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MEMOIR 344

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DEPARTMENT OF ENERGY,
MINES AND RESOURCES

# WAKUACH LAKE MAP-AREA,
# QUEBEC–LABRADOR (23 O)

W. R. A. Baragar

Price, $4.25

1967

WAKUACH LAKE MAP-AREA,
QUEBEC–LABRADOR (23 O)

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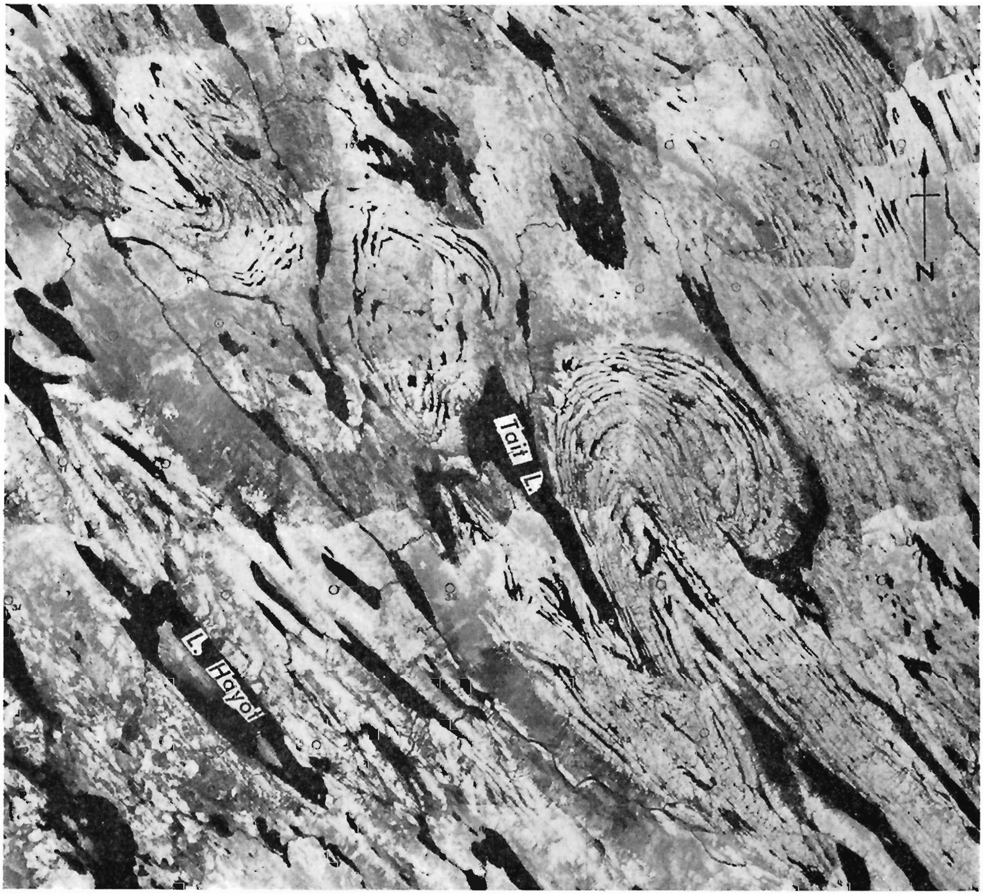


PLATE I. Interlacing folds near Tait Lake; one set plunges generally southeast, the other north to northwest. (Mosaic of R.C.A.F. RE4608 photographs)

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By
W. R. A. Baragar

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# PREFACE

The presence of iron-formation in the Labrador peninsula has been known since the explorations of A. P. Low of the Geological Survey of Canada in the years 1892–95. However, not until 1929 when the first iron ore was discovered near what is now the town of Schefferville was there incentive to explore this remote region further. After the Second World War, a surging demand for iron ore led to intensive exploration in the Quebec–Labrador region and with it the need for systematic geological studies. This report and map represent the results of one such study in an important segment of the Labrador Trough.

Y. O. FORTIER,
*Director, Geological Survey of Canada*

OTTAWA, January 20, 1965.

MEMOIR 344 — Kartenblatt Wakuach Lake (Quebec-Labrador).

Von W. R. A. Baragar

Eine eingehende geologische Untersuchung der Sedimentgesteine und vulkanischen Gesteine der Kaniapiskau-Supergruppe des Unteren Proterozoikums und beigemengter mafischer und ultramafischer Intrusivgänge im Zentralteil der Labradormulde. Im Südwesten des Kartenblatts sind Eisenerzlager aufgeschlossen worden.

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МЕМУАР 344 — Уакюу Лейк лист геологической карты, Квебек-Лабрадор.

В. Р. А. Барагар

Детальное геологическое исследование осадочных и вулканических пород нижнепротерозойской Каниапскауской супергруппы и таковое ассоциированных основных и ультраосновных пластовых интрузий в центральной части Лабрадорского прогиба. Железная руда встречается в юго-западной части листа.

# CONTENTS

CHAPTER I

## CHAPTER VI



## CHAPTER VII

## Illustrations

# WAKUACH LAKE MAP-AREA, QUEBEC–LABRADOR (23 O)


## *Abstract*

Wakuach Lake map-area covers about 5,500 square miles of the Canadian Shield in New Quebec and Labrador. It transects the Labrador Trough, a northwest-trending fold belt of little metamorphosed, Lower Proterozoic geosynclinal rocks, and includes basement gneisses of the Superior Structural Province on the west and schists and gneisses of the Churchill Structural Province on the east.

Strata of the Labrador Trough form the Kaniapiskau Supergroup composed of a lower, mainly sedimentary group (Knob Lake Group) and an upper mainly volcanic group (Doublet Group). The Knob Lake Group of shales, arkoses, quartzites, dolomites, iron-formation, and greywackes underlies the western and central parts of the Labrador Trough; the Doublet Group of pillowed and massive basalts, pyroclastic rocks, and minor sediments underlies its eastern part. Eastward, Kaniapiskau strata thicken to as much as 30,000 feet in the eastern part of the Trough.

Mafic and ultramafic sills with a possible aggregate thickness of 20,000 feet profusely intrude Kaniapiskau strata in the central and eastern parts of the Trough. Differentiation in many mafic sills produced accumulates at their base and pegmatitic segregations and granophyric veins near their tops. Chemical and petrographic studies trace the change in minerals and magma chemistry with differentiation. Both lavas and mafic sills are tholeiitic basalts; the differentiation is characterized by extreme iron enrichment. A conspicuous variety of mafic sill is glomeroporphyritic gabbro marked by 6- to 15-cm aggregates of altered feldspar in a coarse feldspathic matrix. The sill boundaries are normal gabbro and emplacement is envisaged as a type of flow differentiation. Ultramafic sills of great length form a narrow belt intruding the Doublet Group. They are commonly composed of superimposed ultramafic and gabbroic layers, a result attributed to flow differentiation modified by gravity. Tremolite-carbonate rock 20 miles east of the Trough may indicate another, parallel ultramafic belt.

Deformation of Kaniapiskau strata and schists and gneisses to the east was imposed by the Hudsonian Orogeny. Folds are overturned to the southwest and thrust faults dip northeast. Increase of metamorphic intensity eastward is recorded in three metamorphic zones: subgreenschist facies (incipient metamorphism with uneven development of pumpellyite), greenschist facies, and almandine amphibolite facies.

Iron ore has been developed in the southwest part of the map-area, but no other commercially workable deposits are yet known. Copper, nickel, zinc, lead, and asbestos prospects are present.

## *Résumé*

La région du lac Wakuach occupe environ 5,500 milles carrés du Bouclier canadien, dans le Nouveau-Québec et le Labrador. Elle occupe le géosynclinal du Labrador, qui est un faisceau de plis, d'orientation nord-ouest, composés de roches légèrement métamorphisées du Protérozoïque inférieur. Elle est formée à l'ouest de gneiss du socle de la province structurale Supérieure et à l'est de schistes et de gneiss de la province structurale de Churchill.

Les strates du géosynclinal du Labrador forment le surgroupe Kaniapiskau qui est composé d'un groupe inférieur, surtout sédimentaire (groupe du lac Knob), et d'un groupe volcanique supérieur (groupe Doublet). Le groupe du lac Knob, composé de schistes, d'arkoses, de quartzites, de dolomie, de formations ferrifères et de grauwackes, est sous-jacent aux parties occidentale et centrale du géosynclinal du Labrador. Le groupe Doublet, composé de basaltes massifs à structure en coussins, de roches pyroclastiques et d'un peu de sédiments, est sous-jacent à la partie est. Vers l'est du géosynclinal, les strates Kaniapiskau s'épaississent et atteignent près de 30,000 pieds.

Les filons-couches ferro-magnésiens et ultraferro-magnésiens d'une puissance totale possible de 20,000 pieds ont fait intrusion un peu partout dans les strates Kaniapiskau au centre et à l'est du géosynclinal. Dans plusieurs filons-couches ferro-magnésiens, la différenciation a produit des accumulations à leur base, de même que des ségrégations pegmatitiques et des veines granophyriques à leurs sommets. Des études chimiques et pétrographiques indiquent les changements que la différenciation a produits dans les minéraux et dans la constitution chimique des magmas. Les laves et les filons-couches ferro-magnésiens sont des basaltes tholéiitiques; la différenciation est caractérisée par un fort enrichissement en fer. Un genre de filon-couche ferro-magnésien très répandu consiste en un gabbro gloméroporphyritique composé d'aggrégats de 6 à 15-cm de feldspath altéré qui se trouvent dans une matrice feldspathique grossière. Les limites des filons-couches sont du gabbro ordinaire et la mise en place proviendrait d'un genre de différenciation par écoulement. Les filons-couches ultraferro-magnésiens très longs forment une bande étroite faisant intrusion dans le groupe Doublet. Ils sont habituellement composés de couches ultraferro-magnésiennes et gabbroïques surimposées, ce qui serait le résultat d'une différenciation par écoulement modifiée par la gravité. Des roches carbonatées à trémolite, à 20 milles à l'est du géosynclinal, pourraient indiquer la présence d'une autre bande ultraferro-magnésienne parallèle.

La déformation des strates Kaniapiskau, ainsi que des schistes et des gneiss de l'est, provient de l'Orogenèse hudsonienne. Les plis sont déversés au sud-ouest et les failles de poussée sont à pendage nord-est. Trois zones vers l'est révèlent un métamorphisme plus intense: l'une présente un faciès à schiste vert inférieur (métamorphisme de début avec formation irrégulière de pumpellyite), l'autre un faciès à schiste vert et la dernière, un faciès à amphibolite et à almandine.

Le minerai de fer a été exploité dans la partie sud-ouest de la région, mais aucun autre gisement de valeur commerciale n'a encore été trouvé. Il y existe des indices de cuivre, de nickel, de zinc, de plomb et d'amiante.

# Chapter I

## INTRODUCTION

Wakuach Lake map-area occupies about 5,500 square miles in the centre of the Quebec–Labrador peninsula between latitudes 55 and 56°N and longitudes 66 and 68°W. Its importance stems from its coverage of a notable segment of the Labrador Trough. Iron orebodies have been developed in the southwestern part of the Trough and numerous occurrences of other minerals of economic interest are known in other areas of it. In addition the Labrador Trough is of great geological interest, as is shown by the numerous technical papers on it in recent years. The southern five-sixths of Wakuach Lake map-area lies within the original concession areas of Hollinger North Shore Exploration Company Limited and Labrador Mining and Exploration Company Limited.

### Access

The iron-ore mining town of Schefferville, about 13 miles south of the map-area on longitude 66°50′W, is the centre of communication and supply for the area. It is served daily by scheduled airline flights from Quebec City and is connected with Seven Islands on the north shore of the St. Lawrence River by the Quebec North Shore and Labrador Railway. Access to most of the area is dependent on float- or ski-equipped aircraft, but limited access to the iron-ore region in the southwestern part is provided by a few dry weather roads. Within the area travel is best accomplished by aircraft and canoe; this combination of means of transport is particularly effective because of the strategic scattering of elongate lakes and the paucity of navigable interconnecting streams.

### History[1]

The explorations of A. P. Low of the Geological Survey of Canada between 1872 and 1895 (Low, 1896) provide the first recorded accounts of iron-formation in the Quebec–Labrador peninsula. In 1929 the New Quebec Company obtained concessions in Quebec and Labrador along the iron-bearing belt. W. F. James and J. E. Gill, sent to explore the concessions the same year, made the first discovery of commercial-grade iron ore near what is now the town of Schefferville. The economic depression of the early 1930's followed and the concession leases

Manuscript received 7 February, 1964.

[1] Much of the section is taken from Le Bourdais (1957) and Harrison (1952).

lapsed. In 1936 a new concession in Labrador was secured by Weaver (Minerals) Limited of Montreal, and Labrador Mining and Exploration Company Limited was organized to explore it under the direction of Dr. J. A. Retty. Further iron ore was discovered in 1937 and 1938. In 1939 a concession was obtained in Quebec and this also came under the management of Labrador Mining and Exploration Company. Activities were suspended during the early war years but resumed in 1942. Hollinger Consolidated Gold Mines Limited that year acquired controlling interest in Labrador Mining and Exploration Company Limited and reorganized the Quebec interests under Hollinger North Shore Exploration Company Limited. In subsequent years intensive exploration added steadily to iron ore reserves. By 1949 sufficient ore had been proved to warrant a railway stretching from the centre of the iron fields to the port of Seven Islands, some 360 miles to the south. The Iron Ore Company of Canada, formed from Hollinger Consolidated Gold Mines Limited, its two subsidiary companies, and seven American steel and mining companies, was incorporated to work the orebodies leased from the concession-holding companies. The railway was completed and the first ore shipped in 1954. From that time until 1962 production has ranged from about 7 million to 13 million tons per year.

## Previous Work

Few regions in Canada can have been more intensely studied than those of the central and southern Labrador Trough. Publications, reports, and maps relating directly or indirectly to the geology of Wakuach Lake map-area are legion. Accordingly, only work directly antecedent to that contained in this report will be mentioned here; other publications are acknowledged throughout the text.

Much of the Wakuach Lake map is a compilation of previous work contained in maps of the Iron Ore Company of Canada, The Labrador Mining and Exploration Company Limited, Hollinger North Shore Exploration Company Limited, and the Geological Survey of Canada. Figure 1 illustrates the extent of the writer's indebtedness in this respect. Most of the main ore zone and the Walsh Lake zone were mapped by company geologists on a scale of 1 inch to 1,000 feet, and the remainder of the Trough within the concession areas on a scale of 1 inch to half a mile. Geological Survey mapping is published on a scale of 1 inch to 1 mile. Previous mapping in Wakuach Lake map-area provided the basis for no less than 6 doctoral and 1 masters' theses[1]. Several additional graduate theses involving rocks of adjoining regions also contribute substantially to an understanding of geology in this area. A manual of the Iron Ore Company of Canada embodies much of the geological research done in the concession areas, particularly in regard to iron-formation and iron ores.

It might be appropriate to clarify the writer's part in the preparation of the present map. A total of four summers were spent in the map-area, two of which were devoted to mapping the Ahr Lake area (Baragar, 1958) and the other

[1] Kirkand (1950), Holmes (1950), Dufresne (1952), Frarey (1954), Kavanagh (1954), Perrault (1955), Baragar (1960).

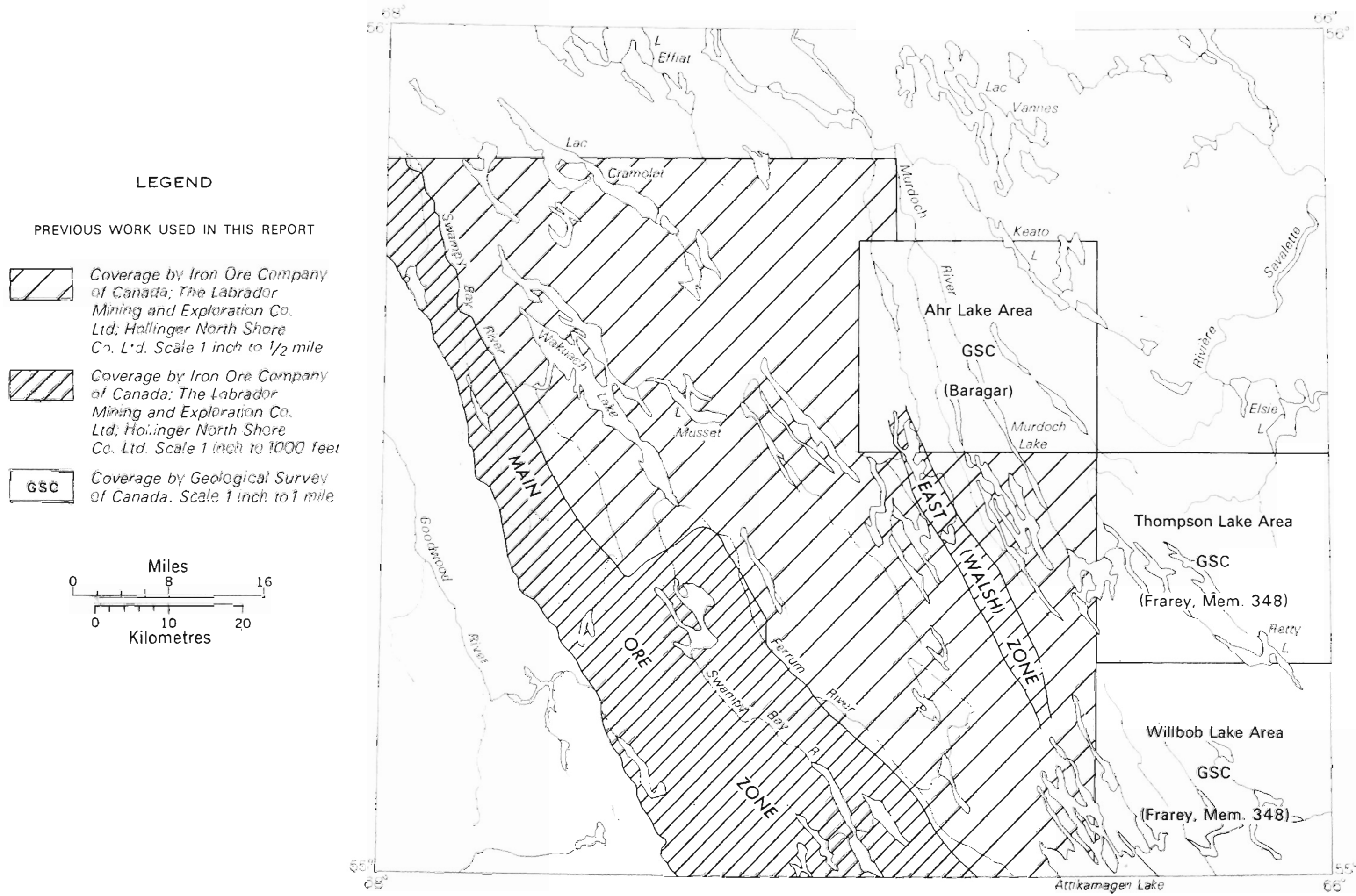


FIGURE 1. Previous mapping in Wakuach Lake map-area.

two to collecting data from remaining parts of the Wakuach Lake area. In the latter case the object was to examine and sample all major rock types at as many points as possible and to study examples of all important relationships. Within this framework priority was given to unmapped areas and areas that seemed to offer promising geological returns. Unfortunately, owing to shortage of time, the objective was not completely fulfilled. Details from company maps have been incorporated in the Wakuach Lake area map with some modification owing to the addition of new data or a different interpretation. The geology of Willbob Lake and Thompson Lake map-areas is reproduced from Frarey's published (1952) and manuscript maps, respectively, and is the subject of a separate memoir (Frarey, 1967).

## Acknowledgments

The writer is deeply indebted to management and staff of Iron Ore Company of Canada and Labrador Mining and Exploration Company for making available their maps and reports and for their numerous courtesies and services. Special thanks are due Drs. R. D. MacDonald and C. Dufresne, and Mr. G. M. Hogg.

Able assistance was rendered in the field by P. R. Lange, N. Jolly, and B. Watson in 1954; by B. E. Felber, J. Rupert, and D. C. Fraser in 1955; by S. C. J. Whelan, G. B. Skippen, and W. R. Barnes in 1958; and by P. W. H. Hay, J. B. Thibaudeau, and S. E. Asano in 1959.

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## *Chapter II*

# PHYSIOGRAPHY AND GLACIATION

## Physiography

Wakuach Lake map-area is on the central Labrador plateau mostly north of the divide separating New Quebec and Labrador. In two places the height of land that marks the boundary protrudes a few miles into the area. Land surfaces average about 2,000 feet above sea-level, but range from about 1,000 to more than 2,800 feet. They culminate in a succession of heights, mainly above 2,500 feet, that cluster about the margin of the Labrador Trough in the southwestern part of the map-area.

Three topographic zones, western, central, and eastern, corresponding to major geological divisions, provide diversity in landscapes. The *western zone,* underlain by granitic gneisses, is characterized by bold, rounded hills, which are approximately equidimensional in plan. Relief changes from 200 to 500 feet over much of the region to more than 1,000 feet among some of the hills adjoining the Labrador Trough. Landforms are influenced to only a minor degree by underlying rock structures. The *central zone* is marked in most places by the expression of folded strata of the underlying Labrador Trough, and as a result it has a distinct northwesterly lineal trend that is most pronounced in the central and eastern parts of the zone, where soft sedimentary rocks profusely intruded by gabbro sills produce a markedly corrugated topography. Cuesta or hogsback gabbro ridges from 100 to 500 feet high accurately trace out the structures of underlying rocks in sharp relief. The western part of the central zone ranges from the marshy lowlands of Swampy Bay River, underlain mainly by shales and dolomites, to the sharp ridges and rounded uplands manifested by regions of iron-formation and quartzite. The highest elevations attend the wide belt of predominantly quartzite and iron-formation adjacent to the western boundary of the Trough. The *eastern zone* is underlain mainly by schists and gneisses, and is marked generally by low relief and smoothly rounded landforms. Much of this region is overlain by muskeg and boulder plains and little can be seen of the underlying structure.

Most of the drainage is provided by four rivers: Goodwood River, near the western border of the area; Swampy Bay River, in the western part of the Trough; Murdoch River in the eastern part of the Trough; and Rivière à la Baleine in the northeastern corner of the map-area. All four rivers ultimately drain into Ungava

Bay, Goodwood and Swampy Bay by way of Kaniapiskau and Koksoak Rivers, and Murdoch by way of Wheeler River and Rivière à la Baleine.

## Glaciation

The map-area is situated near the Labrador centre of Wisconsin glaciation. Accordingly glacial deposits are generally thin, debris is of local origin, and the imprint of glaciation on landforms is marked. Smoothly polished outcrops, *roches moutonnées,* and glacial striae and grooves are ubiquitous on exposures of the more resistant rocks. Valleys and low-lying areas, mostly underlain by shales, are commonly marked by families of drumlinoid ridges. Morainic deposits, eskers, and sand plains are found in a few places in the northern part of the map-area.

Glacial features observed in the area are shown on Figure 2. Symbols denoting ice-movement with known direction of flow are based mainly on *roches moutonnées* and crag-and-tail ridges; those with an indeterminate direction of movement are based mainly on drumlinoid ridges, furrows, and flutings. For most of the area the records of ice-movement indicate a northerly flow. Individual observations largely complement one another and generally produce an integrated picture of ice-movement. Flow appears to have been northwesterly in the southern part of the area and to have swung northerly and even slightly northeasterly in the northern part of the area. In the extreme southeast, observations made by Frarey (1952) indicate two additional directions of movement: to the southeast and to the northeast. The latter direction is considered by Frarey to be the younger.

Interpretation of the glacial movements recorded in Wakuach Lake map-area is aided by reference to the Glacial Map of Canada (1959). The major ice divide forms a large arc, open to the north, that passes through the southeastern corner of Wakuach Lake map-area and has east and west extremities on the east side of Ungava Bay and Cape Wolstenholm, respectively. The opposing directions of movement to northwest and southeast, recorded by Frarey in the southeastern part of the area, presumably mark the division of flow along the ice divide. Evidence of northeasterly flow found in the same region probably indicates a very late stage of movement attributable to a glacial remnant still active in the higher areas of the original ice divide.

Most of the evidence of glacial movement shown on Figure 2 may have been produced in the last major episode of glacial flow. Henderson (1959, pp. 50–51), in a comprehensive study of glaciation in central Quebec–Labrador, cited evidence from near Eclipse and Wakuach Lakes that indicates an early and major period of glacial flow with southwestward movement. Accordingly he suggested (p. 64) that an earlier glacial centre may have existed in a region near the northwestern corner of the present map-area. The writer was not in the vicinity of Eclipse Lake, but observed no evidence of southeasterly movement elsewhere in the Wakuach Lake map-area. However, in most places the evidence of northerly movement is sufficiently marked that records of an earlier movement can be expected to have been obliterated, or at best, obscured.

