic fluvial processes, *by* M. Church, $6.00 (M42-216)
217 The geology and petrology of the alkaline carbonatite complex at Callander Bay, Ontario, *by* John Ferguson and K. L. Currie, $2.00 (M42-217)
219 Lower Cretaceous Bullhead Group, between Bullmoose Mountain and Tetsa River, Rocky Mountain Foothills, Northeastern British Columbia, *by* D. F. Stott, $6.00 (M42-219)
221 Chitinozoa and Acritarcha of the Hamilton Formation (Middle Devonian), southwestern Ontario, *by* J. A. Legault, $4.00 (M42-221)
222 Contributions to Canadian Paleontology, *by* D. E Jackson, *et al.*, $6.00 (M42-222)
223 Ordovician trilobites from the Keele Range, Northwestern Yukon Territory, *by* W. T. Dean, $2.00 (M42-223)
224 Carboniferous and Permian stratigraphy of Axel Heiberg Island and western Ellesmere Island, Canadian Arctic Archipelago, *by* R. Thorsteinsson, $6.00 (M42-224)
225 Quaternary stratigraphy of the Moose River Basin, Ontario, *by* R. G. Skinner, $3.00 (M42-225)
226 Sedimentology of Pleistocene Glacial Varves in Ontario, Canada Part "A", and Nature of Grain-size Distribution, Part "B", by Indranil Banerjee, $2.50 (M42-226)