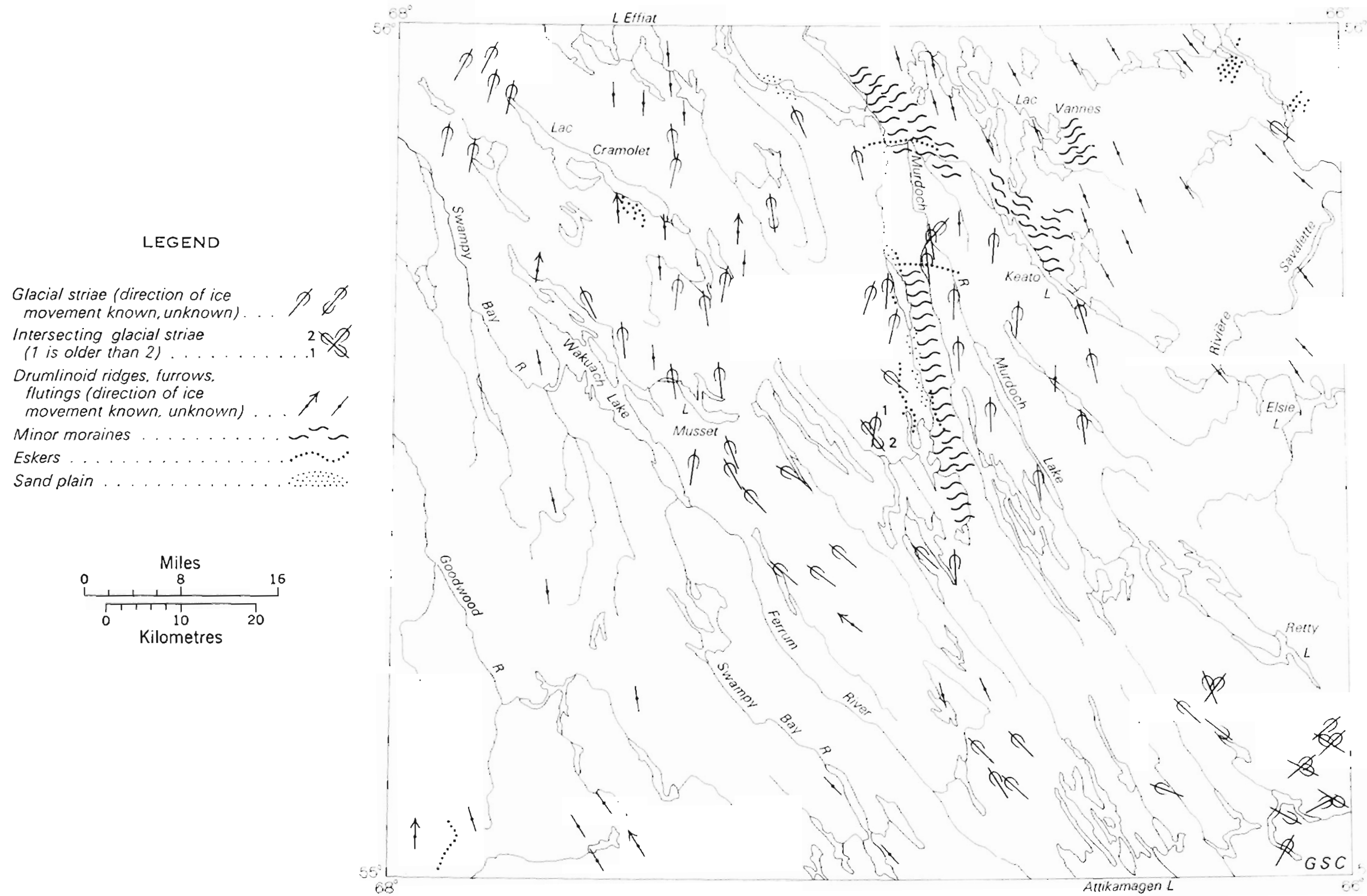


FIGURE 2. Glacial features in Wakuach Lake map-area.

Washboard or cyclic moraines of the type south of Dyke Lake (described by Henderson, 1959, pp. 18–23) occupy part of the Hurst Lake–Chassin Lake valley and the lowlands immediately northeast of the Labrador Trough. These are composed of irregular but rudely subparallel ridges oriented with long axes approximately at right angles to the direction of glacial movement. Henderson ascribed the origin of the Dyke Lake cyclic moraines to annual or polyannual deposition at the margins of active ice tongues and believed that the availability near the ice margin of easily eroded bedrock is a necessary or important factor in their formation. This condition is met in areas in which cyclic moraines have formed in Wakuach Lake map-area. The Hurst Lake–Chassin Lake valley moraine is underlain throughout its length by thinly cleaved shales and greywackes, and the moraines northeast of the Labrador Trough rest on highly fissile quartz–biotite schists.

A sand plain about 14 miles long and a maximum of 1 mile wide occupies the centre of the Hurst Lake–Chassin Lake valley just east of the belt of cyclic moraine. Abandoned stream channels and eskers, trending generally parallel with the long axis of the sand plain, mark its surface. A second sand plain occurs on the south side of Wheeler River valley at the outlet of Lac Low. There, abandoned stream channels radiating northward into the valley suggest that the sand plain was formed at the margin of a glacial lobe resting on higher lands immediately to the south. Esker deposits are rare and, apart from those associated with the sand plain in the Hurst Lake–Chassin Lake valley, are mainly oriented at right angles to the direction of glacial flow. Presumably they formed in lateral fractures near the margin of a stagnant ice sheet.

*Chapter III*

# GENERAL GEOLOGY

All rocks in the map-area are Precambrian. The southwestern corner is underlain by a mixed assemblage of gneisses and granitic rocks of intermediate composition termed here the Basement Complex. These rocks yield K–Ar ages reflecting the Kenoran orogeny of late Archaean time. They are overlain by the Lower Proterozoic Kaniapiskau Supergroup. At the contact Kaniapiskau strata dip gently eastward, but within a few miles they become involved in a northwest-trending orogenic belt of Hudsonian age that persists to the eastern boundary of the area. Deformed Kaniapiskau rocks constitute a strip about 50 miles wide that occupies the major part of the map-area. In the northeastern corner, separated from the Kaniapiskau Supergroup by a fault, is an assemblage of schists and gneisses called the Younger Complex in this report. Its antecedents are of unknown age, but they were completely recrystallized in the Hudsonian orogeny and in that sense are younger than the Kaniapiskau Supergroup.

The Kaniapiskau Supergroup subdivides naturally into lower, predominantly sedimentary, and upper, predominantly volcanic, groups called the Knob Lake and Doublet Groups, respectively (Frarey and Duffell, 1964). The Knob Lake Group is composed of red arkose and quartzite, dark shales and greywackes, dolomite, chert, quartzite, and iron-formation; the Doublet Group consists of basic pyroclastic rocks, pillowed and massive basalts, and modest amounts of dark shale, greywacke, and quartzite. Sills of the Montagnais Group, consisting of the Wakuach Gabbro and the Retty Peridotite, abound in the Kaniapiskau Supergroup, particularly in the shaly members.

The Younger Complex, composed mainly of quartz–feldspar–biotite schists, amphibolites, garnet–staurolite–mica schists, and granodiorite gneisses, is evidently derived from a mixed assemblage of sedimentary and volcanic rocks, which may be, in part, equivalent to the Kaniapiskau Supergroup.

The term "Labrador Trough" is used in this report as a non-genetic designation for the belt of folded Kaniapiskau rocks and the Montagnais Group. Its eastern boundary is taken to be the fault through Keato Lake. Used in this sense it has considerable advantage over the alternate term, "Labrador geosyncline", which refers to an ancestral state that cannot now be defined in space.

*Table of Formations*

| Era | Supergroup or Complex (Orogeny) | Group | Formation (thickness in feet) | Lithology |
|---|---|---|---|---|
| | | | | Diabase dyke |
| Intrusive into Kaniapiskau Supergroup | | | | |
| PROTEROZOIC | Younger Complex (Hudsonian) | | | Quartz–feldspar–biotite schist, garnet–staurolite–mica schists, granodiorite gneiss, amphibolite, tremolite–actinolite rock |
| | Relations Unknown | | | |
| | | Montagnais | Retty Peridotite | Serpentinite, peridotite, may be concurrent with Wakuach Gabbro |
| | | | Wakuach Gabbro | Gabbro, meta-gabbro, glomeroporphyritic gabbro (leopard rock) |
| | Intrusive into Kaniapiskau Supergroup probably with chronological overlap | | | |
| | Kaniapiskau | Doublet | Willbob (15,000+) | Meta-basalt, minor interlayered slate and tuffaceous metasediments |
| | | | Thompson Lake (1500–2300) | Metasedimentary rocks, mainly slate, siltstone, quartzite, greywacke; minor basalt |
| | | | (0–600) | Iron-formation |
| | | | Murdoch (3000–6000)? | Basaltic agglomerates, breccias, lapilli, tuff; minor basalt, acidic tuffs, and metasediments |
| | | Knob Lake | Menihek (0–7000) | Carbonaceous shale, siltstone, and greywacke; some quartzite and subgreywacke; basalts and pyroclastic rocks; minor dolomite and chert |
| | | | Purdy (0–1500+) | Dolomite |
| | | | Disconformity | |
| | | | Sokoman (500–800) | Iron-formation |
| | | | Ruth (50–100) | Ferruginous and carbonaceous slate and shale |
| | | | Wishart (100–300) | Quartzite, feldspathic quartzite, arkose, subgreywacke, siltstone, chert; minor shale, conglomerate |

*Table of Formations* (conc.)

| Era | Supergroup or Complex (Orogeny) | Group | Formation (thickness in feet) | Lithology |
|---|---|---|---|---|
| | | | Local disconformity | |
| PROTEROZOIC | Kaniapiskau | | Fleming (0–400) | Chert breccia, minor siliceous argillite |
| | | | Denault (0 4200) | Dolomite, minor chert |
| | | | Attikamagen (1000+ - 2500+) | Black, grey, red, and green shales, carbonaceous greywackes and subgreywackes. quartzites; basalts and basaltic pyroclastic rocks; minor dolomite |
| | | | (10,000?) | Dolomite, cherty quartzite, carbonaceous shale, cherty greywacke |
| | | | Local disconformity ? with Attikamagen Formation | |
| | | | Seward (2000+– 4500+) | Red beds; arkose, feldspathic quartzite, quartzite, dolomite, siltstone, shale, conglomerate. Minor non-red beds; marl, limestone, dolomite conglomerate. Basic lavas, pyroclastic rocks, tuffaceous sediments |
| | Angular Unconformity | | | |
| ARCHAEAN | Basement Complex (Kenoran) | | | Granite and quartz monzonite |
| | | Intrusive Contact | | |
| | | | | Pyroxene-bearing granodiorite–syenodiorite |
| | | | | Pyroxene-bearing banded gneisses of granodioritic composition |

## Basement Complex

The Basement Complex is an involved assemblage of crystalline rocks that occupies the southwestern part of the map-area and pre-dates the Labrador Trough. The rocks have not been mapped in detail; accordingly, the subdivisions shown are highly generalized. The positions of the contacts were largely interpreted from aerial photographs aided by limited helicopter reconnaissance. With the exception of a single traverse west of Ritchie Lake, ground observations are confined to the region south of latitude 55°20′N and the following remarks apply to rocks within this region.

## Banded Gneisses

This unit underlies most of the region west of the Labrador Trough. It comprises a wide assortment of quartzo-feldspathic gneisses and intimately associated granitic and pegmatitic bodies. Dark, thinly layered, quartz–feldspar–biotite gneisses appear to be the most prevalent rock-types, but they are commonly accompanied by a profusion of white, granitic, or pegmatitic seams, lenses, pods, and dykes (Pl. II). The granitic bodies are generally layers or lenses an inch to a foot thick that parallel the foliation or, more rarely, crosscut it. Where they are closely spaced rocks are typical *lit-par-lit* gneisses. In places discrete granitic masses reach dimensions of several tens and even several hundreds of feet across. All such granitic bodies are grouped with the banded gneisses on Wakuach map-sheet, but genetically they should probably be associated with the granite and quartz monzonite of map-unit 3. The quartzo-feldspathic gneisses are mainly fine grained ($\frac{1}{2}$-1 mm), well-foliated rocks. Commonly the foliation is accompanied by

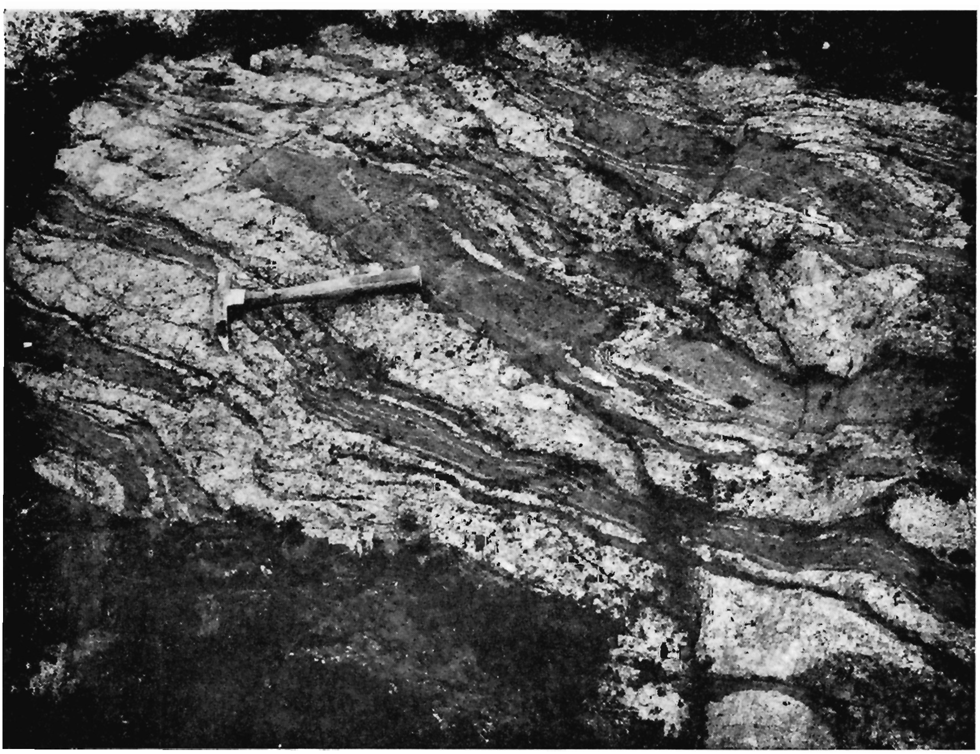

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PLATE II. Banded gneisses seamed with granitic layers. Basement Complex, near the south arm of Canoe Lake.

a distinct compositional layering, which marks a differential concentration of mafic and felsic components. Biotite is the predominant mafic mineral, but rarely pyroxene may be seen in hand specimen and garnet is sporadically present. Generally the rock contains from about 10 to 40 per cent by volume mafic minerals.

Most specimens of gneiss examined in thin section may be classified as granodiorite or quartz diorite. A typical specimen would contain the following minerals within the range of composition given:

| | |
|---|---|
| quartz | 10 to 30 per cent |
| plagioclase (oligoclase–andesine) | 45 to 65 per cent |
| potash feldspar (generally microcline) | 0 to 5 per cent, rarely 15 to 20 per cent |
| biotite | 10 to 30 per cent |
| hypersthene | 5 to 10 per cent |

Garnet and diopside are less regular constituents and accessory minerals include apatite, zircon, and iron oxide. Commonly the mafic minerals are partly or completely replaced by chlorite and amphibole. Textures are typically metamorphic; crystals are anhedral and intergrown. In places plagioclase and microcline show complex intergrowths at their mutual boundaries reminiscent of myrmekitic textures. Perthitic and antiperthitic structures are developed to varying degrees. The presence of hypersthene and the modest content of potash feldspar are the most characteristic features of the gneisses.

The compositions of associated granitic or pegmatitic layers, lenses, or masses contrast markedly with those of the gneisses. They are mainly granites or quartz monzonites. Potash feldspar is a major constituent and biotite is generally minor. Garnet is a common accessory mineral.

The banded gneisses are complexly folded along what appear to be generally northwestward-trending axes. However, the major trends are obscured in the swirls and sinuous forms that the structural elements describe in a more detailed scale. Dips are generally steep northeastwards. Locally the gneisses may be severely crenulated and granitic interlayers commonly show marked thinning and swelling or boudinage structure.

The widespread occurrence of hypersthene and garnet in the banded gneisses is presumably indicative of the granulite facies of metamorphism. On the other hand, biotite does not typically belong to the granulite facies where orthoclase is the major potash-bearing mineral. In these gneisses biotite is generally abundant and potash feldspar rare. Possibly the explanation lies in the amount of water present during metamorphism. If water took part in the reactions but was not present in excess, then it should be considered a component and the resulting rock could be permitted an additional hydrous phase, in this case biotite. Biotite does appear as a stable phase in these assemblages. These rocks also have many of the characteristics of charnockite, e.g., the presence of hypersthene and in places diopside, the distinctive pink to green pleochroism of hypersthene, and the prevalence among the feldspars of perthitic and antiperthitic textures.[1]

The age of the banded gneisses, as determined by the K–Ar method on biotite from a typical specimen taken near the southern border of the area, is 2,365 million years (Lowden, 1960, p. 28). Although this is slightly lower than the average Kenoran age of about 2,500 million years (Stockwell, 1961), the metamorphism of these rocks is obviously to be attributed to this orogeny. Age determinations on other specimens taken from equivalent positions just west of the Labrador Trough, north and south of the present area, gave 2,440–2,505 (Lowden, 1961, p. 73, and Fig. 2) and 2,425 million years, respectively.

## Granodiorite–Syenodiorite

The granodiorite–syenodiorites are a mass of unfoliated or poorly foliated medium-grained rocks that occur within the region of banded gneiss. They have not been well defined in the field; the boundary shown on the accompanying Wakuach Lake map-sheet (*in pocket*) is based entirely on airphoto interpretation, but these rocks have such a distinctive appearance and are so similar in the two widely separated localities where they were observed, that to the writer a separate classification seems warranted.

The rocks are massive, even-grained plutons with a peculiar greenish grey colour. They contain minor mafic minerals and very little quartz that can be recognized in hand specimen. Following are rough estimates of mineral compositions made on thin sections of rocks from the two localities visited:

| | Per cent | |
|---|---|---|
| quartz | 10 | None |
| microcline | 25 | 25 |
| plagioclase ($An_{23-27}$) | 55 | 65 |
| pyroxene | 6 | None |
| hornblende | None | 10 |
| biotite | 3 | None |

In addition to the minerals listed above iron oxide and apatite are accessory minerals. Both hypersthene and diopside are present. Microcline is generally perthite and plagioclase is antiperthite. Biotite and hornblende are both greenish brown varieties. The texture is granitic.

These rocks are similar to the surrounding gneisses in that they contain pyroxene, perthites, and antiperthites, but have few other likenesses. The potash feldspar content is substantial, in contrast to the gneisses, and quartz and biotite

[1] De Waard (1965) recently subdivided the granulite facies into six subfacies that reflect variations in load and water pressures within the granulite field of stability. The mineralogy typical of the banded gneisses, biotite–plagioclase–orthopyroxene, is characteristic of his hornblende–orthopyroxene plagioclase subfacies of intermediate load and water pressures. Clinopyroxene and garnet in addition would signify somewhat higher load pressures according to de Waard.

are relatively minor. Like the banded gneisses they display some of the characteristics of charnockites. The relationship between the granodiorite–syenodiorite mass and the banded gneisses is unknown. Presumably, since they are relatively unfoliated, they are younger.

### Granite–Quartz Monzonite

Granite–quartz monzonite forms a small stock of about 25 square miles within the banded gneisses. It has been outlined on Map 1209A mainly by air-photo interpretation and helicopter reconnaissance, but the rock shows such a marked contrast in appearance against the dark gneisses that its boundaries are probably reasonably well defined.

The rock is massive or, rarely, weakly foliated, white, even-grained granite or quartz monzonite. Generally it is medium grained, but locally it coarsens to pegmatite. In places it is sparsely garnetiferous. The mineral composition, estimated from a few thin sections, is generally as follows:

| | |
|---|---|
| quartz | 25–35 per cent |
| microcline | 25–45 per cent |
| plagioclase (oligoclase) | 20–30 per cent |
| biotite | Under 5 per cent |

One specimen, which proved to be granodiorite, contains minor pyroxene and hornblende as well as nearly 10 per cent biotite. It may have been contaminated with wall-rock material.

Contacts between the granite–quartz monzonite stock and the surrounding banded gneisses were not observed on the ground but appear to be sharp. Near the contact, thick granitic layers are abundant in the banded gneisses, and farther away, thin layers, lenses, and pods of similar composition are an integral part of the country rock. The granitic stock itself contains, in places, numerous rounded inclusions, 10 to 15 feet in diameter, of recrystallized mafic gneiss.

The granite–quartz monzonite appears to be synorogenic in age. Structural trend-lines in the surrounding gneisses closely parallel the boundaries of the stock, giving "wrap-around" structure so typical of synorogenic intrusions. The age determination mentioned on a previous page (2,365 million years) dates the orogeny as Kenoran.

## Knob Lake Group

The Knob Lake group is the lower of the two major rock divisions of the Labrador Trough in this area and rests unconformably on the Basement Complex. The oldest units of the Trough are not found in direct contact with the basement rocks, but appear in the cores of anticlinoria and thrust slices well east of the contact.

## Seward Formation and Underlying Volcanic Rocks

The Seward (red bed) Formation and underlying volcanic unit representing the oldest known rock units in the Labrador Trough are here combined for purposes of description. The volcanic unit is neither sufficiently widespread nor well exposed for one to be able to assess its likely importance as a separate stratigraphic member in the Trough succession. The name, taken from the Seward Grits near Seward Lake, was recently extended by Frarey and Duffell (1964) to include equivalent assemblages elsewhere in the Trough and is therefore used in this report.

### *Distribution and Thickness*

Exposures of the volcanic unit are confined to a few square miles on the south side of Lac Musset. Its thickness is entirely unknown.

The distribution of the Seward Formation, shown on Figure 3, is to a considerable extent a result of structure and does not necessarily reflect the extent of deposition. The outcrop areas are mainly in the cores of anticlines or anticlinoria and are all on the up-thrown side of a major thrust fault. The total thickness of the formation could not be determined, but measurements of partial sections at a few places give some lower limits to possible thicknesses. The formation has a minimum estimated thickness near Lac Cramolet and Lac Musset of 3,500 feet, adjoining Pickup Lake, 4,500 feet, and near Lac Moranbert, 2,000 feet. Probably by its very nature the formation can be expected to vary widely in thickness, particularly with distance from the margin of the basin of deposition.

### *Lithology*

*Volcanic Rocks*

The volcanic rocks of the lower unit are composed of grey to green lavas and mottled red and green or green, poorly defined fragmental rocks. Bedding is rare or absent in the fragmental rocks and the lavas are massive. The fragments are as much as 6 inches long, but most are less than a quarter inch across. Many of the fragments are red sandstone or siltstone. Rounded quartz particles are apparent in the groundmass.

In thin section, the fragmental rocks can be seen to consist of mixed fragments of siltstone or quartzite together with patches of chloritic and sericitic material. Commonly, the siltstone and quartzite fragments are heavily impregnated with hematite, and a few fragments are predominantly hematite. The groundmass is a confused mixture of rounded quartz and feldspar grains cemented with hematite and of shredded masses of chlorite and sericite containing euhedral plagioclase locally. The lava specimens observed in thin section are considerably altered. Plagioclase is now albite and the ferromagnesian minerals are hornblende and chlorite. Probably the lavas were originally intermediate to basic in composition.

The mixed nature of the fragmental rocks in company with known lavas suggests that they are sandy or silty tuffs and that the assemblage represents a period of concurrent volcanism and sedimentation. As hematite is a common cementing

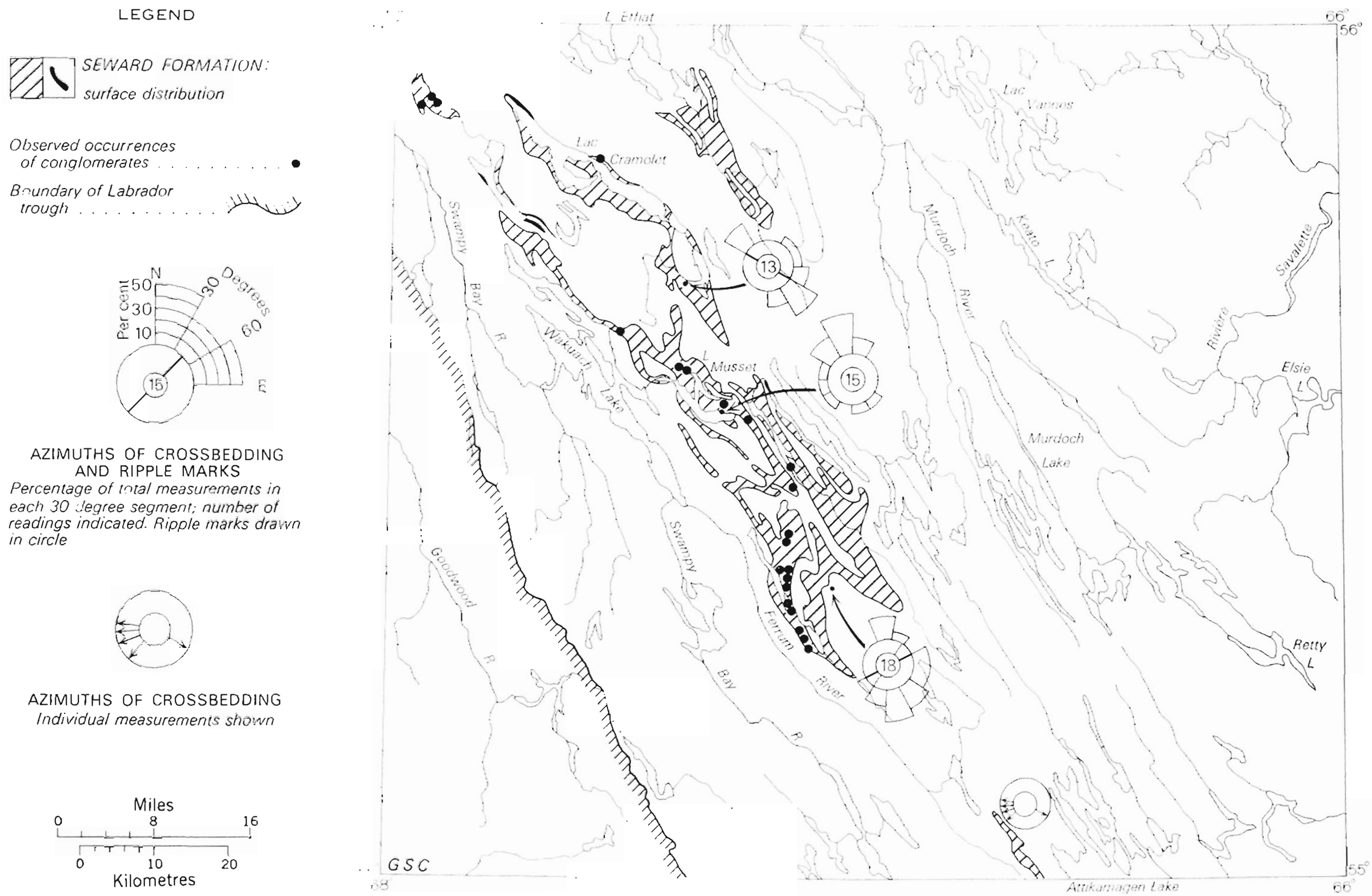


FIGURE 3. Distribution of the Seward Formation in Wakuach Lake map-area and current direction data.

material in both the fragmental rocks and the overlying red beds, the volcanic sequence presumably grades into the Seward Formation. At the south end of Lac Musset the mixed volcanic–sedimentary rocks are clearly overlain by red beds but the base of the volcanic succession is not exposed. The rocks could, therefore, represent either the final phase of a fairly long period of volcanism that merged with the succeeding period of sedimentation, or a limited episode of volcanism concurrent with sedimentation. About 3½ miles north of Lac Billiard, a 10-foot-thick member of similarly intermixed volcanic and sedimentary fragmental rock is interbedded with red sandstones in a thick succession of the latter. Thus the existence of short periods of volcanism contemporaneous with red bed deposition can be demonstrated.

*Seward Formation*

The Seward Formation comprises a wide variety of sedimentary rocks nearly all of which are marked by a red or pinkish colour. These include conglomerate, sandstones, siltstones, shales, dolomite, and limestone. Most common are sandstones and siltstones whose composition ranges mainly from arkosic to quartz-rich. Feldspathic sandstone is probably the most characteristic variety.

Most of the conglomerates are in poorly defined layers or lenses intimately associated with arkoses or feldspathic sandstones and they generally form a subordinate part of the local assemblages. They are mainly composed of sparsely scattered quartz and feldspar pebbles, mostly less than half an inch in diameter, in an arkosic matrix. Rarely are the pebble sizes greater than an inch across. An exception is a conglomerate that outcrops in a brook on the east side of Wakuach Lake about 10½ miles from its north end. There, fairly close-packed pebbles, as much as 3 inches in diameter are composed of pink limestones and shales as well as quartz and feldspar. Generally the conglomerates are poorly sorted and not well bedded.

Sandstones and siltstones are closely associated and in places are finely interbedded. They range in colour from light pink to deep brick or purplish red. Bedding is evident in most places, but is generally not sharply defined and rarely produces a fissility. It is most commonly formed by a gradational change in grain size or a vague colour layering. Crossbedding is common in both the sandstones and conglomerates.

The principal components of the conglomerates, sandstones, and siltstones are quartz, feldspar, plagioclase, sericite, and hematite. Characteristic compositions might be anywhere within the following percentage ranges:

| | |
|---|---|
| quartz | 35–75 per cent |
| potash feldspar | 10–30 per cent |
| plagioclase | 5–25 per cent |
| sericite | 5–20 per cent |
| hematite | 1– 5 per cent |

Chlorite, biotite, and unidentified micaceous minerals are additional, normally minor components. Plagioclase is invariably albite or sodic oligoclase and potash feldspar is generally microcline and may be perthitic. In addition to the characteristic compositional range given above, a few examples of almost pure quartzite and also of highly feldspathic sandstone were observed.

Grains or pebbles in the sandstones, siltstones, and conglomerate range from angular to well-rounded with possibly a slight preponderance of examples in the subangular to subrounded categories. Commonly the rocks are bimodal in grain size. Grains of one order of size are embedded in a matrix of much smaller average grain size. In such rocks quartz appears to be favoured in the coarser grains and feldspar in the matrix. Sericite with minor amounts of chlorite, biotite, and possibly other micaceous minerals is generally an important component of the groundmass. In some rocks it is an interstitial filling, in others it forms a continuous median enclosing the detrital grains. Probably most of the sericite developed during diagenesis; locally sericite crystals can be seen growing into or out from detrital grains. Quartz rarely forms a cementing material in these rocks.

The red colour is due to hematite, which coats most of the grains and is disseminated through the interstitial or groundmass material. It varies in concentration from layer to layer and is generally the most conspicuous marker of bedding planes. In many places it clouds the potash feldspar grains, but is seldom found in other minerals. Rod-like masses up to 1 mm long and blebs 0.1 to 0.2 mm in diameter of both earthy and metallic hematite are common in many of the sandstones and siltstones. Generally the rod-like masses parallel the bedding and may be either evenly dispersed through the rock or concentrated in certain beds. Siltstones, in general, appear to have a greater content of hematite than either the sandstones or conglomerates.

Carbonate rocks appear at many levels in the Seward Formation, but probably compose only a small part of the formation as a whole. Dolomites predominate; they are generally pink and commonly contain some intermixed sand and silt grains. In a few places algal structures are well developed. Sandy dolomites are crossbedded in many places. The carbonate rocks are finely crystalline, generally less than 0.05 mm in crystal size, and dusty hematite is well dispersed through the crystals of the rock.

Thin-bedded red shales are widespread but not abundant. They are generally interbedded locally with sandstones or siltstones and do not themselves seem to form significant thicknesses.

Rocks that outcrop on parts of the shore and islands of Lac Cramolet and Lac Ribero have been included with the Seward Formation but are not typical of it elsewhere. Conglomerate composed of dolomite boulders in a dark quartzite matrix is exposed intermittently along the eastern shoreline of Lac Cramolet for a distance of about 3 miles from the north end of the lake. The boulders range from about a quarter inch to 8 feet in size and from angular to rounded in shape. Most of them are in the size range of 3 to 6 inches. Virtually all are dark grey dolomite or sandy dolomite. Many of the boulders are slab-like

W.R.A.B. 2-31-59

PLATE III. Dolomite conglomerate, Seward Formation, Lac Cramolet, showing interlacing quartzite beds.

and parallel the bedding planes. Quartzite beds up to several feet thick are interlaced with the conglomerate in a manner reminiscent of large-scale crossbedding (Plate III) and in places grade into it along strike. The conglomerate layer is at least 30 to 50 feet thick, the maximum distance exposed above the lake surface. In places it is succeeded by a layer a few feet thick of well-sorted white quartzite, in others by shale containing quartzite pebbles. A gabbro sill just above the white quartzite and shale obscures the section, but the conglomerate seems to be succeeded ultimately by dark shales of the Attikamagen Formation. About 6 miles south on the same shore of Lac Cramolet opposite the prominent central peninsula, conglomeratic arkose, fairly typical of the Seward Formation elsewhere, contains beds, 3 to 5 feet thick, of dolomite conglomerate. These differ from the dolomite conglomerates farther north in that the matrix is conglomeratic arkose like the adjoining rock. Nevertheless they are sufficiently similar to link the dolomite conglomerates to the Seward Formation. These outcrops are also overlain by dark grey shales considered to be part of the Attikamagen Formation.

On islands just south of the prominent central peninsula of Lac Cramolet, and on a headland of the west shore about 2 miles farther north, a sequence of thin- to thick-bedded, greenish grey and white marls and limestones outcrops. These rocks occur within the core region of the major anticline that lies along Lac

Cramolet and underlie a thick assemblage of typical red beds. Despite their lack of reddish colour they have been grouped with the red bed formation for convenience, but possibly should be given separate status. On the islands where they closely overlie a gabbro sill the beds have been severely drag-folded and have undergone some degree of contact metamorphism.

At the northeast tip of Lac Ribero and on the nearby islands, a chaotic assemblage of largely metamorphosed, carbonate fragmental rocks is exposed. The fragments or boulders are rounded to sharply angular, and range from a quarter inch to about 5 feet across. Most of the fragments were probably dolomite at first but are now coarsely crystalline lime silicate minerals, mainly actinolite. Commonly blocks or boulders have a rim of lime silicate minerals and a carbonate interior (Pl. IV). Some of the matrix is nearly pure quartzite; elsewhere it must have been dolomitic quartzite, for it is now largely lime silicate minerals. In a few places the thick lime silicate rims of large boulders contain masses of magnetite several inches across. Some of the boulders are well-bedded quartzitic rocks, probably dolomitic quartzite. On one of the small islands, blocks of pink dolomite or limestone appear in a matrix of white dolomite. Metamorphism of these rocks is most likely attributable to thermal effects from a nearby gabbro sill. Overlying

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PLATE IV. Metamorphosed fragmental rock, Lac Ribero. Fragments are mainly meta-dolomite in a siliceous matrix. Note reaction rims.

gabbro can be seen in contact with the fragmental rocks in outcrop on shore and undoubtedly they are also closely underlain by gabbro. The fragmental or conglomeratic assemblage is estimated to be at least 300 feet thick. It is completely isolated from other stratigraphic units, but much red siltstone is found on the east shore of this arm of the lake about half a mile distant along strike. On a point of the east shore about 2½ miles to the south, an outcrop of coarse red sandstone, typical of the Seward Formation, contains layers and blocks of dolomite. The fragmental rocks, therefore, are probably part of the Seward Formation. If so, they must be near its upper contact, for intersill sediments outcropping a short distance to the east are grey siltstones thought to belong to the Attikamagen Formation. The similarity of general composition, structure, and stratigraphic relations between the metamorphosed fragmental rocks of Lac Ribero and dolomite conglomerates of Lac Cramolet is noteworthy. Possibly they are correlative.

### *Crossbedding and Ripple-Marks*

Crossbedding ranges from units 1 inch or 2 inches to units 10 feet in thickness, but is generally less than 1 foot thick. The thickest crossbedded units noted are at the southeastern extremity of Lac Musset where several are in excess of 7 feet. Another area of thicker-than-normal crossbedding is near Lac de l'Axe, where units 2 to 4 feet thick are common. The crossbedding is of several types, from minutely lenticular to uniformly tabular. Many of the thicker units are of the "torrential type" in which the foreset beds show little curvature and meet topset and bottomset beds at similar angles (Pl. V). Current directions determined by measuring the foreset and true bedding at a number of places and restoring the latter to a horizontal position are shown in histogram form on Figure 3. The technique is similar to that used by Pelletier (1958, pp. 1043–1044) and detailed by Ramsay (1961). The tabular variety of crossbedding was used for this purpose and corrections were made for dip and plunge. The values used for plunge, however, were average regional values and in some instances little more than informed guesses. Nevertheless, as the regional plunges are rarely greater than 25 degrees, the maximum probable error in current direction according to Ramsay's graph for flexure-folds (1961, p. 89) would be about 25 degrees. The error in most cases would be considerably less and probably the current directions shown in the diagram are roughly valid. The folding was considered to be simple flexure folding. The indicated location of each histogram is approximately the centre of the region from which measurements were derived.

Ripple-marks were noted in a few places but are not plentiful. The directions of three sets of ripple-mark crests in tilt- and plunge-restored strata are shown in the histogram figures of Figure 3. In the southernmost histogram on this figure the ripple-marks are nearly at right angles to the dominant current directions as might be expected. In the only other histogram with ripple-marks recorded they appear to correlate with a minor current direction.

W.R.A.B. 4-4-59

PLATE V. Torrential crossbedding in red feldspathic quartzites, Seward Formation, Lac de l'Axe. The topset beds can be seen in the upper left corner of the photograph above the head of the pick. The top of the bottomset beds is marked by the plane that crosses the foreset beds near the point of the lower pick.

### *External Structural Relations*

Fragmental rocks or conglomerates at the top of the Seward Formation at Lac Cramolet and Lac Ribero are indicative of some terminal tectonic disturbance even though structural relations with overlying Attikamagen Formation are generally conformable. The contact was not actually observed, however, and modest or local disconformities may not be apparent in regional relationships. Apart from the volcanic rocks near Lac Musset, relationship between the Seward Formation and older strata (if they exist in the map-area) are unknown.

### *Mode of Origin*

The Seward Formation evidently accumulated rapidly from a mainly granitic source. The moderate angularity of its grains and the importance of fairly fresh feldspar as a constituent preclude deep weathering of the source and long transportation. The conditions of deposition implied by the various types of cross-bedding, ripple-marks, algal structures, and intercalated dolomite beds range from terrestrial, through deltaic to estuarine or shallow marine.

The origin of red beds is controversial; the main problem is whether the red colour is due to oxidation of iron compounds at the source of the sediments or at their point of deposition (Krumbein and Sloss, 1955, pp. 369–370). In recent classifications of red beds different origins were assigned to different types of red sediments (Krynine, 1949; Clark, 1962). The red beds of this report would be classified under Clark's "Brick Red Beds", which are assumed to be "deltaic or littoral to estuarine" sediments deposited under desert conditions (Clark, 1962, Table 1). Clark (p. 426) maintained that the presence of fresh feldspar was difficult to reconcile with the deep weathering necessary to produce lateritic material in the source area, and proposed instead laterization of the sediments at or near their site of deposition. This would not preclude the formation of some laterite at the source and perhaps at intermediate points, with its ultimate deposition in the main body of sediments. In view of the abundance of fresh feldspar in red beds of Wakuach Lake map-area Clark's explanation applied to the problem in this area would seem to have considerable merit.

The location of source material for the red beds in this area is not clear. In the central and northern parts of the area the meagre crossbedding data indicate longitudinal currents with opposing directions of flow (Fig. 3). A plot on the same diagram of all conglomerate beds sighted, however, shows a preponderance of conglomerate occurrences on the western side of the outcrop area, which suggests a western source. The following explanation is believed to utilize both sets of data. The source was a highland area to the west and sediments were carried into the trough by easterly flowing streams. Within the trough distribution was by currents, either stream or marine, flowing along the axis of the trough. Differential subsidence, or differential accumulation of sediments in the course of time resulted in periodic reversals of flow. In the southern part of the area the few data available indicate a westerly flow of currents. Possibly these sediments were derived from a basement source to the east or southeast, as postulated for sediments of the Seward Formation in the type area, in Michikamau Lake map-area (Wynne-Edwards, 1960[1]).

Volcanism accompanied the formation in part and may have preceded it. Some of the minor mafic constituents of the red beds (biotite, chlorite, and possibly even part of the hematite) may therefore have had a volcanic origin.

### *Age and Correlation*

The Seward Formation is clearly older than the Attikamagen Formation for it appears in the centres of anticlinoria that involve the latter. Further, it is composed mainly of granitic material and might reasonably be considered to rest in large part on a granitic basement. It is, therefore, thought to be the oldest sedimentary

[1] The formation is not named on Wynne-Edwards map but is at the locality designated by Frarey and Duffell (1964, p. 3) as the type locality of the Seward Formation.

formation in the Trough assemblage. Both Fahrig (1956a) and Roscoe (1957, pp. 3–6) recorded reddish arkosic rocks as their basal unit in the Trough assemblage of Cambrian Lake map-area, immediately northwest of the present map-area. In its type area, about 50 miles southeast of the southernmost exposure of red beds in this map-area, the Seward Formation, an assemblage of crossbedded conglomeratic arkoses, arkosic quartzites, and greywackes (Wynne-Edwards, 1960) occupies an equivalent stratigraphic position. All these units are probably correlative although they may not necessarily be continuous nor even exact time equivalents.

## Pre-Attikamagen Dolomite and Associated Quartzite, Greywacke, and Shale

### *Distribution and Thickness*

This group of rocks has been mapped as a separate unit continuously for 40 miles in the western part of the map-area. It is unknown elsewhere in the map-area. If one assumes little repetition of beds by internal folding and faulting the formation has a thickness of about 10,000 feet near the north arm of Wakuach Lake.

### *Lithology*

Adjacent to Wakuach Lake the formation is mainly dolomite with interlayers of quartzite, black carbonaceous shale, and black cherty greywacke. Non-carbonate rocks comprise about 20 per cent of the formation and are distributed throughout the section. West of Lac Ribero and Lac des Deux Iles the lithology is reversed; quartzite, shale, and greywacke predominate and minor dolomite is dispersed through the section.

The dolomite is a white or light grey, fine-grained (0.05 mm) rock with generally buff weathering surfaces. Commonly it is thick bedded, but bedding may not be well defined unless the dolomite is also sandy or silty. In many places the bedded dolomites show evidence of deformation penecontemporaneous with their deposition; highly contorted beds are truncated by undisturbed overlying beds. Chert layers are interbedded with dolomite in places but are not common.

The quartzites are light to dark grey rocks composed generally of 80 to 90 per cent quartz. Locally they exhibit faint crossbedding. Some beds are cemented with carbonate and weather to a rusty, crumbly surface. The shales are black, generally highly fissile, and thinly bedded. A few shaly layers intimately associated with dolomite beds are composed of thinly interlayered black chert and siltstone. The greywackes are dark, quartz-rich rocks containing numerous grains, fragments, and discontinuous layers of black chert. In thin sections, the quartzite can be seen to be composed generally of rounded to well-rounded grains of quartz and possibly 5 to 10 per cent potash feldspar set in a matrix of one or more of chlorite, sericite, cryptocrystalline quartz, carbonate, and finely divided opaque iron oxide. Grains of chert or cryptocrystalline quartz are locally present. With increasing quantities of chert or cryptocrystalline quartz in the form of grains, matrix, or lithic fragments, quartzite passes into the rock that is here termed greywacke. A typical specimen is composed as follows:

| | |
|---|---|
| cryptocrystalline quartz | 50–55 per cent |
| quartz grains | 5–10 per cent |
| carbonate | 35 per cent |
| chlorite | 1– 2 per cent |

Other specimens contain, in addition, minor potash feldspar grains and in the matrix small quantities of sericite, iron oxide, and biotite. The chert is generally brown and may be either cryptocrystalline or submicrocrystalline. It is commonly microscopically layered in a highly irregular fashion.

### *External Structural Relations, Correlation, and Age*

Dolomites and associated non-carbonate rocks clearly underlie slates and greywackes assigned to the Attikamagen Formation where the relationships are exposed at the major bend of Swampy Bay River. The folded dolomite formation plunges southward beneath the Attikamagen Formation. Along the west side of the fold the strike of bedding in both formations is roughly conformable; elsewhere the relationship between the two formations is not clear. The writer has tentatively assumed that the dolomite formation is everywhere older than the Attikamagen Formation and has made his interpretations accordingly. Thus a fault is postulated along the contact between the dolomite formation and Attikamagen Formation in the east arm of Wakuach Lake. Southeast of the lake the same part of the dolomite formation is assumed to plunge beneath the Attikamagen Formation as an overturned anticline. Alternatively, the dolomite may be interbedded with the Attikamagen Formation and therefore bounded by it above and below. Possibly the facies change northward represents an interfingering of the dolomite formation into the surrounding Attikamagen slates and greywackes. There is no definite evidence to support either theory.

Age relations between the dolomite formation and the Seward Formation are unknown.

## Attikamagen Formation

### *Distribution and Thickness*

The outcrop pattern of the Attikamagen Formation together with an interpretation of its probable distribution before folding is shown on Figure 4. As shown by this figure the Attikamagen Formation does not extend to the present western boundary of the Labrador Trough but falls 2 to 4 miles short of it. Most of the rocks included with the Attikamagen Formation here are dated by their stratigraphic relationship to other formations. However, as is almost inevitable in any Precambrian succession, uncertainties in correlation arise and a few of the rocks included with this formation may in fact belong to some other stratigraphic unit(s).

Total thicknesses of the formation are not readily obtainable owing to the structural complexity of the rocks and the lack of suitable marker beds, but estimates of minimum thicknesses are probably of the right order of magnitude. In the main ore zone, west of Swampy Bay River, direct measurements of several partial sections gave thicknesses ranging from 400 to 600 feet and by combining sections a thickness of 800 feet was obtained. This latter figure approaches the thicknesses of 1,000 feet (Dufresne, 1952, p. 74) and 1,200 feet (Harrison, 1952, p. 8) previously given for this formation in the main ore zone. Eastward the formation appears to thicken. Near Tait Lake where it includes a 1,000-foot thickness of volcanic rocks, it is estimated to be at least 2,000 to 2,500 feet thick; near Hurst Lake, a minimum of 2,000 feet.

### *Lithology*

#### *Sedimentary Rocks*

The lithologies of the sedimentary rocks in three broad regions show distinctions that can best be contrasted by discussing each region separately. The lithologies undoubtedly merge into one another or overlap in intermediate areas.

*West of Swampy Bay River and Wakuach Lake.* Shales and slates are the dominant lithology. In the main ore zone thin-bedded, red, green, light greenish grey and black shales are all common, but the coloured varieties predominate in the upper part of the formation. Several beds of dolomite, 1 foot to 8 feet thick, are interbedded with the shales over a wide stratigraphic range. In one place, thick-bedded, impure quartzites about 100 feet thick were found in the shales. On the west side of Wakuach Lake sparsely scattered outcrops are mainly dark carbonaceous slates and subordinate siltstones.

*Northwest of Attikamagen Lake.* The sedimentary rocks are composed of dark grey to black shale or slate and greywacke in about equal parts. The greywackes are generally fine grained, but commonly contain numerous chips of black shale ranging from microscopic to 1 cm or more in length. They are generally thickly bedded, but in most places are nevertheless intimately interbedded with shales. Graded bedding was not observed. Thin sections of typical specimens of greywacke tend to be dominated by angular rock fragments. These are mainly carbonaceous shale, but a few are basic volcanic rocks, recognizable by their plagioclase laths, and many more are indeterminate fragments of chloritic material. Chert fragments are less common. Quartz grains are widely scattered through the rock and do not exceed 10 per cent of it. They are subrounded to rounded in marked contrast with the angular but rather indistinctly defined grains and fragments that make up the enclosing matrix. Altered plagioclase, chlorite, carbonate, iron oxide, carbon, biotite, and sericite, as well as much indeterminate material compose the latter. Figure 5A is typical.

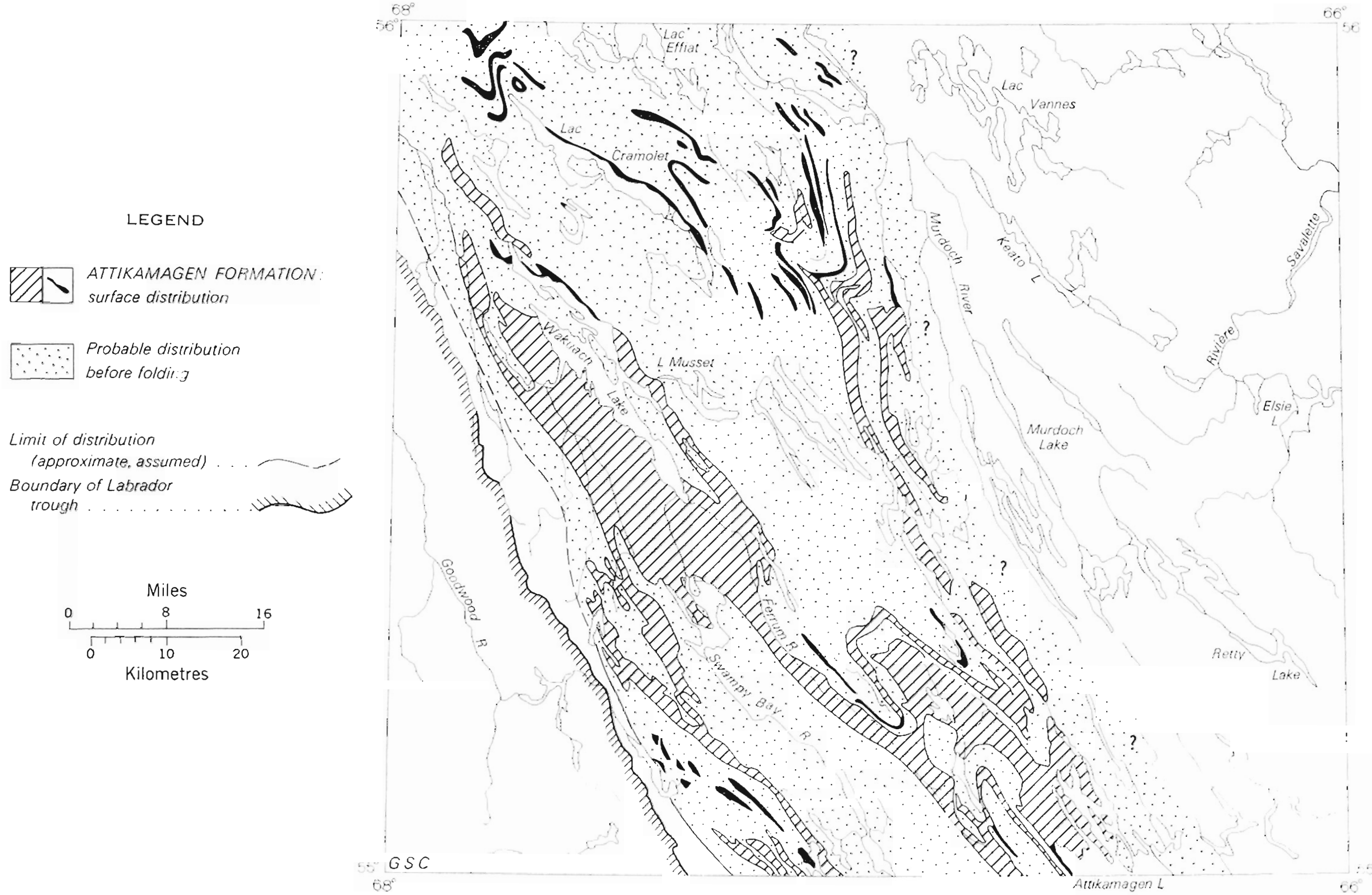



FIGURE 4. Distribution of the Attikamagen Formation in Wakuach Lake map-area.

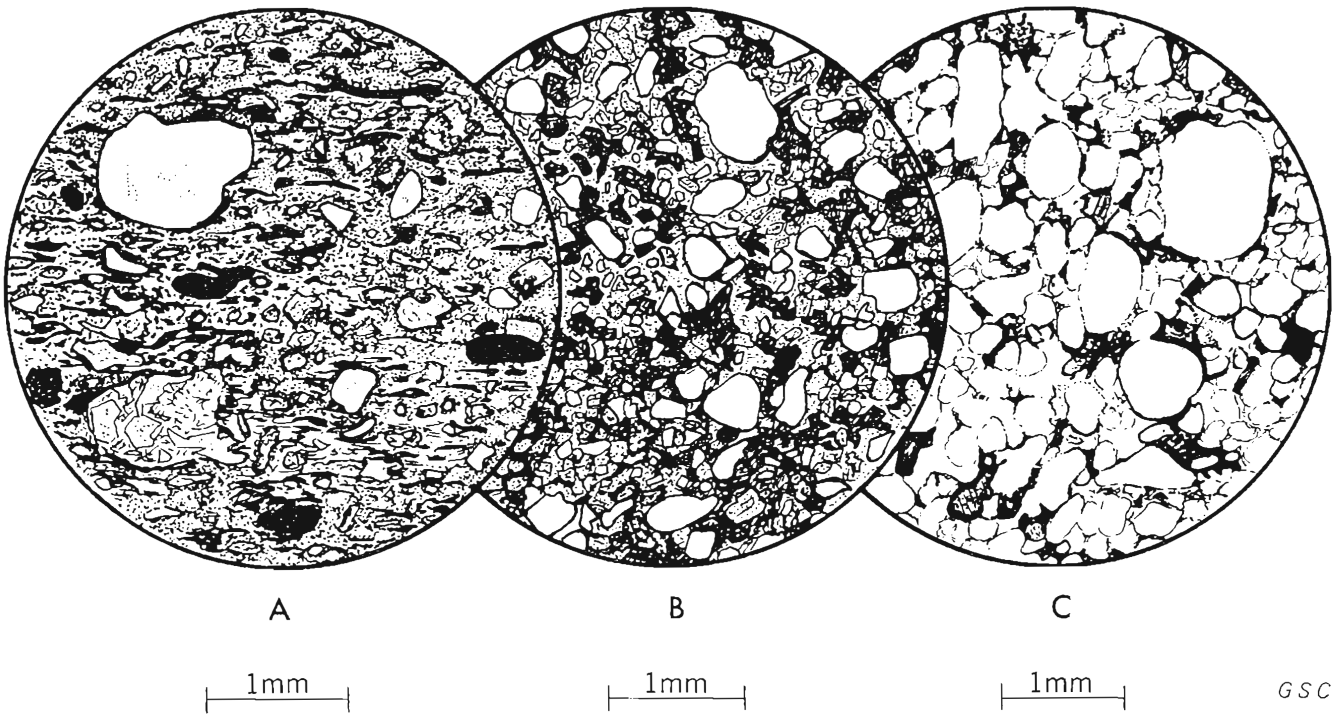



FIGURE 5. Greywackes and subgreywackes of the Attikamagen Formation.

A. Greywacke, composed mainly of lithic fragments (carbonaceous shales, chloritic material) and a largely indeterminate matrix, contains scattered, subrounded to rounded quartz grains. Feldspar in various stages of decomposition blends with the matrix. Bedding is marked by carbonaceous streaks. Specimen BL9-19.

B. Greywacke composed of subrounded to rounded quartz grains in an indistinct matrix of biotite, chlorite, altered feldspar, and indeterminate material. Specimen R140.

C. Subgreywacke consisting mainly of rounded quartz grains in a sparse matrix of altered feldspar, chlorite, and dusty iron oxide. Specimen B316.

*West of the Hurst Lake–Lac Chassin Valley.* In this region Attikamagen sedimentary rocks are predominantly slates and quartz-rich greywackes, subgreywackes, and to lesser extent impure quartzites. All but the impure quartzites are jet black in colour. The greywackes and subgreywackes have a distinctive appearance characterized by a sooty matrix thickly dotted with glassy quartz. Their bedding is in general fairly thick, commonly ranging from 6 inches to 4 or 5 feet, and they are intimately interbedded with slates. No graded bedding was observed. Under the microscope the greywackes and subgreywackes appear essentially as quartzites with a matrix that ranges from about 15 to 50 volumetric per cent of the rock. The quartz grains are generally subrounded, but commonly have serrated edges, presumably due to irregular quartz overgrowths. The matrix is composed of angular grains of altered plagioclase, minor quartz, and varying amounts of chlorite, sericite, carbonate, biotite, iron oxide, carbon, and sphene. Fragments of shale are not uncommon. The quartz grains appear sharply defined and thus dominate the appearance of the rock in thin section whereas the components of the matrix tend to blend into one another and provide an indistinct background. With a reduction in the percentage of matrix material the rocks grade into quartzites. Figures 5B and 5C illustrate typical specimens of greywacke and quartzite from this region.

Greywacke and subgreywacke similar to the ones of this region are also found in a thin strip along the east shore of Wakuach Lake.

*Quartzites.* A unit of predominantly impure quartzite or subgreywacke (shown as map-unit 7a on the Wakuach Lake sheet) found along the east side of Wakuach Lake is included with the Attikamagen Formation. It is continuous for a strike distance of about 25 miles and reaches an apparent maximum thickness of about 4,000 feet. Massive thick-bedded, grey quartzite with minor interbedded black shale is the characteristic rock type. Near the base of the unit the proportion of interbedded shale and greywacke increases, and the unit grades into shales and greywackes more typical of the Attikamagen Formation. In one place the grey quartzites exhibit pronounced graded bedding. About half a mile east of the east arm of Wakuach Lake, 8 miles from its south end, at least five successive beds ranging from 2 to 4 feet in thickness are distinctly graded from fine-grained conglomerate (2 to 5 mm) at the base to siltstones at the top. This is the only known graded bedding in the Attikamagen Formation in this map-area and one of the few occurrences in the entire Trough succession.

These grey quartzites are composed of about 70 to 80 per cent quartz in subrounded to rounded grains set in a matrix of sericite, chlorite, biotite, iron oxide, and indeterminate fine-grained material. Plagioclase grains are scarce. Rare scattered chips of shale and siltstone are visible in places. These quartzites are lithologically similar to the small quartzite units found interbedded with known Attikamagen Formation, and might reasonably be thought to represent a localization of this phase of Attikamagen sedimentation. They have therefore been grouped with the Attikamagen Formation.

*Mixed Volcanic–Sedimentary Rocks*

A single outcrop area of mixed volcanic–sedimentary fragmental rocks interbedded with greywacke and slates of the Attikamagen Formation was found at a point about 2½ miles west of the north end of Hurst Lake. The rock is composed. of blocks and slabs of greywacke and shale as much as 3 feet long in a confused matrix of shale, greywacke, and volcanic debris. The unit is at least 150 feet thick. A thin section of a matrix specimen shows a number of indistinctly outlined blebs of finely crystalline basic lava with scattered plagioclase phenocrysts in an indeterminate groundmass. Under crossed nicols the groundmass is essentially opaque except for irregular dusty areas that yield hazy pinpoints of light. Possibly the groundmass was at first glassy and may now be, in large part, finely crystalline chlorite or serpentine that has very low birefringence.

*Volcanic Rocks*

Volcanic rocks occur at several levels within the Attikamagen Formation, but the major volcanic member is evidently near the base of the formation. Near Tait Lake it is doubtful whether more than 200 or 300 feet of sedimentary rock intervene between the volcanic member and the Seward Formation. The volcanic rocks are mainly basaltic lavas. They are characteristically brown weathering, light grey aphanitic rocks. Pillows are found at several stratigraphic levels, but in those sections observed probably constitute less than one-third of the mass of volcanic rocks present. Pyroclastic rocks are minor and most of those seen occur at the base of the main volcanic member, most notably near Tait Lake and southeast of Lac Moranbert. These range from lapilli to coarse agglomerates and/or breccias.

In thin section the aphanitic basalts present a most distinctive appearance. An extremely fine grained, 'fuzzy' brownish groundmass is crowded with forked and hollow plagioclase microlites each about 0.2 to 0.5 mm long (Fig. 6A). The groundmass seems to owe its fuzzy appearance to a ubiquitous cloud of brownish dusty particles, possibly sphene or leucoxene. Where it is relatively clear, the groundmass is composed mainly of tiny rods, or fan-like clusters of pyroxene, commonly extending from the plagioclase microlites as if these had formed a nucleus for crystallization. Most sections contain from about 1 to 5 per cent microphenocrysts of plagioclase and pseudomorphs of chlorite (or serpentine) after olivine and pyroxene. Coarser grained specimens are commonly subophitic. Pyroxene crystals, which in the aphanitic basalts line the plagioclase microlites or protrude from them in tufts, appear to have merged in the coarser grained rocks into single large crystals, which surround or partly surround plagioclase laths (Fig. 6B). In places the intermediate stage can be seen; plagioclase crystals are surrounded by continuously adjoining discrete pyroxene crystals each with a slightly different optical orientation. Where plagioclase can be determined it is generally labradorite, but in a few sections it is albite. The albite basalts are more likely the products of local alteration or metamorphism than that they are spilites. Augite is the only pyroxene recognized, but most sections were not examined exhaustively and the presence of pigeonite is probable.

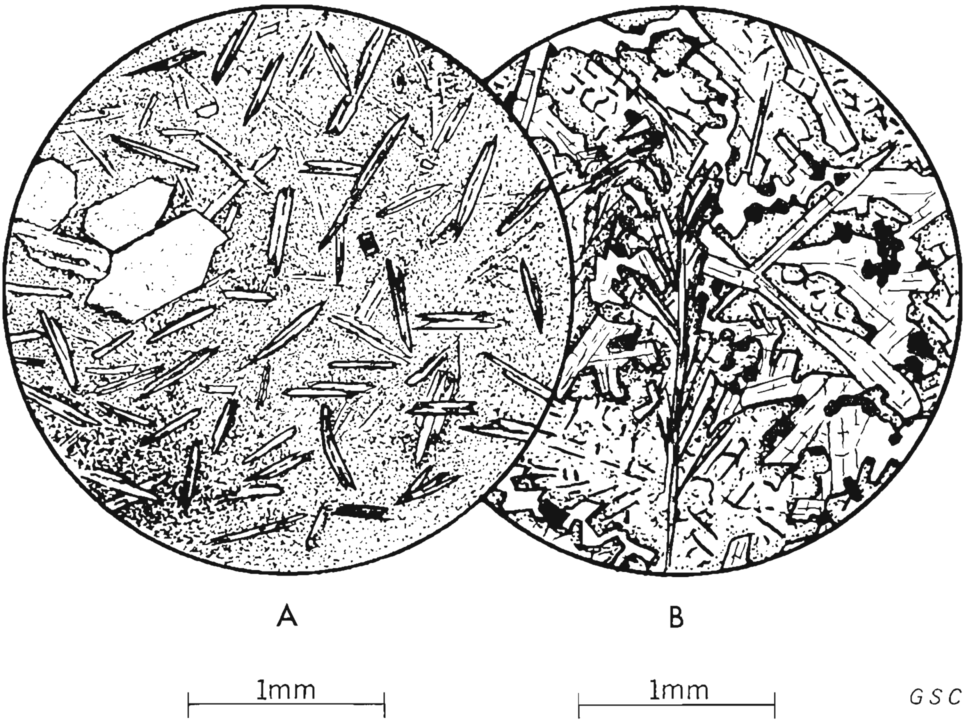



FIGURE 6. Basalts of the Attikamagen Formation.

A. Aphanitic basalt composed of a very fine grained, indistinct groundmass crowded with forked and hollow plagioclase microlites. The cluster of microphenocrysts contains serpentine pseudomorphs after pyroxene and/or olivine and a crystal of altered plagioclase. Specimen BL9-76.

B. Coarse-grained basalts consisting mostly of intergrown augite and plagioclase. Clusters of sphene and opaque iron oxide tend to rim the augite crystals. Specimen BL9-21.

*Internal Structural Relations*

Small-scale folds (i.e. with interaxial distances several hundred feet or less) are found in most places where bedding is sufficiently well defined to mark the structure; presumably they are characteristic of much of the formation. The small internal folds reflect the attitudes of neighbouring major folds and are therefore of dependent type. Slaty cleavage is most common in the shales but is by no means ubiquitous. Much of the greywacke or siltstone is all but free of cleavage.

*External Structural Relations*

The Attikamagen Formation grades upward from shales or greywackes through dolomitic shales to shaly dolomites and dolomites of the overlying Denault Formation. This is well illustrated just west of a small lake 2 miles north of the north tip of Annabel Lake, where red and green shales of the Attikamagen Formation are separated from Denault dolomite by an 85-foot thickness of light green, highly fissile, dolomitic shales. Dufresne (1952, p. 74) and Sauvé (1953, p. 14) found similar relationships in the main ore zone and in the Astray Lake area, respectively.

### *Mode of Origin*

The contrast noted between the quartz and matrix components of greywackes in the eastern regions is indicative of a mixed origin. The quartz grains are generally subrounded to rounded whereas the matrix material and lithic fragments are mainly sharply angular. Moreover the matrix material (composed as it is of a high proportion of ferromagnesian minerals, basic volcanic fragments, and chips of shale and siltstone) is not entirely consistent with a source that produces quartz in considerable abundance. On a larger scale, the contrast in materials is represented by interbedded quartzites and dark shales and greywackes.

At least part, and probably much, of the greywacke matrix is of volcanic derivation and a local source for this is readily apparent in the associated volcanic members. Similarly the shale and greywacke chips must be of local derivation. On the other hand the subrounded to rounded quartz grains, fairly uniform in size, have evidently been subjected to considerable transportation and a distant source is indicated. Since the Attikamagen Formation follows directly the richly quartzitic Seward Formation it seems reasonable to suggest that the same source continued to supply quartz to the sedimentary basin during Attikamagen times. Thus the various rock types of the Attikamagen Formation can be explained by a shifting balance between local and distant supplies.

The almost complete lack of graded bedding in the greywackes of this region, in view of its prevalence in similar environments elsewhere in the Canadian Shield is, perhaps, noteworthy. It probably denotes conditions unfavourable for the formation of turbidity currents, possibly an unusually slow rate of sediment accumulation and/or a paucity of periodic tectonic disturbances.

### *Age and Correlation*

The age of the Attikamagen Formation, like that of other formations of the Labrador Trough, can only be dated in detail by reference to its stratigraphic position in the Trough assemblage (*see* Table of Formations). In common with other members of the assemblage it is Lower Proterozoic.

The Attikamagen Formation of this area has been unequivocally correlated with that of the classical stratigraphy of the Knob Lake section by the work of Dufresne (1952), Kirkland (1950), and others.

## Denault Formation

### *Distribution and Thickness*

The present outcrop pattern of the Denault Formation together with an interpretation of its distribution before folding is given on Figure 7. It is noteworthy that, like the Attikamagen, its western boundary is well east of the present western boundary of the Labrador Trough. In the northern part of the map-area, the interpretation rests partly on the presence of a pre-iron-formation dolomite in Cambrian Lake map-area to the northwest (Fahrig, 1956; Roscoe, 1957).

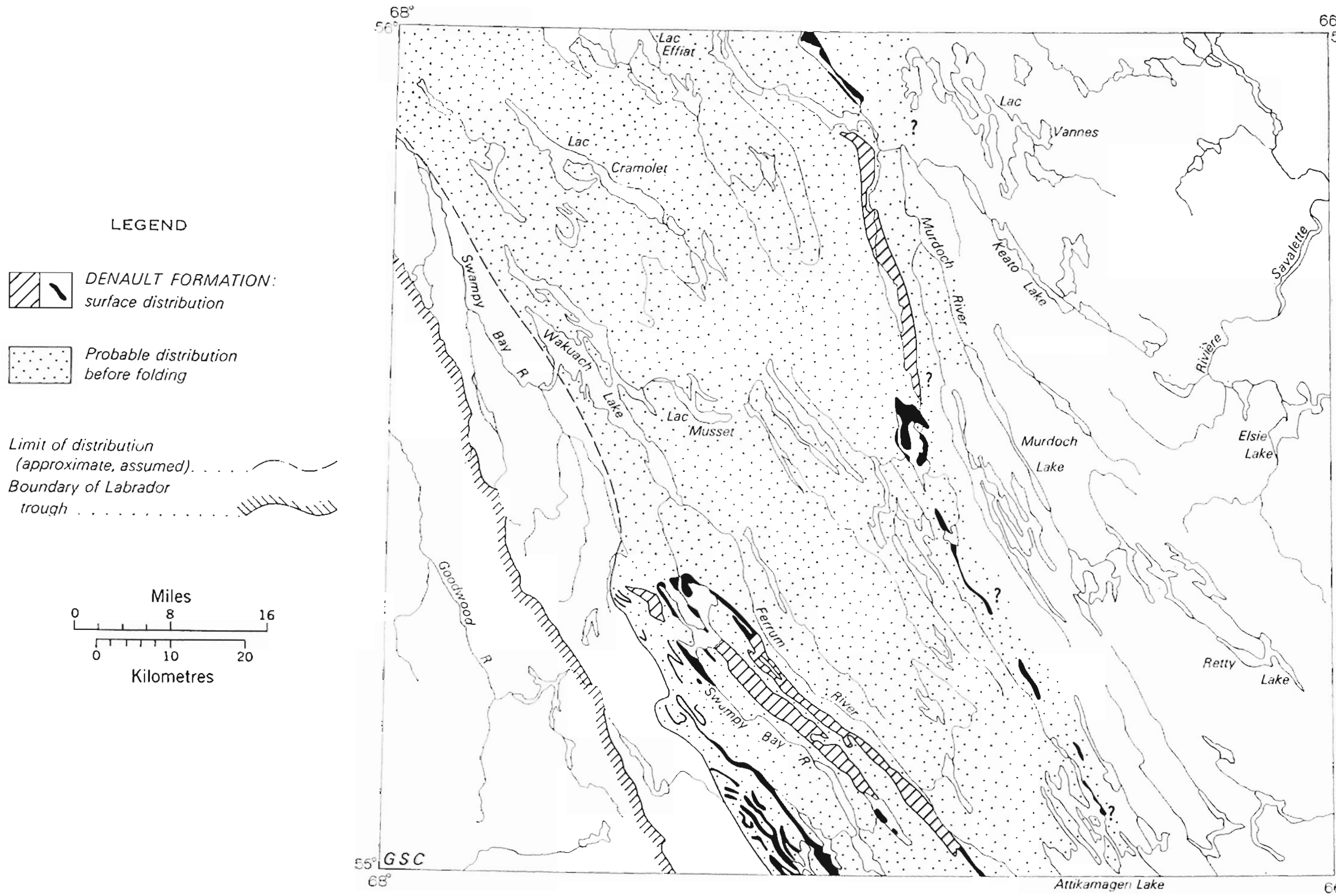



FIGURE 7. Distribution of the Denault Formation in Wakuach Lake map-area.

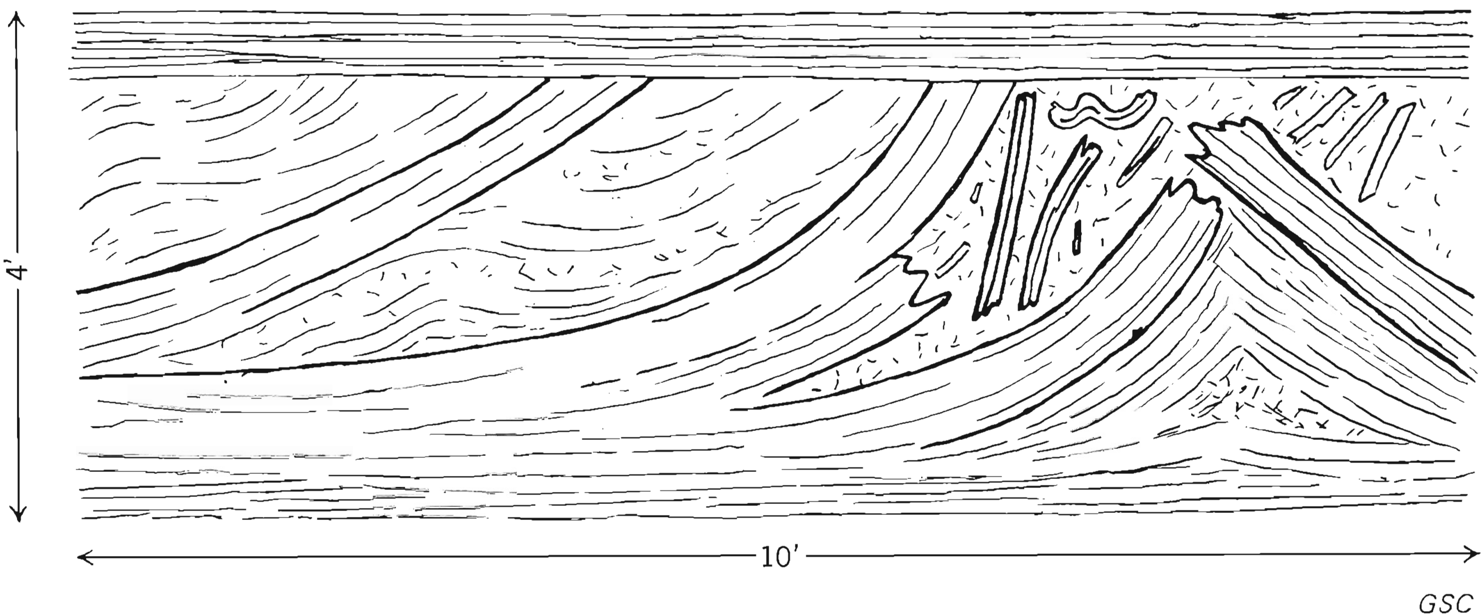



FIGURE 8. Intraformational deformation structures in the Denault dolomite. From a field sketch made about a mile east of Lac Chassin.

The thickness of the formation varies widely. Dufresne (1952, p. 26) gave a range of thicknesses in the main ore zone of 0 to 400 feet. The writer observed one section about 2 miles north of Annabel Lake that has a thickness of about 400 feet. Kirkland (1950, p. 34) reported a range of thicknesses in the area northwest of Attikamagen Lake of 0 to a minimum of 250 feet. Just east of Lac Chassin the Denault Formation reaches a thickness of about 4,200 feet.

*Lithology*

The Denault Formation is predominantly light grey to grey, buff-weathering dolomite. It is generally sufficiently finely crystalline that a fresh surface appears aphanitic. In some places the dolomite contains minor sand and in others it is slightly argillaceous and slaty. Dark grey chert in thin-bedded layers and lenses and as nodules and veins is a very common associate and locally constitutes as much as 20 to 25 per cent of the rock. The bedding ranges from a fraction of an inch to several feet in thickness and in a few places is crossbedded. Harrison (1952, p. 8) reported grain gradations in the Denault Formation of the Burnt Creek strip.

Structures that are indicative of deformation contemporaneous with deposition are widespread throughout the formation and are one of its most characteristic features. Commonly these consist of highly contorted or fractured and tilted beds confined to bedding layers from about 1 foot to 6 feet thick. Underlying beds may be involved in the contortions with diminishing effect downward, but undisturbed overlying beds sharply truncate the contorted strata (Fig. 8).

Algal structures[1] are common in the eastern belt of dolomite, particularly between Hurst Lake and Lac Low. These range from squat, bulbous forms not more than a few inches high (Pl. VI) to columnar structures several feet high (Pl. VII). The structures are circular in sections parallel with the bedding and are invariably convex upwards. At the place shown on Plate VII, colonies of closely crowded algal columns extend for at least a few tens and possibly hundreds of feet along strike and are confined to a limited stratigraphic range.

*External Structural Relations*

The Denault Formation is in contact above with both the Fleming and Wishart Formations. The Fleming Formation is of very limited extent in this map-area so that in most places the Denault Formation is directly overlain by Wishart quartzites. The contact is probably a disconformity in part (Kirkland, 1950, p. 26; Harrison, 1952, p. 7; Dufresne, 1952, pp. 51–54), but whether the disconformity is to be placed at the top of the Denault Formation or the base of the Wishart Formation is a problem that will be discussed more fully later in this

[1] A detailed account of stromatolites in the Denault Formation of the Marion Lake area by J. A. Donaldson (1963) has recently appeared. According to the classification used by Donaldson (p. 7) the columnar structures shown in Plate VII are also "Digitate" stromatolites, although considerably larger than those in Marion Lake area, and the squat bulbous structures (Pl. VI) are "Hemispherical" stromatolites.

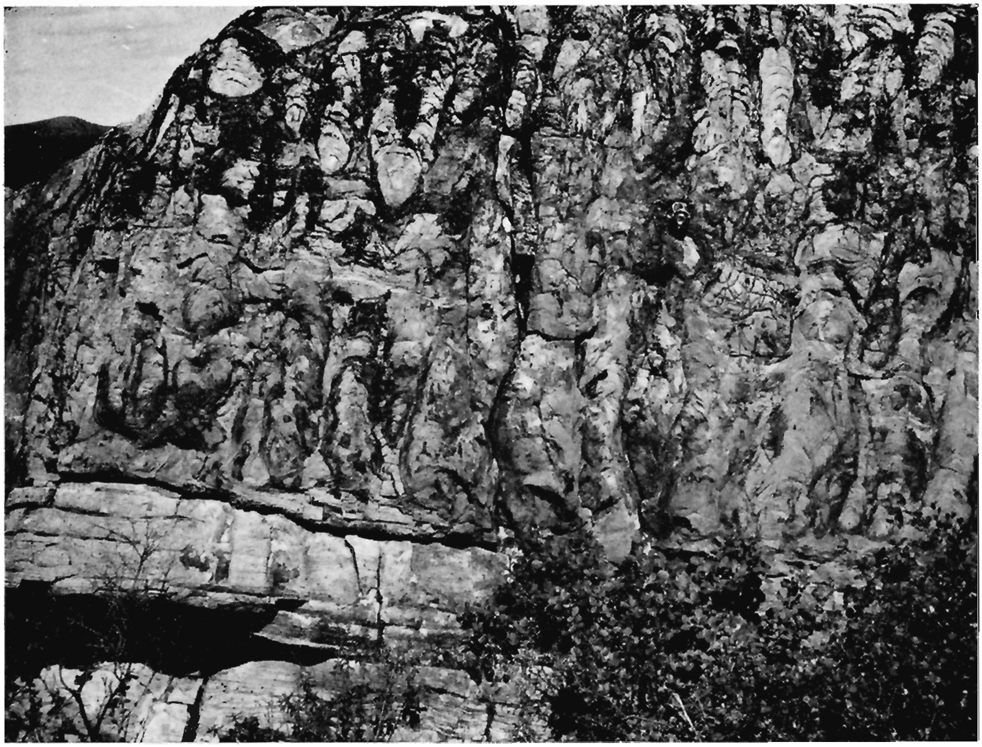
W.R.A.B. 2-2-54

PLATE VI. Squat, bulbous algal structures, Denault Formation, Lac Chassin.

report under the heading Fleming Formation. Regardless of where the disconformity is placed it may be limited geographically to the western part of the Trough. East of the south end of Bacchus Lake, dolomite is interbedded with Wishart quartzite a few feet above the upper contact of the Denault Formation. Thus, there at least these two formations appear to be conformable.

### *Mode of Origin*

The wide variability in thickness, the prevalence of disrupted bedding structures amongst undisturbed strata, and the evident clastic nature of at least part of the formation (grain gradation and crossbedding) all suggest deposition during a period of some tectonic disturbance. On the other hand, extensive algal structures of the eastern outcrop belt denote episodes of reef-formation under presumably quiet, shallow-water conditions. Possibly periodic differential sinking with local uplift resulted in widespread slumping throughout the basin, and provided clastic materials for redeposition. The presence of minor sand and silt in the dolomite formation suggests that some clastic sediment was still being contributed from outside the sedimentary basin.



W.R.A.B. 2-3-59

PLATE VII. Columnar algal colonies, Denault Formation near Lac Low. Note bifurcating columns.

## Fleming Formation

### *Distribution and Thickness*

The approximate distribution of the Fleming Formation, both at present and prior to folding, is shown on Figure 9. The formation has not been separated on the coloured geological map accompanying this report because in Wakuach Lake map-area it outcrops in only a few places. Its original distribution may not have been the isolated patches shown on Figure 9, for these are based on the presence or absence of the formation at the present surface; the formation could have had interconnections at levels not now exposed. According to Dufresne (1952, p. 35), the Fleming Formation in the main ore zone ranges in thickness from a few feet to more than 400 feet.

### *Lithology*

The Fleming Formation, where seen by the writer, consists of angular fragments of light to dark grey chert thickly packed in a matrix of dark chert, dense cherty quartzite, or quartzite. The fragments range from a millimetre to several inches across and are composed of thinly layered or massive chert. In places the fragments are platy or tabular, as if they were parts of a broken layer or bed.

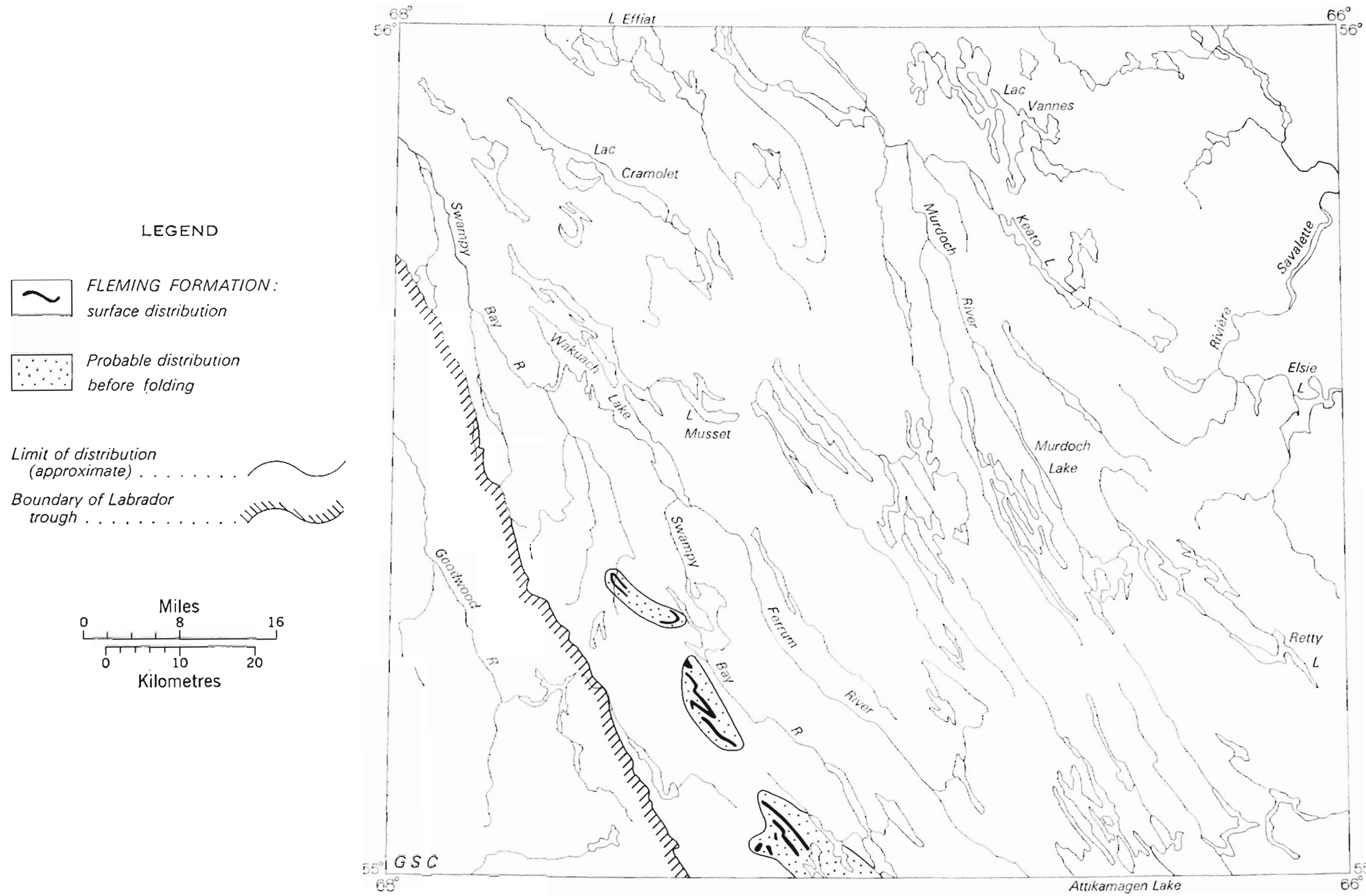



FIGURE 9. Distribution of the Fleming Formation in Wakuach Lake map-area.

In one place at or near its base, the formation contains a layer of thinly bedded, greenish grey chert or siliceous argillite. The contact relationships were not observed by the writer. However, Harrison (1952, p. 9) reported that in the Burnt Creek strip a band of slate 1 foot to 4 feet thick separates chert breccia from underlying dolomite, and Dufresne (1952, p. 54) described a section southeast of Burnt Creek where dolomite grades through massive beds of siliceous argillite into the Fleming Formation.

### *Mode of Origin*

The mode of origin postulated for the Fleming Formation determines also the interrelationships of the Denault, Fleming, and Wishart Formations. Early geologists of the Iron Ore Company held that the Fleming Formation was formed of the debris resulting from prolonged erosion of the cherty Denault dolomite (Harrison, 1952, p. 9). Therefore, a disconformity was considered to exist at the top of the Denault Formation. Dufresne (1952, p. 54) proposed that the Fleming chert breccia represents the deformation of a newly deposited chert member due to tilting of the sea-bottom prior to its emergence and subsequent erosion. According to this view the disconformity should be placed at the base of the Wishart Formation. Evidence cited by Dufresne includes: (a) the angularity of chert fragments composing the Fleming Formation; (b) the presence of local conglomerates containing chert pebbles at the base of the Wishart Formation; and (c) an exposure near the base of the Fleming Formation showing an upward gradation from siliceous argillite through highly crenulated thin chert beds to fragmented chert layers embedded in chert. The evidence is all but compelling.

## Wishart Formation

### *Distribution and Thickness*

The present and interpreted pre-folding distribution of the Wishart Formation is shown on Figure 10. The Wishart Formation overlaps all earlier formations and is the first in the succession to appear at the western edge of the Trough.

The thickness of the Wishart Formation ranges from 100 to about 300 feet. Figure 10 shows approximate thicknesses at a number of localities. Dufresne (1952, p. 36), whose map-area encompasses the main ore zone as far north as the vicinity of Harris Lake, reported the thickness of the Wishart Formation on the western and northern edges of his area as 100 to 120 feet. Although the thickest section of Wishart quartzite noted is in the easternmost zone there is little evidence of a progressive thickening in this direction. Harrison (1952, p. 9), however, noted that in the Burnt Creek strip the Wishart Formation thickens gradually eastward, from 80 to 100 feet, where it rests on gneisses at the west side of the Trough, to a maximum of 160 feet 3 miles northeast of Knob Lake.

### *Lithology*

The Wishart Formation becomes darker and apparently more mafic in composition to the southeast and eastward. The contrast is emphasized by describing the formation in three general areas: the main ore zone, Lac Hayot, and the eastern belt.

*Main Ore Zone*

The Wishart Formation is predominantly light-weathering grey quartzite ranging from thick- (1 foot or more) to thin- (1 inch or less) bedded. In places the thin-bedded members grade into grey siltstones or shale with marked bedding fissility. A rock highly typical of the formation is a thin-bedded, poorly fissile, light yellowish grey quartzite in which bedding is marked by thin (less than 2 mm) discontinuous seams of dark argillaceous material. These commonly diverge along strike to give the rock uneven or wavy bedding. Probably this is in part a poorly represented form of crossbedding. Well-defined crossbedding is visible in a few places. Local conglomeratic arkose beds a few inches to about 3 feet thick occur at a number of stratigraphic levels. Massive black chert ranging from 5 to 55 feet in thickness caps the formation at all localities visited in the main ore zone and at Ritchie Lake. In detail the stratigraphy is highly variable and few generalizations can be made. Commonly thin-bedded quartzite predominates in the upper part of the formation whereas the middle is thick bedded or massive. Both thin-bedded siltstones and thick-bedded quartzites are found at the base of the formation and in one place the base was occupied by a thin bed of conglomeratic arkose. Dufresne (1952, p. 38) reported that coarse arkosic grit a few inches to a few feet thick generally occurs at the base of the formation and noted one locality where the basal member is a conglomerate containing pebbles of chert and dolomite. Typical stratigraphic sections of Wishart quartzite from a number of widely separated localities are given on Figure 11 and their locations are shown on Figure 10.

Microscopic study shows that rocks of the Wishart Formation are predominantly quartzites, but range in composition to both arkoses and subgreywackes. Feldspathic quartzite is a common type. Subgreywackes are generally confined to the thin-bedded, fine-grained quartzite or siltstone phases. Characteristic compositions may fall anywhere within the following percentage ranges:

| | |
|---|---|
| quartz | 60–95 per cent |
| feldspar (mainly K-feldspar) | 5–40 per cent |
| chert | 1– 5 per cent |
| matrix | 1–25 per cent |

Matrix material is mainly fine-grained chlorite, sericite, biotite, dusty opaque minerals, and locally goethite and hematite. It appears as clots, thin selvedges

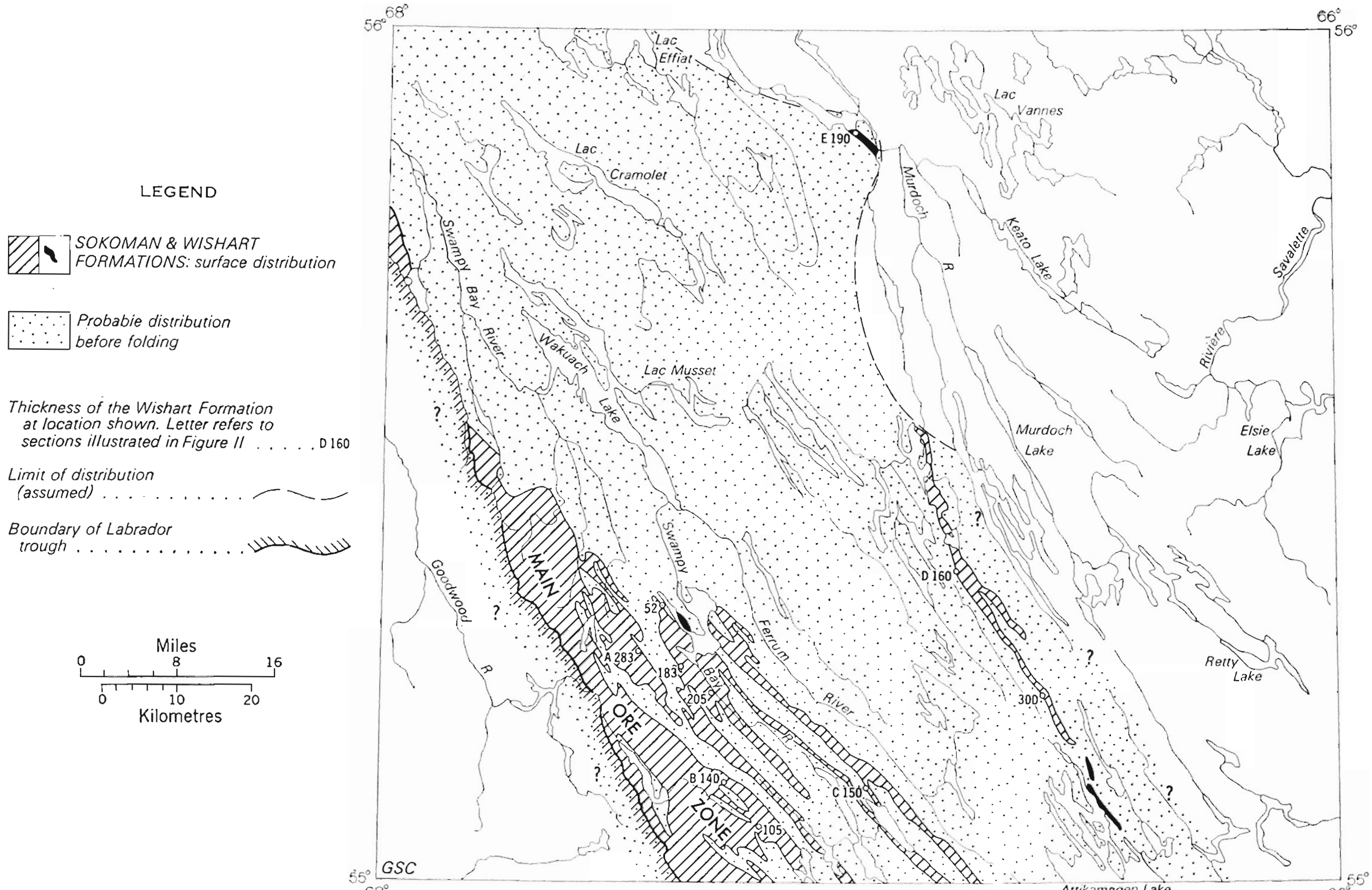



FIGURE 10. Distribution of the Sokoman and Wishart Formations in Wakuach Lake map-area showing thicknesses of the Wishart Formation at a number of localities.

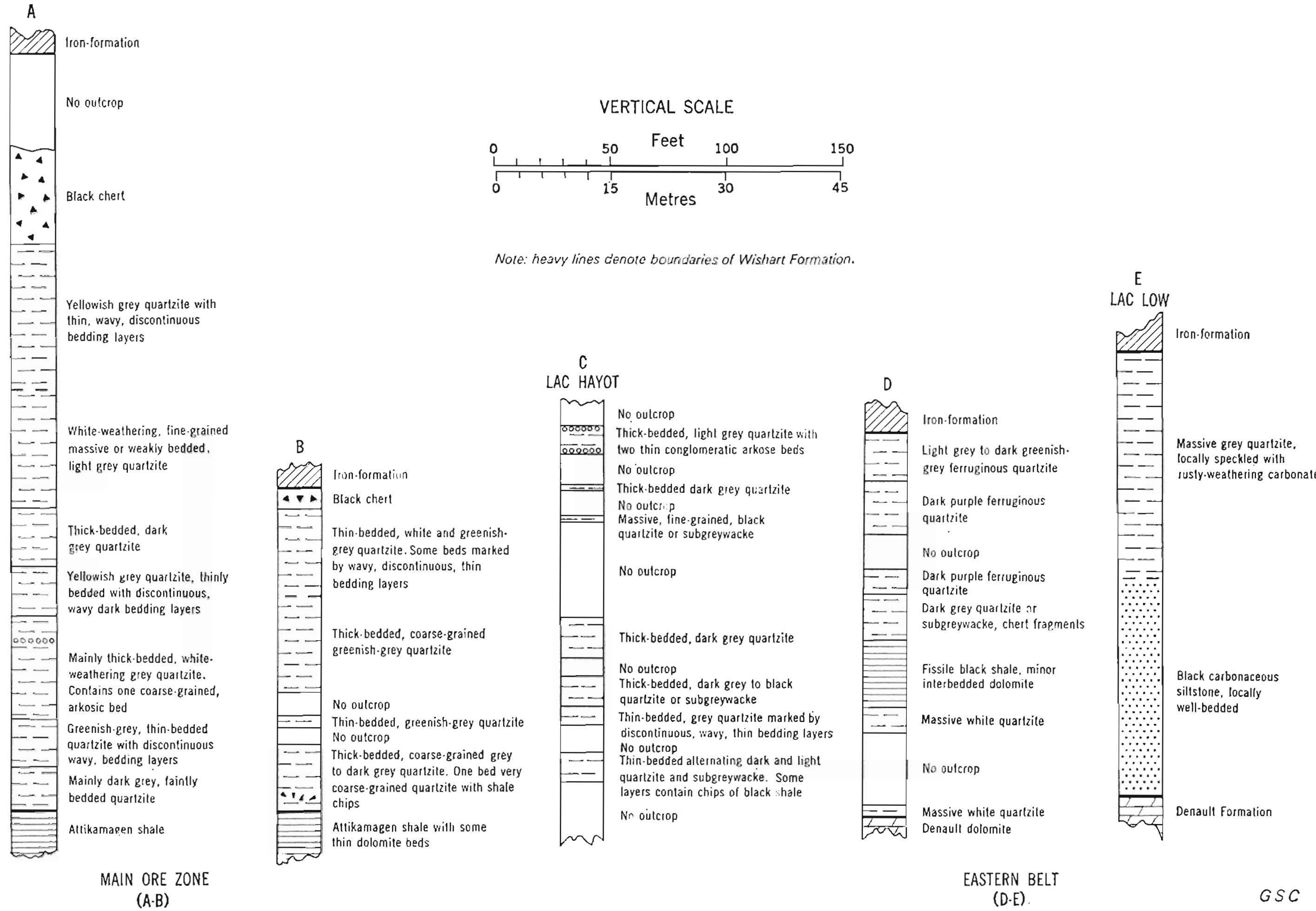



FIGURE 11. Stratigraphic sections of the Wishart Formation, Wakuach Lake map-area.

about grains, interstitial filling, and discontinuous, hair-like seams. The amount of matrix material varies from bed to bed, but is generally more abundant in the finer grained beds. In a few places dark quartzite layers contain as much as 10 to 15 per cent of iron oxides. Quartzite grains are generally subrounded to rounded, but in many specimens this is partly obscured by quartz overgrowths. The grain size commonly ranges from 0.1 to 0.5 mm. Minor chert is ubiquitous and occurs both as grains and locally as cementing material. Figure 12A shows a typical specimen of Wishart quartzite as seen in thin section.

Thin-bedded siltstones of the Wishart Formation are invariably more mafic than the associated quartzites. Characteristic compositions may be represented as follows:

| | |
|---|---|
| quartz | 30–50 per cent |
| feldspar (K- feldspar dominant) | 5–30 per cent |
| chert | 1– 3 per cent |
| matrix | 40–60 per cent |

The matrix material is similar to that of the quartzites. Grains are generally angular to subangular and range from about 0.01 to 0.06 mm.

The chert layer that caps the formation is composed of tightly packed, clouded pellets of cryptocrystalline quartz, 0.3 to 0.5 mm in diameter, in a clear, slightly coarser grained matrix of the same material. The contact with underlying quartzites is marked in one thin section by a clear, structureless zone of cryptocrystalline quartz, about 1 mm thick, that is locally continuous with the cryptocrystalline matrix of adjoining quartzite.

*Lac Hayot*

Northwest of Lac Hayot, the Wishart Formation is a mixed assemblage of light grey quartzites, similar to types in the main ore zone, and dark grey or black quartzites or subgreywackes. The latter type seems to predominate (*see* for example, section C, Figure 11). Locally the subgreywackes abound in angular chips of dark argillite and, in one place, hematite. Kirkland (1950, p. 40) noted that south of Hook Lake pure quartzite is less common than north of it and that chlorite, biotite, argillaceous material, white mica, and some feldspar, are abundant constituents of the formation. The feldspar is chiefly microcline and rarely exceeds 5 per cent of the rock. Kirkland also reported (p. 41) that the top of the formation is marked by a 3- to 20-foot-thick layer of black chert.

*Eastern Belt*

The Wishart Formation of the eastern belt also contains a preponderance of dark, mafic-rich quartzites or subgreywackes over pure quartzites (Fig. 11).

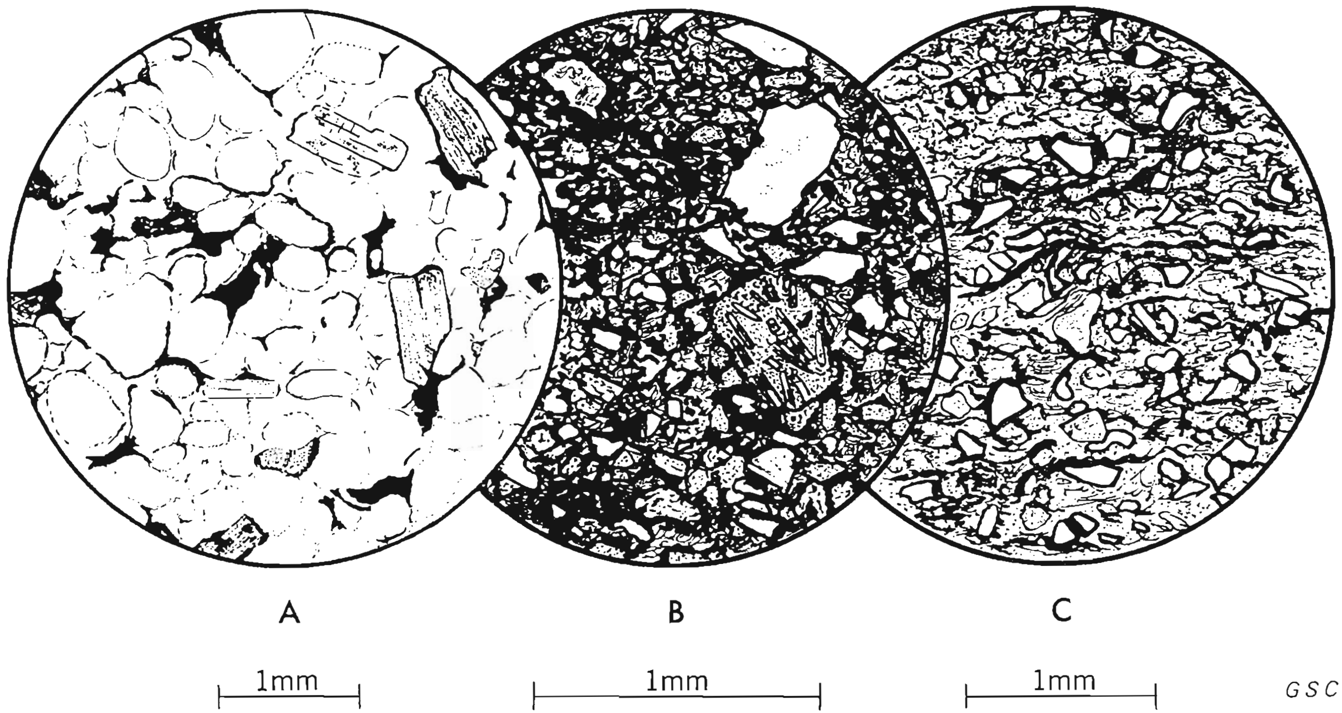



FIGURE 12. Wishart quartzite and Menihek greywackes.

A. Wishart quartzite composed mainly of rounded to well-rounded quartz grains cemented by quartz overgrowths. A few orthoclase grains clouded with dusty hematite are scattered through the section, chlorite and dusty iron oxide minerals fill some of the interstices. Specimen BL-208.

B. Menihek greywacke dominated by angular grains of quartz and feldspar, lithic fragments, and dusty, black opaque material. A fragment of volcanic rock contains microphenocrysts of plagioclase. Specimen BL-290.

C. Menihek (formerly Howse Group) greywacke composed mainly of sharply angular grains of quartz and feldspar in a finely divided, largely indeterminate, groundmass. The feldspar (both potash feldspar and plagioclase) shows varying but commonly severe, alteration, which tends to blend it with the groundmass. Bedding is marked by wisps of carbonaceous matter. Specimen B329.

Purple to grey or black ferruginous quartzite contains the following estimated amounts:

| | |
|---|---|
| quartz | 70 per cent |
| chert | 2–3 per cent |
| feldspar (mainly K–feldspar) | 5 per cent |
| goethite cement | 20 per cent |
| carbonate (ankerite?) | 3–4 per cent |

### *External Structural Relations*

The relationship of the Wishart to underlying formations was discussed under the heading Denault Formation. The Wishart Formation obviously truncates the older formations in the western part of the area for it overlaps them westward by several miles but this should not be interpreted as a disconformity of more than minor significance. Little evidence of a disconformity is found in the eastern outcrop belt where dolomite seems to be interbedded with Wishart quartzite. Similarly east of Astray Lake, 30 to 40 miles south of the present map-area, Sauvé (1953, p. 19) found evidence of interbedding between Wishart quartzites and the Attikamagen Formation. Therefore, as Sauvé pointed out, an unconformity may exist near the margin of the Trough but not in its interior (a fairly common feature of depositional basins).

The Wishart Formation is conformably and almost continuously overlain in the western part of the Trough by the Ruth Formation. In the eastern belt it is succeeded directly by the Sokoman Formation. Harrison (1952, p. 10) assigned the chert layer at the top of the Wishart quartzite to the Ruth Formation; this interpretation is supported in the eastern belt, for there, where the Ruth Formation is absent, the chert layer is also absent. Nevertheless, because of the consistent chert content of the Wishart quartzite and the continuity of the chert (cryptocrystalline) matrix between the chert layer and quartzite, the writer prefers to follow the interpretation of the Iron Ore of Canada geologists, who regard the chert as part of the Wishart Formation.

### *Mode of Origin*

The relative purity of much of the quartzite, the presence of crossbedding and ripple-marks (Harrison, 1952, p. 10), and the constant thickness of the formation denote deposition under shallow water conditions during a tectonically quiet period. The seas evidently expanded westward during this period onto the basement complex from where much of the clastic material contained in the formation was probably derived. In the southeastern and eastern parts of the map-area, an increasing amount of mafic and argillaceous constituents indicates a change in sedimentation. At least part of the change might be attributed to normal, off-shore, deeper water sedimentation but probably the greater abundance of iron and magnesian minerals contained in these rocks also denotes some admixture of volcanic material. According to Sauvé (1953, p. 42), the Nimish basic volcanic rocks near Astray Lake cover a wide stratigraphic range and are, in part, interbedded with the Wishart Formation. It would be strange if contemporary volcanism at distances of 30 to 50 miles did not contribute some material to the Wishart of this area.

## Ruth Formation

The Ruth Formation is thin, unevenly distributed, and weathers readily; accordingly it was rarely seen by the writer and much of the following account is summarized from the literature. It has not been shown separately on the map, but is included in the closely associated Sokoman Formation.

### *Distribution and Thickness*

The Ruth Formation is limited to the main belt of iron-formation on the western side of the Trough and according to detailed Iron Ore of Canada Company maps it is absent in much of this belt directly west of Annabel Lake. Dufresne (1952, p. 40) noted that the formation is as much as 100 feet thick. Just west of Lac Le Fer it is 65 feet thick.

### *Lithology*

The Ruth Formation is predominantly thinly fissile, brown to black, ferruginous slate. It contains some interbedded chert and greenish iron silicate layers. Gross (1951, pp. 12 and 35) reported that the less ferruginous members can be seen to consist of angular quartz and feldspar grains sparsely distributed through a matrix of chlorite, white mica, other platy minerals, abundant iron oxides, and finely disseminated opaque minerals. Feldspar is mainly orthoclase in contrast with the feldspar in both the Attikamagen and Menihek Formations. Differential thermal analyses led Gross (1951 p. 89) to conclude that the iron oxides were mainly the two minerals lepidocrocite and goethite, and of these probably lepidocrocite was more abundant. Rhombohedral-shaped masses of iron oxide indicate the former presence of iron-bearing carbonate (Gross, 1951, p. 35). Carbon is moderately abundant. Iron silicate layers locally contain up to 95 per cent minnesotaite (Dufresne, 1952, p. 41).

### *External Structural Relations*

Just west of Lac Le Fer, one of the few localities where outcrops of Ruth Formation were seen by the writer, it is interlayered for several feet with the silicate–carbonate iron-formation of the overlying Sokoman Formation. According to Dufresne (1952, p. 40) this relationship is common.



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### *Mode of Origin*

The presence of moderately abundant carbon together with lack of cross-bedding or ripple-marks led Gross (1951, pp. 96–97) to suggest that deposition took place in deep, quiet water where oxidizing conditions were lacking. Orthoclase may have been contributed from the same source as that which supplied the Wishart Formation (Gross, 1951, p. 36). The abundance of iron and the presence of chert marks the beginning of conditions that culminated with deposition of the iron-formation. Presumably the Ruth Formation represents a period during which the influx of clastic sediments into the Labrador Trough declined to zero.

## Sokoman Formation

The Sokoman Formation is not only of major economic importance but is the marker bed of most significance in the Trough assemblage. It has been subdivided into a number of stratigraphic units, which are sufficiently distinctive to be useful in detailed structural analyses. The extent to which structural and stratigraphic problems of the Labrador Trough have been solved is largely a result of these properties.

### *Distribution and Thickness*

The present and interpreted pre-folding distributions of the Sokoman Formation in Wakuach Lake map-area are similar to those of the Wishart Formation and are shown on Figure 10. The formation outcrops most extensively in the western part of the Labrador Trough where it appears at the surface repeatedly owing to numerous shallow folds and thrust faults. In the east-central region of the Trough it appears again on the eastern limb of the regional anticlinorium. The eastern zone (Walsh zone) is not continuous. North of Hurst Lake the Sokoman Formation is absent except for a strip a few miles long immediately south of Lac Low. South of this the Denault Formation is directly succeeded by the Menihek Formation and north of it by the Murdoch. Either non-deposition or erosion could account for the missing sections of the Sokoman Formation, but direct evidence is lacking.

The thickness of the Sokoman Formation was reported by Stubbins, *et al.* (1961, p. 44) to range from 400 to 800 feet. Only a few sections observed by the writer in Wakuach Lake map-area were sufficiently complete to afford measurements of total thickness. On the southwest shore of Lac Le Fer the formation is 750 feet thick, just south of Bruin Lake it is about 550 feet thick, and east of Gulch Lake in the eastern zone, about 800 feet. Elsewhere measurements of partial sections confirm the general range of thicknesses reported by Stubbins, *et al.* Near Lac Low the thickness of iron-formation exposed is about 110 feet, but the total thickness is probably much greater and may approach the same general thickness as elsewhere.

TABLE I

*Stratigraphy of Sokoman Iron-formation in the Schefferville Mining District*

(from Stubbins, *et al.* (1961, p. 44))

| Member (in feet) | Name of Subdivision (in feet) | Description |
|---|---|---|
| Upper member (80–200) | Lean chert (30–80)<br>Red upper (10–30)<br>Grey upper (40–120) | Thick bedded lean chert.<br>Mainly ferruginous slate.<br>Thick bedded ferruginous chert. |
| Middle member (300–500) | Upper red cherty (80–150)<br><br>Brown cherty (10–40)<br>Grey cherty (50–100)<br>Pink cherty (30–60)<br>Lower red cherty (30–60) | Jasperoid, carbonate-bearing metallic horizon.<br>Silicate-rich metallic horizon.<br>Thick bedded ferruginous chert.<br>Thin bedded ferruginous chert.<br>Thin bedded jasperoid metallic horizon. |
| Lower member (30–100) | Silicate-carbonate (30–100) | Silicate-rich horizon. |

TABLE II

*Stratigraphy of Sokoman Iron-formation in Wakuach Lake Map-area*

| Member (in feet) | Name of Subdivision (in feet) | Description |
|---|---|---|
| Upper member (50–170) | Lean cherts (50–170) | Thick bedded pink and grey chert with low content of metallic minerals. |
| Middle member (350–600) | Cherty metallic iron-formation (350–550): Jasper metallic iron-formation; Grey cherty metallic iron-formation<br>Thin bedded jaspilite (40–80) | Alternating layers of jasper and metallic-rich rock 1″–4″ thick.<br>Mainly thick layered grey chert with heavily dispersed metallics.<br>Thin layered jasper, metallic-rich, and chert rock. |
| Lower member (40–80) | Silicate-carbonate iron-formation (SCIF) (40–80) | Thin layered, olive-green, minnesotaite-rich rock. |

## *Subdivision of the Sokoman Formation*

A detailed subdivision of the Sokoman Formation in the Schefferville Mining District provided by Stubbins, *et al.* (1961, p. 44) is reproduced in

Table I. The breakdown of the middle member results from the work of Schwellnus (1957, pp. 22–31) in the Ruth Lake area near Schefferville. The writer was guided in Wakuach Lake map-area by a more generalized subdivision used by Iron Ore Company geologists in their regional mapping of the iron-formation (on the scale of 1 inch to 1,000 feet) and it proved to be entirely practical. The main units are given in Table II. Additional units noted by the company geologists but rarely observed by the writer in his more abbreviated examinations include carbonate iron-formation, magnetite greywacke, and magnetite shales. These do not appear to be persistent statigraphic units such as those noted in the table. A comparison of various sections and partial sections is summarized on Figure 13.

The stratigraphy of the iron-formation is more complex than the simple tabulation of units can convey. The boundaries of units are not necessarily well defined and units may range widely in thickness. Rock types characteristic of one unit may appear within another; thus silicate–carbonate iron-formation identical with that composing the lower member was observed interbedded with cherty metallic rocks of the middle member in several places. Parts of each unit may consist of rocks of a general character that are indistinguishable from unit to unit and are therefore stratigraphically indeterminate when observed in isolation. Nevertheless, the units are generally sufficiently distinctive to constitute useful stratigraphic marker beds.

*Silicate–Carbonate iron-formation*

The rock is generally thin bedded, fine grained, and ranges in colour from dull olive-green to dark grey with an olive cast. Bright orange-brown weathering is characteristic and differential weathering of individual layers produces a corrugated surface. In places the rock is speckled with minute crystals of magnetite.

Silicate–carbonate iron-formation is composed mainly of minnesotaite, quartz, iron carbonate[1], and opaque iron oxide or, more rarely, goethite. Stilpnomelane was observed in only one thin section. The relative abundance of each constituent varies widely from layer to layer in the same rock but commonly falls within the following percentage ranges (volumetric): minnesotaite, 50–90; quartz, 10–40; iron carbonate, 0–30; and iron oxides, 5–20. A few of the rocks are predominantly iron carbonate. In the only polished section studied the iron oxide is entirely magnetite.

Minnesotaite[2] occurs in clusters of radiating crystals each of which is from 0.05 to 0.1 mm long. Some layers are composed almost entirely of close-packed rosettes and tufts of the mineral. Other layers are predominantly fine-grained quartz and contain only isolated clusters of minnesotaite. In one thin

[1] The composition of carbonates varies considerably (Gross, 1961, pp. 31–32) in different iron-formations. Siderite, ankerite, and ferruginous dolomite are the most common. Here no specific determinations were made: hence they are referred to under the general term iron carbonates.

[2] $Si_{7.6}\ (Al,Fe^{3})_{0.35}\ (Mg,Fe^{+2},Mn)_{5.55}\ O_{18.5}\ (OH)_{5.5}$ according to Deer, *et al.* (1962, p. 124). Minnesotaite is considered to be the iron analogue of talc.

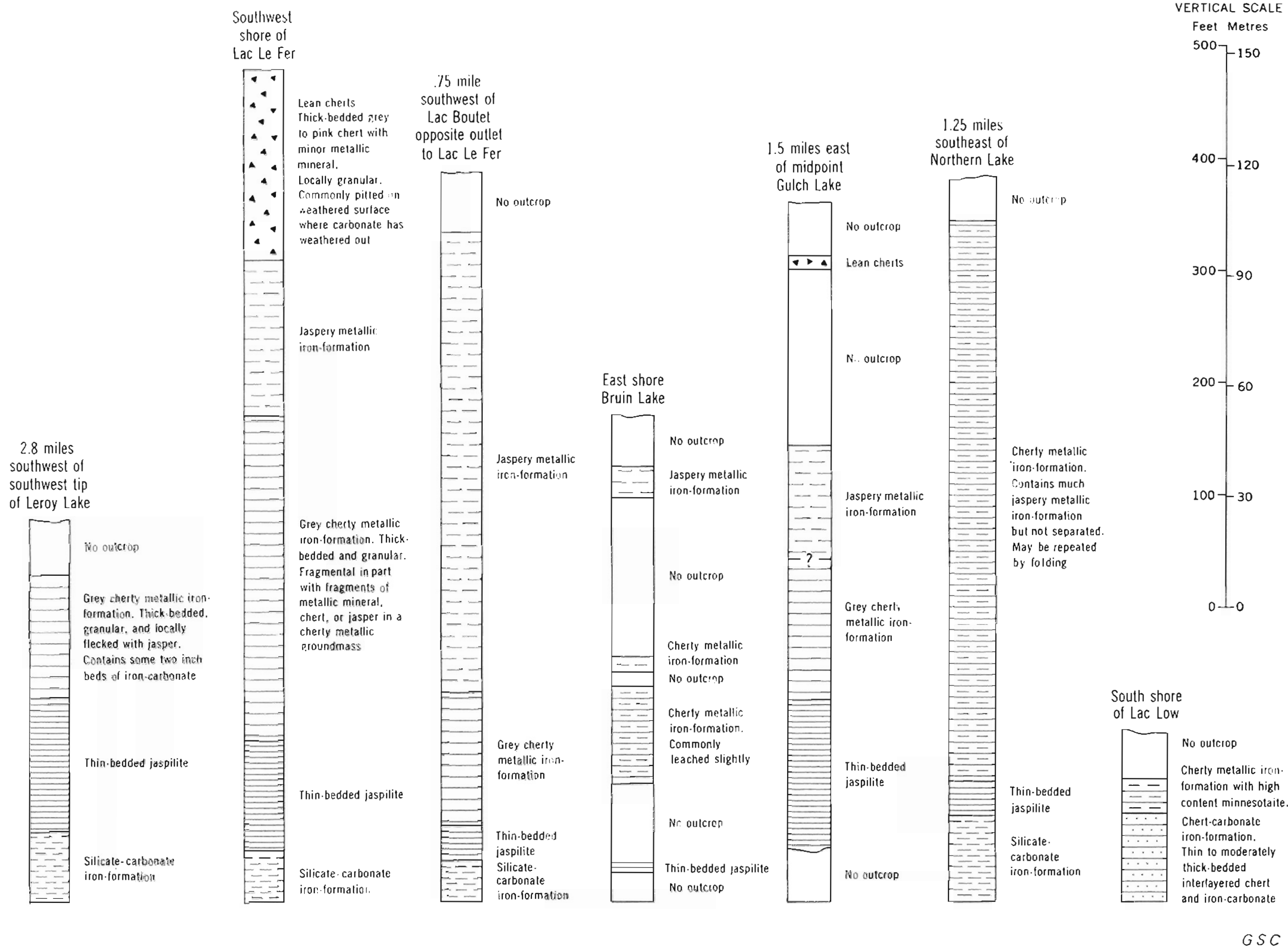



FIGURE 13. Stratigraphic sections of the Sokoman Formation.

section, fractures that cross the layering of the rock and penetrate a metallic-rich bed are seen to be thickly lined with tufts of minnesotaite regardless of the composition of the host layer. Quartz appears as a fine-grained mosaic in which individual crystals range in size from submicroscopic to 0.1 mm but most commonly are in the size range 0.005 to 0.05 mm. Magnetite (possibly martite in part) is mainly disseminated in euhedral crystals 0.01 to 0.1 mm across or in clots 0.1 to 0.5 mm in diameter. The latter may represent poorly formed granules[1]. Some layers are composed largely of metallic minerals. Carbonate is present generally in discrete irregular masses that range from 0.005 to 0.2 mm across. Granules appear to be uncommon in this unit but may occasionally be seen in chert-rich[2] layers.

Both iron carbonate and minnesotaite readily weather to goethite, but iron carbonate appears to be more susceptible and is replaced by goethite to a greater depth from the weathered surface than is minnesotaite. The metallic minerals appear to be much less affected by weathering. Silicate–carbonate iron-formation overlies the Ruth Formation conformably and in places near the contact can be seen interbedded with Ruth slates.

*Thin-bedded Jaspilite*

Thin-bedded jaspilite is composed of layers 1 mm to about 3 cm thick with varying proportions of chert and iron oxide minerals. The colour of individual layers ranges from red through brown to light or dark greys, depending on whether the dominant mineral is jasper, goethite, grey chert, or metallic iron oxides, respectively. Commonly the rock consists of alternating layers of jasper or grey chert nearly free of metallic minerals and metallic or goethite-rich layers. Layers pinch, swell, and commonly lense out along strike. The rocks are in general thinly fissile and may be much contorted locally. Goethite appears to be more abundant in this unit than elsewhere in the iron-formation, a feature that may be attributable to its greater fissility and internal deformation.

The main constituents of the thin-bedded jaspilites are metallic iron oxides, goethite (plus limonite), and quartz. Minor constituents include iron carbonate, riebeckite, white mica, and pyrite. A polished section of one of the metallic-rich specimens shows a wide variation in the relative abundances of hematite and magnetite from layer to layer. The layers richest in metallic minerals contain a preponderance of magnetite. Other layers contain thickly scattered magnetite octahedra in a ground mass of quartz and irregular blebs of hematite. One brownish submetallic layer is composed of goethite with minor magnetite in dispersed euhedral crystals.

Quartz ranges in grain size from 0.005 to 0.1 mm, but is commonly nearer the lower size limit. Magnetite crystals are generally 0.01 to 0.03 mm in diameter.

[1] Granule in this report is generally used to denote a completely metallic pellet, whereas oolite is applied to a pellet formed of concentric metallic shells separated by non-metallic minerals. Hybrid pellets make it difficult to apply the terms rigidly. Granules may also be used as a general term to describe all types of pellets as in megascopic descriptions.

[2] Chert is now invariably fine-grained quartz.

Granules can be seen in the jasper layers of some specimens, but were not observed in thin section. Thin-bedded jaspilites overlie silicate–carbonate iron-formation and pass into it fairly abruptly. Some interbedding of the two units may take place at the contact as suggested by Harrison (1952, p. 11).

*Cherty metallic iron-formation*

This unit may be divided into two parts, which, in the writer's experience, have varying qualities of distinctiveness and a rather indefinite mutual boundary. They have in common a tendency to be thick bedded or massive and to possess a metallic mineral content generally in excess of about 25 per cent by volume. The name 'cherty metallic' derives from the rock's characteristic metallic appearance. Grey cherty metallic iron-formation, the name given to the lower part of the unit, is typically a thick-bedded rock of grey, pinkish grey, or, where riebeckite-bearing, bluish grey colour. The metallic minerals are generally evenly dispersed through the rock and granular structures are common. Jaspery metallic iron-formation, the upper part of the unit, is a highly distinctive rock in its characteristic form. Bright red layers of jasper, 1 inch to 4 inches thick, alternate with grey, metallic-rich layers of similar thickness. The jasper layers pinch and swell along strike and locally separate into strings of isolated pods and lenses. Some of the jasper beds are thickly packed with granules. In addition to its characteristic types, the jaspery metallic sub-unit contains varieties that are similar to those typical of the lower sub-unit.

The major minerals of cherty metallic iron-formation are quartz, hematite, magnetite, and goethite. Minor minerals of more local distribution are riebeckite, iron carbonate, minnesotaite, stilpnomelane, and white mica. The iron oxide content of the rock ranges from about 20 to 80 per cent by volume, but is most commonly 30 to 50 per cent. Polished sections of grey cherty metallic rock generally show a complex mixture of magnetite, goethite, and hematite in varying proportions. Volumetric analyses of two such specimens, one (BL-243) from the lower sub-unit and the other (BL-251) of a similar rock type from the upper unit, are as follows:

| | BL-243 | BL-251 |
|---|---|---|
| magnetite | 23.4 | 12.8 |
| hematite | 6.4 | 1.6 |
| goethite | 24.6 | 18.4 |
| non-metallic | 45.6 | 67.2 |
| | 100.0% | 100.0% |
| Point counter analyses | 1395 points | 1475 points |

Although these specimens appear to be fairly typical of grey cherty metallic iron-formation in hand specimen, the mineral proportions may not be representative. Other specimens seem to contain mainly hematite. In typical jaspery metallic iron-formation, the metallic mineral is almost entirely hematite, and may constitute 70 per cent by volume of the metallic-rich layers.

Granular structures are common throughout the cherty metallic member, but appear to be more abundant in grey cherty metallic rock-types. They range from 0.05 to 2 mm, but are commonly in the size range 0.3 to 0.5 mm. Three main types are present: 1) chert oolites; 2) oolites comprising alternating chert and iron oxide shells; and 3) granules entirely of iron oxide. Intermediate types include granules with outlying oolitic shells and granules with quartz-filled centres. Generally one type predominates in a single specimen. Granules or oolites appear to be most abundant in rocks with small or moderate amounts of iron oxide and are rare in metallic-rich layers.

The ground mass of cherty metallic iron-formation is a fine-grained mosaic of interlocking quartz crystals that range from submicroscopic to about 0.3 mm across. The texture is generally seriate or patchy, such that patches of the finest grain size, typically 0.001 to 0.005 mm, alternate throughout the rock with patches in which the grain size increases gradationally to a maximum near the centre of the patch. An average grain size is, therefore, difficult to estimate, but a size range from 0.005 to 0.1 mm is common. Chert oolites, in places, preserve a grain size that is barely perceptible whereas surrounding quartz has a normal size range. No significant variation in grain size from place to place could be detected.

Hematite and magnetite crystals are commonly 0.005 to 0.05 mm across, but aggregates of crystals may be as much as 3 mm in diameter. Goethite and hematite replace and commonly pseudomorph magnetite crystals. Martite is particularly widespread and denotes a substantially higher initial magnetite content in cherty metallic iron-formation than is now present. The age relations between hematite and goethite are generally indecisive, but under some conditions, at least, goethite clearly replaces hematite. For example, in slightly leached specimens of jaspery metallic iron-formation, the iron mineral is predominantly goethite whereas in fresh specimens of the same unit it is predominantly hematite. Secondary magnetite was observed in one specimen where a hair-line fracture with associated coarse magnetite crystals crosses a hematite-rich layer.

Slightly leached cherty metallic iron-formation is not uncommon. In such rocks chert and jasper are converted to light buff to white, crumbly, sugary-textured quartz, and metallic minerals are replaced by dense brown goethite. The resulting rock shows a heightened contrast between layers, which commonly produces a markedly striped appearance. Boundary relationships between cherty metallic iron-formation and underlying thin-bedded jaspilite were not noted in detail, but the change from one to the other appears to be abrupt.

*Lean Cherts*[1]

Lean cherts forming the upper member of the iron-formation were seen by the writer in few places, where they were not well exposed. The following

[1] 'Lean cherts', the term applied to this unit by Iron Ore Company of Canada geologists, derives from its low iron content.

description is, therefore, far from complete. The cherts are mainly massive, grey to pink rocks containing less than 25 per cent by volume of metallic minerals and commonly much less. They are generally pitted and stained where clumps of iron carbonate have weathered out and iron has been redeposited as limonitic coatings. Granular structures are fairly common. Dufresne (1952, p. 120) observed layers of carbonate-rich chert up to 20 feet thick within the lean chert member, particularly in the western part of the Goodwood–Burnt Creek area and he also noted that black graphitic chert is common. Ferruginous slate, referred to as the red upper unit by Stubbins, *et al.* (Table I) was not observed by the writer.

The only two thin sections studied showed granular and oolitic textures. One of them contains almost 25 per cent by volume metallic minerals mainly in poorly formed granules with quartz-filled centres. The other, and probably more typical rock, is composed mainly of chert containing chert and carbonate oolites and granules with thin rims of sparse, dusty iron oxide. Some of the carbonate oolites seem to be partly replaced by chert. In both specimens the chert shows a seriate texture similar to that in cherty metallic iron-formation.

### *Regional Comparison of the Sokoman Formation*

The Sokoman Formation is remarkably similar throughout the two major regions in which it is exposed: the Walsh zone, and the western region south and west of Lac Le Fer. All stratigraphic units of Table II are found in both regions and corresponding lithologies may be readily matched. Locally one unit or more may be missing in either region. The total thickness of the formation is generally similar throughout.

At Ritchie Lake the silicate–carbonate and thin-bedded jaspilite units are missing. The cherty metallic iron-formation contains both jaspery metallic and grey cherty metallic rock types, but these do not seem to be separable into upper and lower subdivisions. Lean cherts are present and may be near the top of the formation, but the overlap of the Purdy Formation makes this uncertain. The departure there from the normal stratigraphic succession may indicate either a northward change in the stratigraphy of the Sokoman Formation or simply a local variation.

Iron-formation near Lac Low bears little resemblance to that in most of Wakuach Lake map-area. Neither the silicate–carbonate nor thin-bedded jaspilite members are present and the basal unit is interbedded chert and iron carbonate. It is succeeded by cherty metallic iron-formation with a low metallic mineral content but significant amounts of minnesotaite. The remainder of the section is not exposed, but an isolated outcrop, 1.8 miles southeast of the southernmost tip of the lake, appears to be fairly high in the section and is composed of interbedded iron carbonate and cherty metallic layers. The cherty metallic phase of the rock contains up to 30 per cent by volume of metallic iron oxide and 25 per cent minnesotaite. Thus the Sokoman Formation in the Lac Low region may, to a large degree, be an iron carbonate–iron silicate facies. Gastil, *et al.* (1960, p. 30) recognized a facies change in the iron-formation of the southern Labrador geosyncline. On the inner or northwestern side of this part of the geosyncline, the iron-formation is mainly an iron oxide facies, whereas on the outer or southeastern side it is predominantly a silicate–carbonate facies. No persistent facies change is evident in Wakuach Lake map-area, but the composition of the Sokoman Formation in Lac Low region may herald such an eastward change. However, the eastern limit of iron-formation deposition may not be much farther east if the missing segments of iron-formation north and south of Lac Low are interpreted as the result of non-deposition.

### *Metamorphism*

Recrystallization appears to be the only reliable guide to metamorphism in the Sokoman Formation of this area. James (1955, p. 1475) provided a rough scale by which this may be judged. In the chlorite zone of northern Michigan the grain size of quartz derived from chert is typically 0.03 mm with a maximum of 0.05 mm; in the biotite zone it is characterized by a range from 0.05 to 0.10 mm. The Sokoman Formation in Wakuach Lake map-area contains recrystallized chert with grain sizes ranging mainly from less than 0.001 mm to 0.1 mm. The texture is seriate and minutely patchy so that an average grain size is difficult to determine and may be of doubtful value. Perhaps of greater significance is the heterogeneity of grain size that characterizes these rocks. Microscopic patches in which the grain size is less than 0.001 mm exist side by side with patches in which the grain size may reach a maximum of 0.3 mm. A comparison with James' scale of grain sizes would lead to the conclusion that the metamorphism is of biotite grade or less but the heterogeneity and size range casts doubt on whether they might be so simply classified.

The significance of the silicate minerals, most notably minnesotaite, is uncertain. James thought they might be derived from a pre-existing silicate mineral of similar composition, but was in doubt as to whether they were of diagenetic or metamorphic origin (James, 1955, p. 1474). Yoder (1957, p. 234) deduced that with rising temperatures minnesotaite should form by reaction of greenalite and quartz, but since greenalite and minnesotaite could not be synthesized (Smith, 1957, pp. 230–231) reaction temperatures are uncertain. No evidence was obtained during the present study to indicate that minnesotaite was formed from an earlier mineral, but the growth of minnesotaite along fractures noted previously does warrant the conclusion that at least some of it is secondary.

No systematic geographical variation in mineralogy or textures that might be related to regional metamorphism could be detected. Hence the variation in metamorphic intensity throughout the region in which the Sokoman Formation outcrops must have been low.

### *Mode of Origin*

The origin of siliceous iron-formation is still in question and it is beyond the scope of this report to contribute substantially to the problem. A few general observations from this region, however, may be helpful in its ultimate solution.

Two major theories are held on the source of iron and silica: 1) they were derived from a landmass of low relief by deep weathering and transported to the site of deposition in solution or colloidal suspension and precipitated by chemical or organic means (Gruner, 1922; Moore and Maynard, 1929); 2) they were contributed directly to the basin of deposition from volcanic sources (Van Hise and Leith, 1911). Kirkland (1950, p. 100), on the basis of studies conducted near Hook Lake and southward, concluded that most of the iron and silica of the Sokoman Formation was derived from weathering of a landmass either east or west of the depositional basin but probably east of it. He suggested that part of the iron may also have been contributed from volcanic sources. Dufresne (1952, pp. 168–170) similarly concluded that the iron and silica were derived by intensive weathering of a landmass probably situated east of the Trough.

Features pertinent to the problem are as follows:

1. The geometry and internal structures of the Sokoman Formation and its stratigraphic precursors denote a long period of sedimentation under stable, mainly shallow water conditions. Both the Wishart and Ruth Formations are thin widespread units with fairly constant thicknesses. The Denault Formation thickens markedly in the Hurst Lake–Lac Chassin region, but shows evidence of shallow-water deposition throughout. The iron-formation has a modest range of thicknesses and such shallow water features as oolites and fragmental structures analogous to intraformational conglomerates.

2. The Wishart Formation, composed mainly of detrital quartz with lesser potash feldspar and albite, is most probably derived from a granitic landscape. The close association and similarity of distribution of the Wishart and Sokoman Formations makes it unlikely that the configuration of the depositional basin was altered to any degree during the time that these formations were being deposited.

3. Volcanic rocks are interbedded with the Wishart, Ruth, and Sokoman Formations a few miles south of Wakuach Lake map-area (Frarey, 1961; Sauvé, 1953, p. 42).

The evidence is far from conclusive but raises doubt that sufficient iron could have been obtained by deep-weathering of the sedimentary source material exposed at that time. The source material during Wishart time appears to have been granitic. Since the depositional basin shows no evidence of tectonic movement between Wishart and Sokoman times it seems unlikely that the same source area could have changed composition markedly during the interval. Granitic rocks are not a rich source of iron and in weathering might be expected to produce large amounts of other materials. Basalts on the other hand are contemporaneous with the Sokoman Formation, are a rich source of iron themselves, and might be expected to be accompanied by iron-rich emanations. On the basis of evidence available in Wakuach Lake map-area, the writer favours the view that iron and silica of the Sokoman Formation are from volcanic sources.

## Purdy Formation

### *Distribution and Thickness*

The Purdy Formation is exposed along the western side of the Labrador Trough for a strike distance of about 25 miles. It is not known elsewhere in the area. At Ritchie Lake it has a minimum thickness of 1,500 feet, but the total thickness is obscured by folding.

### *Lithology*

The formation is composed essentially of dolomite although one outcrop of greenish siltstone was observed at the centre of a small anticline. The dolomite ranges from thinly bedded to massive and may be black, grey, or pink in colour. Some beds are sandy or silty. Algal structures and penecontemporaneous slump structures are both common. A characteristic rock-type is of thinly layered (less than 1 inch) dolomite in which dark, light, and commonly pink layers alternate. All layers are slightly wavy. Chert is not abundant in this formation, but a few outcrops are severely dissected by crisscrossing quartz veinlets.

The dolomites observed in thin section have a heterogeneous microtexture. Patches of dolomite with crystal sizes as large as 0.5 mm alternate with patches in which the crystal size appears to be submicroscopic. In at least one specimen the heterogeneity may be the result of brecciation. Sandy dolomites contain as much as 50 per cent rounded to subrounded clastic grains, mainly quartz with minor potash feldspar.

One algal structure studied in thin section consists of elongate dolomite crystals radiating from a central vertical axis. The core is a jumble of coarsely crystalline, randomly oriented dolomite. Radiating crystals may be several millimetres long but parts of the array are separated into tiers of concentric, discontinuous thin seams of finely crystalline quartz. Vague colour banding that passes from crystal to crystal but cannot be detected under crossed nicols also marks the concentric nature of the structure.

### *External Structural Relations and Mode of Origin*

The Purdy Formation truncates the Sokoman Formation and for part of its length rests directly on the Basement Complex. An unconformity at the base of the Purdy Formation is, therefore, indicated but little is known of its extent and significance. The Menihek Formation overlies the Purdy Formation with approximate concordance on a regional scale, but the contact was not observed so the actual relationship is unknown.

The Purdy dolomites were evidently deposited in shallow water where algal colonies could flourish and wave-action could disrupt newly formed strata. Lack of such features as crossbedding and graded bedding together with the generally fine-grained size of the primary dolomite suggest that deposition was predominantly chemical. Minor quartz and potash feldspar grains in the sandy beds were probably contributed by the granitic areas to the west.

## Menihek Formation

The Menihek Formation in this report includes much of the former Howse Group. Although increasing recognition has been given to the equivalence of these units in past years (Harrison, 1952, p. 13; Baragar, 1960, p. 1595; Frarey, 1961), only recently has formal correlation been made and the name Howse Group dropped (Frarey and Duffell, 1964).

### *Distribution and Thickness*

The present and pre-folding distributions of the Menihek Formation are shown on Figure 14; the former Howse Group is represented by the eastern belt of outcrop areas, the traditional Menihek Formation by the remainder. In both areas the respective units immediately overlie the Sokoman Formation. The former Howse Group rocks are thickly intruded by gabbro sills so that much of the unit is presently represented by a succession of thin inter-sill strips that outcrop only intermittently.

The thickness of the Menihek Formation in the western part of the area was not determined. Dufresne (1952, p. 26) listed its thickness in the area between Burnt Creek and Goodwood as 1,500(?) feet, and Harrison (1952, p. 13) reported the maximum thickness exposed in the Burnt Creek strip as 1,000 feet, with limits unknown. West of Murdoch Lake the former Howse Group has a thickness of about 7,000 feet. North of Lac Low the unit seems to be missing entirely.

### *Lithology*

*Sedimentary Rocks*

In the western part of the Labrador Trough, according to Dufresne (1952), Harrison (1952), and Gross (1951), the Menihek Formation is mainly carbonaceous slates or shales and fine-grained siltstones. Minor interbedded dolomite is found locally (Harrison, 1952, p. 13). In the Lac Gillard–Annabel Lake valley, the most westerly exposures of significance observed by the writer, the formation is predominantly a greywacke–shale assemblage. Dark grey to jet black, fine-grained greywackes with generally subordinate interbedded dark shales are characteristic. Bedding is not commonly conspicuous nor is the rock generally fissile. The lower part of the former Howse Group, west of the belt of closely spaced sills, is composed of a similar assemblage of dark greywackes and shales, but inter-sill exposures higher in the formation are mainly black shales with some quartzites and subgreywackes. Commonly the inter-sill shales are partly, and in places almost completely, composed of fine-grained sulphide minerals, mainly pyrrhotite and pyrite. Some of the sulphide-rich bands west of Murdoch Lake are more than 100 feet thick and were traced for several thousand feet along strike (Baragar, 1958, p. 6). They are commonly fine bedded and the sulphide components are intimately intermixed with the shaly material. The sulphide minerals may have been an integral part of the original deposit. In places, the bedded

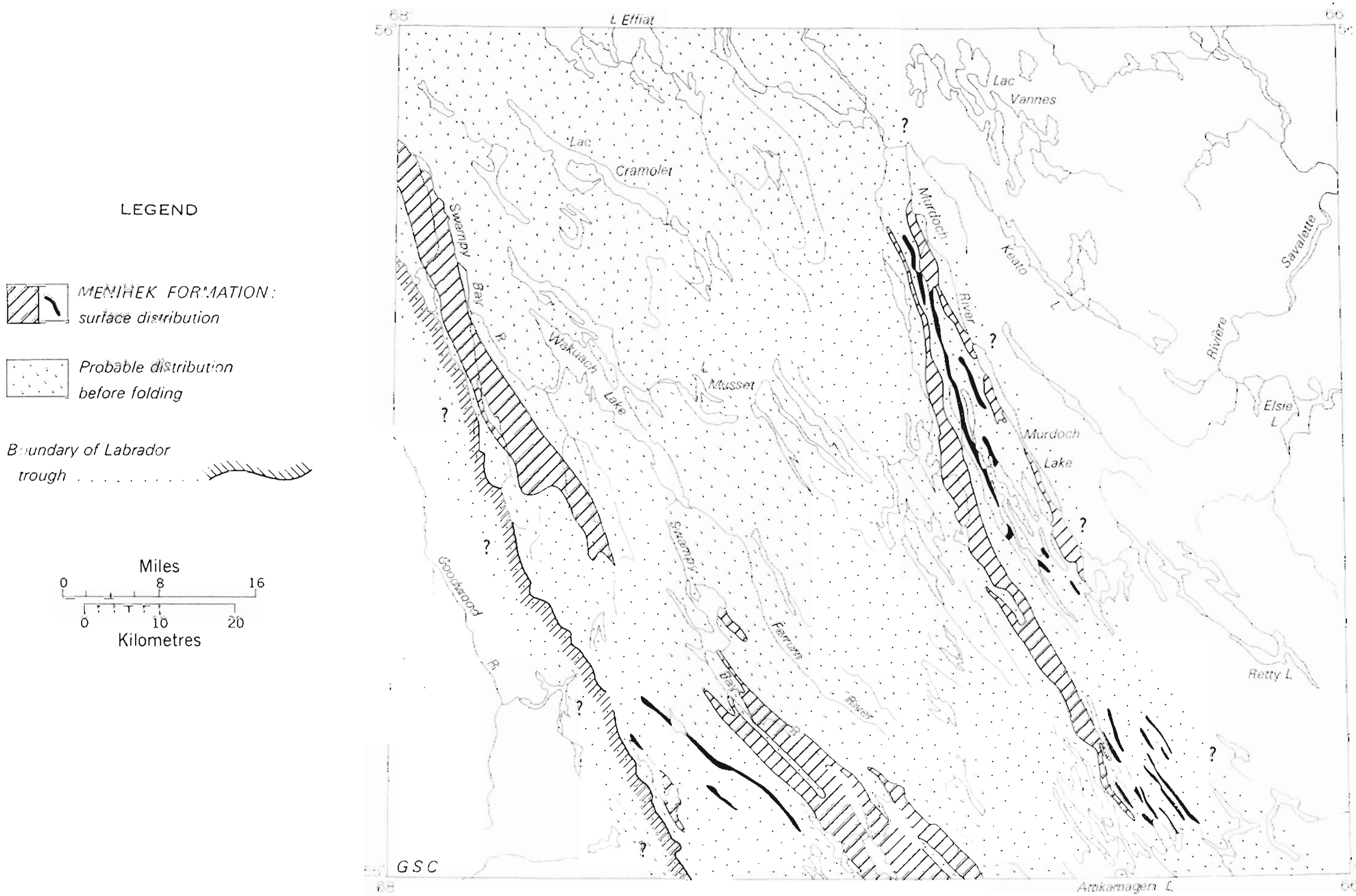



FIGURE 14. Distribution of the Menihek Formation in Wakuach Lake map-area.

sulphide minerals are cut by veinlets of coarser grained pyrite and pyrrhotite. Graded bedding appears to be absent in the greywacke–shales of the Menihek Formation, except for a small area 1 mile to 2 miles northeast of Hurst Lake where a few exposures of rhythmically graded beds were observed.

The greywackes are generally composed of sharply angular fragments of quartz and feldspar in an indistinct groundmass of altered feldspar grains, chlorite, biotite, sericite, carbon, hornblende, pyroxene, and possibly other minerals. Fragments of shale, and, less commonly, basic or intermediate volcanic rocks are important constituents in some of the thin sections studied. Staining of three thin sections from widely separated points in the lower part of the eastern belt (Howse Group) showed that 5 to 10 per cent of these rocks consist of potash feldspar. Following is a fairly typical range of compositions for the greywackes examined from both the Lac Gillard and eastern belts:

| | |
|---|---|
| quartz (minor chert) | 15-25 per cent |
| feldspar | 30–50 per cent |
| indistinct groundmass | 20–40 per cent |
| rock fragments | 5-10 per cent |

Estimates of compositions are difficult since groundmass and altered feldspar and rock fragments blend into one another to varying degrees. Figure 12 B and C shows typical Menihek greywackes, as seen in thin section in the petrographic microscope.

Quartzite or subgreywacke from one inter-sill exposure in the upper part of the formation northeast of Hurst Lake is composed of well-rounded quartz grains up to half a mm in diameter in a sparse chloritic matrix.

*Volcanic Rocks*

Buff weathering grey basalts with a maximum thickness of about 3,000 feet cap the Menihek Formation in the northern part of the eastern belt. The lava pile appears thickest northwest of Irene Lake and interfingers southward into underlying Menihek sediments. The only known pyroclastic member associated with these volcanic rocks is at the base of the thickest part of the volcanic pile, where it continues southward into the sedimentary assemblage for at least 10 or 12 miles. Pillow lavas are rare; one thin layer of pillows occurs near the bottom of the volcanic sequence west of Irene Lake.

The lavas are similar to those of the Attikamagen Formation. A characteristic type in both formations is a light grey, densely aphanitic rock composed of minute, hollow and forked plagioclase needles thickly scattered through an indistinct brownish groundmass of pyroxene and dusty sphene. The finer grained lavas of this formation closely resemble the Attikamagen basalts (Fig. 6 A). Similarly the coarser grained varieties in both formations are indistinguishable. Unaltered plagioclase ranges in composition from about $An_{50}$ to $An_{65}$. Augite is the main pyroxene; pigeonite was not recognized but would, in any event, be difficult to distinguish in the fine-grained lavas. In some of the coarser grained lavas, pigeonite may be represented by tiny patches of serpentine or chlorite in the centres of some of the augite crystals. Clusters of microphenocrysts are common in the finer grained lavas. Generally they are plagioclase and pyroxene with or without pseudomorphs of olivine and pyroxene. A chemical analysis of lavas from the west side of Murdoch Lake is given in Table IX.

### *External Structural Relations*

In the area about Knob Lake, Harrison (1952, p. 13) and Dufresne (1952, p. 47) found the Menihek Formation and underlying Sokoman Formation to be conformable and locally, at least, gradational at the contact. In the western part of the present map-area not only does the Menihek Formation overlap the Sokoman Formation but the Purdy dolomite intervenes. Elsewhere in the area the two formations appear to be roughly concordant, but the missing segment of Sokoman Formation between Hurst Lake and Lac Low may indicate local emergence and erosion. In other parts of the Labrador Trough considerable evidence for a disconformity has accumulated. Sauvé (1953, p. 39) noted that lava flows intervening between the Sokoman and Menihek Formations in the Astray Lake area are partly eroded. Dufresne (1952, p. 57) drew attention to the marked difference in sedimentation represented by the lithologies of the two formations, and noted that in the Sawyer Lake area they are separated by a thick assemblage of ferruginous siltstones. In the Lac Herodier area, the Sokoman Formation on the west side of the Trough shows marked erosion and is succeeded by a coarse, angular jasper conglomerate (Fahrig, 1955, p. 4). In Wakuach Lake map-area, a jasper-bearing conglomerate is found near Walsh Lake at the base of the Murdoch Formation on the east side of the Walsh Lake fault (Frarey, 1952, p. 3; Kavanagh, 1954, p. 31). Kavanagh (1954, pp. 134–136) suggested that this conglomerate marks the unconformity between the Sokoman and Menihek Formations. A variety of evidence, therefore, points to a widespread erosional unconformity separating the two formations but there is little indication of more than minor discordance between them. An angular unconformity at the base of the Menihek Formation in what must be the Willbob Lake and Thompson Lake map-areas was reported by Gastil, *et al.* (1960, p. 35) from the work of Kavanagh (1954). The writer was unable to find reference to such an unconformity in Kavanagh's thesis and Frarey (1967) gives no evidence to suggest that an angular unconformity exists in these areas.

### *Mode of Origin*

Dark shales, siltstones, and greywackes of the Menihek Formation are indicative of relatively deep-water sedimentation. Volcanic particles recognizable in the greywackes together with abundant chloritic and other ferromagnesian material testify to the importance of the volcanic contribution to the origin of the formation.

The source of angular quartz, plagioclase, and potash feldspar is uncertain, but because these are the major constituents of granitic rocks, the granitic basement would be a reasonable source to postulate.

### *Correlation*

Gastil, *et al.* (1960, p. 25) correlated the Menihek Formation with the Larch River Series (Bergeron, 1957) of the northern part of the Labrador Trough and with a post-iron-formation complex of schists and gneisses at its southern end.

## Doublet Group

### Murdoch Formation

The rocks formerly called Murdoch Group are redesignated in this report as the Murdoch Formation, in accordance with earlier usage by the writer (Baragar, 1958) and the revised stratigraphy of the Labrador Trough proposed recently by Frarey and Duffell (1964). Apart from minor conglomerate localized at and near Walsh Lake, the unit is predominantly composed of pyroclastic rocks and is sufficiently homogeneous and distinctive to be consistent with the definition of a formation, as set forth by the American Commission on Stratigraphic Nomenclature (1961, p. 650).

#### *Distribution and Thickness*

The formation is represented by a single belt in the eastern half of the map-area, which is evidently continuous from the south to the north boundary of the area. Its thickness near Murdoch Lake is estimated to be 6,000 feet, whereas in the south-central part of the area Kavanagh (1954, p. 30a) gave the thickness as 3,200 feet. Both estimates are likely to be in error, for bedding is uncommon and the rocks appear to have been highly deformed.

#### *Lithology*

The Murdoch Formation is predominantly composed of basic pyroclastic rocks and green schists presumably derived from them. Massive basaltic flows, argillites, acidic tuffs, and conglomerate are minor components.

The pyroclastic rocks include tuffs, lapilli, agglomerates, and breccias, with the coarser fragmental rocks dominant. Fragment size most commonly ranges from half an inch to 3 inches, but fragments as much as 2 feet long have been noted. The fragments are mainly subrounded and are composed of green basalts, commonly highly amygdaloidal. In places they are elongate and tend to parallel the formation. Bedding is rare in the coarser pyroclastic rocks, but can occasionally be seen in the lapilli and tuffs where it generally appears as inconspicuous layering half an inch to about 2 inches thick.

Basaltic flows appear to form a very small proportion of the formation but are evidently widespread. They were noted by Frarey (1952) and Kavanagh

(1954) in the Willbob Lake area and are found in scattered localities east of Murdoch and Irene Lakes and north of Lac Low. Generally the flows are massive, but in one locality east of Lac d'Argent they are well pillowed. Highly porphyritic flows occur in two places east of Murdoch Lake.

The basic volcanic rocks are composed of amphibole, albite, chlorite, epidote–zoisite, sphene, and magnetite commonly in this approximate order of abundance. Biotite was reported to be an important constituent in parts of this formation in the Willbob Lake area (Frarey, 1954, p. 28). The amphibole is generally light green or colourless and is probably tremolite–actinolite. According to Kavanagh (1954, p. 39), chlorite predominates over amphibole in the highly schistose rocks, but the reverse is normal in the more massive rocks. Igneous texture is still preserved in some of the basaltic fragments although the rocks have all been completely recrystallized, and with the exception of one relic pyroxene no primary minerals were observed. In the greenschists, chlorite and amphibole commonly show pronounced preferred orientation and primary textures are rarely recognizable.

Acidic tuffs are found in numerous lenses interbedded with the basic pyroclastic rocks in a belt that extends for about 8 to 10 miles north of the north end of Murdoch Lake. Several such lenses outcrop on shore at the northern tip of the lake. The lenses range from a few inches to about 200 to 300 feet in thickness, and some are known to have a strike length of at least 1,000 feet. The acidic tuffs are generally thinly layered, pink to buff rocks associated in places with thin-bedded, grey to black argillites. They are composed largely of very fine grained ($\pm$ 0.01 mm) alkali feldspar and quartz with lesser biotite, chlorite, iron-bearing carbonate, sphene, magnetite, and pyrite. Pyrite is locally abundant, and together with iron-bearing carbonate, may give outcrops of this rock a rusty appearance. One boulder of acidic tuffs found near the northern end of Murdoch Lake contains several per cent chalcopyrite.

Conglomerates are restricted to a small area in the vicinity of Walsh Lake, but because of their stratigraphic implications are an important unit of the Murdoch Formation. They were not seen by the writer and the following description was summarized from accounts by Frarey (1952, p. 3; 1954, p. 28) and Kavanagh (1954, p. 31). The conglomerate contains pebbles ranging from a quarter inch to 3 inches across in a quartzite groundmass. Pebbles are composed of quartz, jaspery iron-formation, chert, quartzite, altered slate, and acidic porphyry. The groundmass consists mainly of rounded quartz grains with minor feldspar grains in a sparse sericitic matrix.

### *Internal Structural Relations*

Schistosity is widespread throughout the formation. In many places it completely obliterates all primary structures and few of the rocks are entirely free of it. It strikes roughly parallel with the formation and dips mainly between 50 and 75 degrees east.

### External Structural Relations

The stratigraphic significance of the conglomerate at the base of the Murdoch Formation at Walsh Lake was mentioned in the section on the Menihek Formation. Since this conglomerate contains pebbles of iron-formation it is later than the Sokoman Formation at least. Kavanagh (1954, pp. 134–136) concluded that the conglomerate marks the unconformity noted elsewhere in the Labrador Trough between the Sokoman and Menihek Formations and that the remainder of the Murdoch Formation could pre-date, be contemporaneous with, or post-date the Menihek Formation. West of Murdoch and Irene Lakes, presumably on the west side of the Walsh Lake fault, typical Murdoch agglomerates in places outcrop within 4 or 5 feet of the top of the Menihek (Howse) lavas. Intervening conglomerates are, therefore, unlikely and the Murdoch Formation appears to succeed the Menihek Formation directly. If the conglomerates near Walsh Lake are to be related to the unconformity at the top of the Sokoman Formation, and yet the Murdoch Formation in another place is to succeed the Menihek Formation directly, the latter must be assumed to have pinched out in the interval. This is not impossible. The conglomerates are on the east side of the fault and might be assumed to have come from a point originally much farther east. Moreover the Menihek Formation is known to pinch out northward, since it is absent north of Lac Low, and might well do the same eastward.

### Metamorphism

Rocks of this formation are predominantly or entirely metamorphic. Mineral assemblages are typical of the greenschist facies, but the prominent schistosity may be a modification caused by movement on the subjacent Walsh Lake fault. Kavanagh's observation (1954, p. 39) that chlorite predominates in the schists and amphibole in the more massive rocks suggests a retrograde effect due to movement.

### Mode of Origin

The predominance of pyroclastic material in the Murdoch Formation with only local, widely scattered flows indicates that volcanism during this period was extraordinarily explosive.

## Iron-formation

### Distribution and Thickness

Lenses of iron-formation found in the contact area between the Murdoch and Thompson Lake Formations just east of the north end of Murdoch Lake interfinger and are interbedded with both formations. The main lens is about 2 miles long and a maximum of about 600 feet thick. At its northern end it is interbedded with Thompson Lake sedimentary rocks, and at its southern end with Murdoch volcanic rocks. Smaller lenses a few hundred feet north, and $1\frac{1}{2}$ miles south, of the main lens are wholly within the Thompson Lake and Murdoch Formations respectively. The southern lens may be continuous with the main body of iron-formation; outcrop in the intervening distance is lacking.

### *Lithology*

The major rock types are in general: (1) thin bedded 'cherty'[1] rock rich in hematite or magnetite, and (2) interlayered iron carbonate and 'chert'[1] with rare metallic minerals. The carbonate facies in at least part of the main lens predominates in the upper one-third to one-half of the formation.

Generally the metallic-rich part of the iron-formation consists of alternating hematite- or magnetite-rich layers and white, red, green, or brown quartz, quartz–silicate, or quartz–carbonate layers. The layers commonly range from a quarter inch to 4 inches but are mostly less than 2 inches thick. Some beds appear to be predominantly metallic and locally these may be cut by veinlets of massive hematite a quarter to half an inch wide. Crosscutting quartz veinlets are ubiquitous.

In thin section, the metallic-rich rocks are seen to be composed essentially of finely crystalline quartz with a metallic mineral content ranging mainly between 5 and 50 per cent by volume from layer to layer. Iron-bearing carbonate generally composes about 5 per cent of the rock. Other minerals appearing in varying quantities from place to place are: riebeckite–arfvedsonite amphibole, tremolite–actinolite amphibole, acmite, and stilpnomelane. Magnetite, hematite, and minor quantities of hydrous iron oxide are the metallic minerals present. The relative proportions of magnetite and hematite are uncertain because only two polished sections were examined and each contains one of the minerals exclusively. In several thin sections euhedral magnetite (or possibly martite) appears to be dominant. Quartz ranges in grain size from 0.01 to 0.5 mm, but most is in the range 0.03 to 0.05 mm. The metallic minerals are generally between 0.05 and 0.1 mm in individual crystal-size, but aggregates of crystals range up to 1 mm or more across.

The mineral composition of the metallic-rich part of the iron-formation is particularly interesting. Most specimens contain minor bluish, sodium–iron amphibole; a few contain a green amphibole that is nearly colourless in thin section. In places, zoned amphibole seems to be composed of both varieties. Markedly zoned crystals grade abruptly from a colourless interior to a lavender or bluish rim. Other optical properties of the two amphiboles appear to be as follows:

| Interior | Rim |
|---|---|
| 2V—moderately large, negative | moderately large, negative? |
| Dispersion—r $>$ v strong | r $>$ v strong |
| Optic plane—010 | $\perp$ 010? |
| Z subparallel with c | X subparallel with c |

[1] Chert and cherty are used on the assumption that the original material was chert. It is now a sugary-textured quartz.

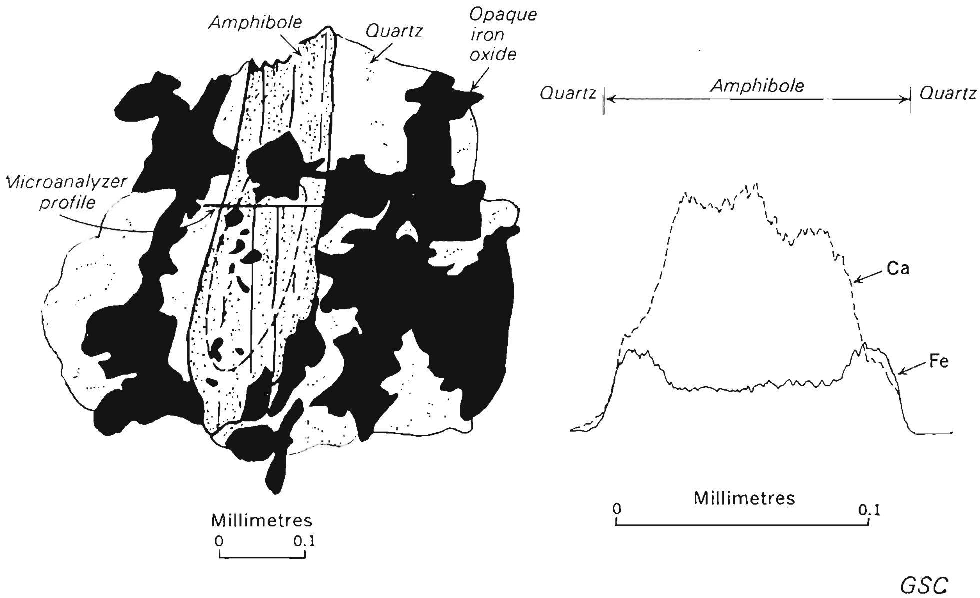



FIGURE 15. Zoned amphibole from Doublet Group iron-formation with electron beam microanalyzer profiles for calcium and iron.

Figure 15 shows a sketch of one of the zoned amphibole crystals (specimen B 116) in thin section and profiles of iron and calcium content across it[1]. Note that iron is enriched at the rims and that calcium increases more or less steadily from the rim to the interior. Thus a calcic amphibole, probably one of the hornblendes, forms the core of the zoned crystals and grades rapidly outward to an iron-rich phase, presumably riebeckite–arfvedsonite, at the margin. The pleochroic colours of the latter suggest that it is mainly riebeckite according to recent data by Ernst (1962). The zoning probably results from progressive substitution of ferric iron and sodium for calcium and magnesium towards the exterior of the crystal. Acmite[2] occurs in small (.01 mm), equidimensional, golden yellow crystals, thickly aggregated in some layers. Stilpnomelane is confined to carbonate-rich layers where it appears as a brown to yellow micaceous mineral associated particularly with patches and bands of interstitial hydrous iron oxide. This association suggests that it may have resulted from weathering or groundwater activities or possibly from post-metamorphic hydrothermal activity.

The carbonate facies of the iron-formation typically comprises alternating layers, generally half an inch to 8 inches thick, of chocolate brown weathering carbonate and sugary grey 'chert'. Carbonate forms from 10 to 50 per cent of the rock. Metallic minerals are rare or absent. Other minerals present in minor amounts in the predominantly carbonate layers include stilpnomelane, goethite or limonite, carbon, and quartz. Probably both stilpnomelane and the hydrous

[1] Electron beam microanalysis by G. R. Lachance, Geological Survey. The instrument used is Elion D. E. M. 301 microanalyzer.

[2] Confirmed by X-Ray Diffraction Laboratory, Geological Survey.

iron oxides are of secondary origin. The grain size of the carbonate mineral ranges generally from about 0.03 to 0.3 mm. A chemical analysis of a composite sample of two specimens of carbonate-rich layers is given in Table III. From the analysis can be calculated the amounts, types, and compositions of carbonate minerals present if one assumes that all the calcium is to be recast into a dolomite-type mineral, that excess $CO_2$ will be in siderite, and that MgO and FeO will enter dolomite in the same ratio as they are present in the rock when the amount of FeO for siderite is deducted. Ferric oxide present in the analysis is considered to represent hydrous iron oxides and therefore involves no ferrous oxide. This leaves minor FeO and MgO for stilpnomelane. Because of the uncertainty of the stilpnomelane formula a complete recast of the analysis into the modal minerals present has not been attempted. Most of the remaining mineral is obviously quartz.

TABLE III

*Chemical Analysis of Iron-formation Carbonate from the Doublet Group*

| | |
|---|---|
| $SiO_2$ | 35.26 |
| $Al_2O_3$ | 0.72 |
| $Fe_2O_3$ | 1.45 |
| FeO | 12.70 |
| CaO | 15.93 |
| MgO | 5.94 |
| $Na_2O$ | 0.19 |
| $K_2O$ | 0.09 |
| $H_2O+$ | 0.39 |
| $H_2O-$ | 0.09 |
| $TiO_2$ | 0.11 |
| $P_2O_5$ | 0.04 |
| MnO | 0.19 |
| $CO_2$ | 26.70 |
| C | 0.10 |
| | 99.90 |

Carbonate minerals present based on partial recast of the analysis.
$Ca(Mg_{.52}Fe_{.48})(CO_3)_2$
(Ankerite) = 56.6%
$FeCO_3$(Siderite) = 4.5%

### *External Structural Relations*

The iron-formation is closely associated with greywacke–siltstones of the lower part of the Thompson Lake Formation. North of the main body of iron-formation, metallic-rich beds and siltstone layers as little as 4 feet thick are intimately interbedded and some of the siltstones carry up to 10 or 15 per cent metallic minerals. The southern lens of iron-formation, although enclosed by Murdoch volcanic rocks, is immediately accompanied by similar grey siltstones.

The iron-formation thus marks the transition from a predominantly volcanic to a predominantly sedimentary environment. Whether or not this was a necessary factor in its development, however, is a matter for speculation.

### *Metamorphism*

The iron-formation is within the greenschist facies zone of metamorphism. The most significant change to be noted as a result is a generally coarser and more uniform grain size for both quartz and metallic minerals in comparison with the minerals in the Sokoman Formation farther west. Both formations might reasonably be considered to have been mixtures of chert and fine-grained iron minerals originally. Now the Sokoman Formation contains quartz ranging from submicroscopic to about 0.1 mm in grain size and metallic crystals ranging generally from 0.005 to 0.05 mm across. Corresponding grain-size ranges for quartz and metallic crystals in the Doublet Group iron-formation are 0.01 to 0.5 mm and 0.05 to 0.1 mm, respectively, and none of the extremely fine-grained fractions are represented. Riebeckite occurs in both formations, but in the latter it may attain a crystal-size of 2.5 mm long and 0.5 mm thick in contrast with the hair-like crystals observed in the Sokoman Formation. Acmite and tremolite–actinolite amphibole are confined to the Doublet Group iron-formation and are considered direct developments of metamorphism.

## Thompson Lake Formation

### *Name, Distribution, and Thickness*

The Thompson Lake Formation takes its name from Thompson Lake in the southeastern part of the map-area where it is well exposed (Frarey and Duffell, 1964). The formation is essentially co-extensive with a belt of closely spaced sills that is continuous from the southeastern corner to the northern boundary of the map-area. Accordingly it appears mainly as thin inter-sill strips that are easily eroded and underlie the valleys. Exposure is generally poor and as a result of this and of the repeated partition of the formation by sills, estimates of thickness are difficult to make and likely to be imprecise. The thickness of the formation as estimated by Kavanagh (1954, p. 30) in the Hyland Lake area is 2,300 feet; by Frarey (1954, p. 31) in the Willbob Lake area, 2,000 feet; and by the writer in the Ahr Lake area, 1,500 feet. The formation thins northward, and northeast of Lac d'Argent it probably does not exceed 300 feet in thickness.

### *Lithology*

A definitive description of the lithology is not possible owing to the fragmentary nature of data available. Rock types most frequently encountered are thin-bedded, black or grey slates and impure quartzites. Greywackes, greywacke–siltstones, chert, and conglomerate are less common.

In the northern half of the map-area, where the formation was mapped by the writer, more than 75 per cent of the outcrops observed consist of grey to black, thin-bedded slates, shales, or siltstones. Most of the remaining outcrops are composed of quartzites, and impure quartzites or subgreywacke. Argillaceous members rich in pyrite and/or pyrrhotite are fairly common. The quartzites range from thick- to thin-bedded and in at least one place are crossbedded. In thin section the argillaceous rocks are seen to be composed essentially of a fine-grained mosaic of quartz and feldspar with varying but generally substantial amounts of chlorite, white mica, and epidote–zoisite. Biotite, iron oxide or sulphide minerals, carbon, and sphene are plentiful in some specimens and tourmaline is an ubiquitous accessory mineral. The textures are typically metamorphic. Only two quartzite specimens were observed in thin section and both contain sufficient feldspar, or its epidote–zoisite equivalent, to be classed as arkoses. No sedimentary textures survive and the rocks are now mainly a mosaic of interlocking quartz and feldspar crystals or intergrown patches of epidote–zoisite and quartz.

In the southern half of the map-area arenaceous rocks seem to predominate. Kavanagh (1954, p. 45) reported that in the Hyland Lake area this formation is composed mainly of interfingering greywacke and quartzitic rocks. Most prevalent of these is a quartz-rich type containing 70 to 80 per cent quartz with most of the remainder mica and chlorite. In the Willbob Lake area, Frarey (1952, pp. 3–4; 1954, pp. 31–33) found that the formation comprises quartzite, subgreywacke, greywacke, minor conglomerate, and intimately interbedded quartzite and slate. The quartzite is composed of angular to rounded grains cemented by quartz, whereas the greywacke is composed mainly of angular quartz grains and slate fragments in a foliated matrix of biotite, quartz, and chlorite. Conglomerate found in one place, contains quartzite boulders and pebbles in a carbonate-rich matrix. Both Kavanagh and Frarey reported rare to moderately abundant crossbedding.

### *Internal Structural Relations*

In the northern part of the map-area, the formation forms an east-dipping, homoclinal assemblage. From Lac Harvut southward it becomes involved in a succession of gently southeast-plunging folds, which strike at a small angle to the trend of the formation and pass into the overlying Willbob Formation. Many of these folds appear to be rooted at décollement surfaces within the Thompson Lake Formation for they do not involve some of the lower members of the formation nor its contact with the volcanic rocks below.

### *External Structural Relations*

Interfingering of the Thompson Lake and underlying Murdoch Formations is evident near the north end of Murdoch Lake. Elsewhere the Thompson Lake Formation appears to overlie the Murdoch Formation conformably.

### *Origin*

The association within the Thompson Lake Formation of dark shales and siltstones, quartzites, subgreywackes, and greywackes indicates a mixed source of sediments and possibly a changing environment of deposition. Dark carbonaceous shales and siltstones with intermittent pyritic members indicate anaerobic conditions, whereas the quartzites were probably deposited in a well-aerated, shallow-water environment, where they would have been subject to continued winnowing by wave action. A similar contrast is evident in the origins of the sediments. Quartzite, composed in part of fairly coarse grains of quartz, must have been derived from a quartz-rich phaneritic rock, most likely granitic gneisses of the Basement Complex. The dark argillaceous sediments, on the other hand, contain a significant amount of ferromagnesian minerals, which might be expected to have had a volcanic origin. The interfingering of the Thompson Lake Formation with Murdoch pyroclastic rocks and its position between two major volcanic units makes it seem probable that the formation contains at least some volcanic material.

The Thompson Lake Formation may, therefore, be the result of intermixing of local volcanic and externally derived materials during a time of fluctuating sea-bottom levels.

## Willbob Formation

### *Name, Distribution, and Thickness*

The name of the formation is taken from Willbob Lake where, north and east of the lake, it is particularly well exposed (Frarey and Duffell, 1964). The Willbob Formation is the easternmost stratigraphic member of the Labrador Trough in this area and forms a single continuous belt between the north and south borders of the map. Its thickness in the Ahr Lake area is estimated to be at least 15,000 feet. Unrecognized repetition by faulting is possible, but is not thought to be significant. Harrison (1952, p. 15) reported the thickness of the Doublet volcanic rocks (Willbob Formation) in the Burnt Creek strip as at least 10,000 feet.

### *Lithology*

The Willbob Formation may be divided into three main lithological types: (1) pillowed and massive basaltic flows, (2) porphyritic basalts, and (3) inter-flow sedimentary rocks. The first-named is by far the most abundant.

#### *Pillowed and Massive Basalts*

Pillowed and massive basaltic flows alternate throughout the succession in a monotonous repetition of dark green, aphanitic to fine-grained, generally unfoliated rock. Each pillowed or massive sequence may be several hundred feet thick. Pyroclastic rocks associated with the flows were rarely observed by the writer, although some of the sedimentary interlayers, to be described shortly, may be tuffaceous. Frarey (1952, p. 4; 1954, p. 35) found minor pyroclastic beds to be fairly common throughout the lava sequence in the Willbob Lake area and at least one extensive area of breccia south of Retty Lake.

Irregular gabbroic masses, rarely more than a few hundred feet in longest dimension, are intimately associated with the lavas in many places. Little is known of their detailed relationship to the enclosing basalts, but presumably they are more closely related genetically to the extrusive rocks than to the profusion of sills that invade the region.

Lavas in the northern half of the region present a monotonously uniform appearance in thin section. They are typically composed of thickly intergrown tufts of shreddy actinolite with lesser chlorite in a scattered, confused matrix of fine-grained epidote–zoisite, albite, and, in places, quartz. The albite is generally clear and commonly untwinned. Finely granular sphene or leucoxene is an ubiquitous accessory mineral. Carbonate is locally present but is rarely abundant. Chloritoid was observed in one specimen of moderately foliated basalt from a point about 3 miles northeast of Lac d'Argent.

Willbob lavas from the southern half of the area were not seen by the writer but Kavanagh's description (1954, p. 52) of corresponding lavas in the Hyland Lake area is similar to that above. Frarey (1954, pp. 36–37) described, in addition, some fairly fresh basalts that he found in this formation in the Willbob Lake area. In these, the plagioclase is labradorite and augite is present.

*Porphyritic Basalts*

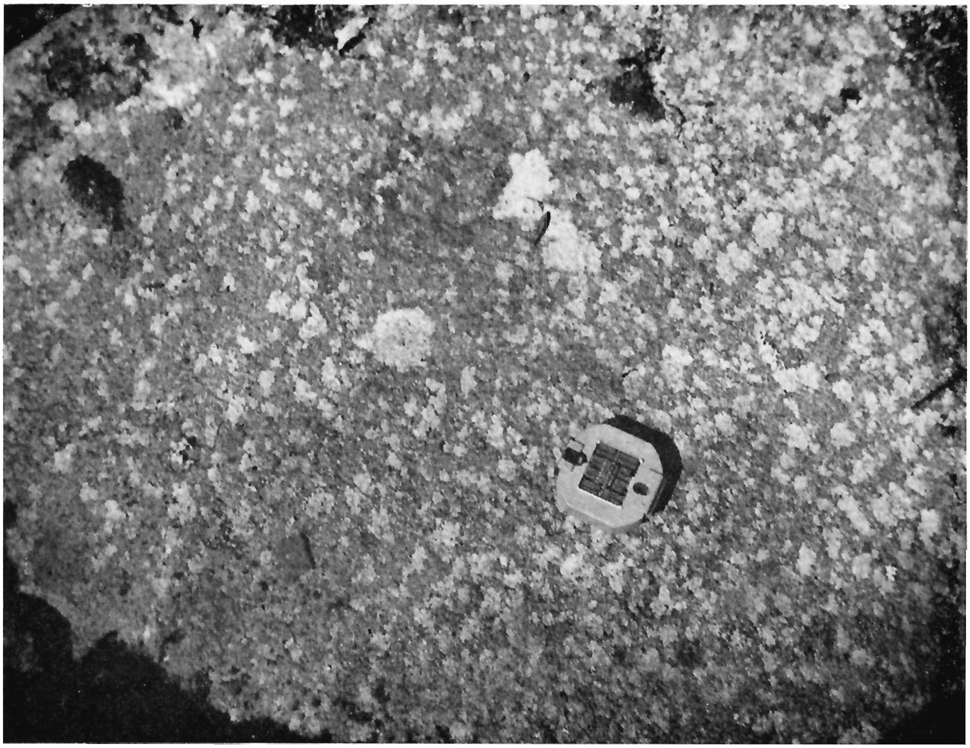
Three discrete layers of porphyritic basalts, ranging from about 300 to 1,500 feet in thickness, are found at different stratigraphic levels in the Willbob Formation of the Ahr Lake area. The layers are each separated by about 4,000 stratigraphic feet of pillowed and massive flows. The two lower members are predominantly massive basalt whereas the upper member is mainly pillowed lava. Chalky white clots of altered feldspar, 0.2 to 5 cm in diameter, are sparsely distributed through the rocks with a density generally in the range of 1 clot to 12 clots per square yard of outcrop. Where the density is less than this the rock is difficult to distinguish from normal massive or pillowed basalt. Gabbroic phases, with or without feldspathic clots are commonly associated with the porphyritic lavas. In two places the gabbroic phase is spectacular. White aggregates of altered feldspar up to 8 cm in diameter are sparsely but regularly distributed through a coarse-grained feldspathic groundmass (Pl. VIII). The similarity of this rock to the glomeroporphyritic sills to be described later is striking (cf. Pls. VIII and IX).

The porphyritic basalts are generally similar in mineral composition and texture to the normal massive and pillowed basalts already described. The feldspathic clots are composed of one or several individual plagioclase crystals. Most are now a dense, fine-grained intergrowth of epidote–zoisite and albite, and crystal outlines are obscure. More rarely a clear selvedge of albite rims or partly rims each feldspar crystal so that the structure of the composite mass is apparent.

In the Griffis Lake area, adjoining the southeastern side of the present map-area, Fahrig (1951) also found three layers of similar porphyritic basalt and associated gabbro in the Willbob Formation. Frarey (1952, p. 4) referred to porphyritic basalts of similar type in his description of Willbob lavas in the Willbob Lake area but did not map them separately. Possibly these porphyritic members are distinct stratigraphic units that are continuous for 50 miles or more.

*Inter-flow Sedimentary Rocks*

Several layers of fine-grained sedimentary rocks are interbedded with the Willbob lavas. In the Ahr Lake area these evidently furnished favourable sites



*W.R.A.B. 6–2–55*

PLATE VIII. Porphyritic gabbro associated with porphyritic flows of the Willbob Formation.

*W.R.A.B. 4–2–54*

PLATE IX. Glomeroporphyritic gabbro from a sill immediately west of Murdoch Lake.

for sill intrusions and now they are almost invariably found adjoining gabbro and ultramafic sills. The thickest of the sedimentary layers ranges from 500 to 700 feet in thickness and together they would not likely exceed 1,500 feet in aggregate thickness. They range from black, carbonaceous shales or slates composed mainly of finely recrystallized quartz, feldspar, and white mica to greenish grey, ferromagnesian-rich sediments formed predominantly of actinolite, chlorite, epidote–zoisite, and sphene. Presumably the latter are tuffaceous. Pyrite and pyrrhotite are sufficiently abundant in these sediments that outcrops are commonly distinguished by conspicuous gossan-coatings.

### *Internal Structural Relations*

The Willbob Formation in the northern half of the map-area forms a homoclinal assemblage of steeply east-dipping volcanic and associated rocks. Numerous top-determinations from well-formed pillows indicate that the flows face consistently eastward throughout the sequence. In the southern half of the area, the formation is involved with the Thompson Lake Formation in a succession of gently southeast-plunging folds. Numerous faults, particularly strike faults, are recognized in all parts of the formation, but most of them are of unknown significance. Many undoubtedly have rather minor displacements.

### *External Structural Relations*

The Willbob Formation overlies the Thompson Lake Formation with apparent conformity. However, in most places the actual contact is obscured by intervening sills or lack of outcrop.

### *Origin and Mechanism of Extrusion*

The formation is evidently the result of a continuing episode of effusive, submarine volcanism with brief intervals of normal, deep-sea sedimentation. Explosive eruptions appear to have been minimal. Small gabbro bodies intimately associated with the volcanic rocks are probably parts of the intricate chain of sub-surface conduits, which fed the volcanoes. Less certain is the relationship between the flows and the vast system of deeper seated sills, which pervade the rocks of the area. A direct linkage between flows and sills is suggested by a textural and compositional trait common to a variety of each. The porphyritic members of the Willbob Formation, particularly the gabbroic phases, are strikingly similar to the unusual glomeroporphyritic sills that invade the eastern belt of the Menihek Formation. That two manifestations of such extraordinary rock in the same magmatic province should have had completely independent origins seems hardly credible; most probably they derive from a common source and are nearly contemporaneous.

The assumption that glomeroporphyritic gabbro and the Willbob porphyritic basalts are directly related leads to some interesting speculation on the mechanism of extrusion. The sills may represent reservoirs of magma at intermediate depth. According to evidence compiled by MacDonald (1961, pp. 677–678), most volcanoes are underlain by, and directly connected with, magma chambers at depths of 4 to 7 km. These are thought to be laccolithic, lopolithic, or lens-shaped bodies. The swarm of glomeroporphyritic sills at some 3 to 6 km stratigraphically below the porphyritic basalts admirably fits the requirements of the theoretical magma chamber of MacDonald. Lava extrusion may have proceeded in two stages: first, to the level of the Menihek Formation where it spread widely through the region along bedding planes; and second, to the surface along one or many openings connected with the sills. Possibly the considerable lateral continuity of the porphyritic basalts is a direct reflection of the regional distribution of their magma in the underlying sills. The second stage of lava extrusion might be brought about by earth movements, which by changing the volume or shape of the magma chambers could increase the magmatic pressure, and at the same time provide fissures along which the magma could escape to the surface. Three levels of porphyritic basalts in the Willbob lavas imply three periods of porphyritic magma generation, but this may not be so. The magma may have been generated once and emplaced as sills. Thereafter it could have been driven to the surface at intervals as, from time to time, earth movements involved the sills. Otherwise the lavas would be normal basalts derived from depth or possibly from magma chambers at other levels.

#### *Age*

The Willbob Formation is considered the youngest stratigraphic unit of the Kaniapiskau Supergroup. The evidence for this, already presented under the various formation headings, is mainly stratigraphic and hinges largely on the relationship of the Murdoch Formation to the underlying Menihek Formation.

## Montagnais Group

All rocks intrusive into the Kaniapiskau Supergroup in the present map-area are sills. Several varieties are present, but all were emplaced prior to major folding and are therefore associated with the late geosynclinal or early orogenic phase (Harpum, 1960) of mountain-building. The field relations and generalized petrography of the sills are dealt with below; the detailed petrology of the gabbroic rocks is reserved for Chapter IV where it can be considered in conjunction with that of the volcanic rocks.

The major varieties of intrusive rocks are ultramafic rocks, glomeroporphyritic gabbros, and normal and meta-gabbros. The two latter are considered to be generally equivalent, but each is composed of sub-varieties of prime petrological interest. The ultramafic sills have been named the Retty Peridotite, and the gabbroic intrusions, the Wakuach Gabbro (Frarey and Duffell, 1964).

### Ultramafic Sills

#### *Distribution and Thickness*

Ultramafic sills are confined to a single continuous belt of one to six sills that intrudes the Thompson Lake and, less commonly, Willbob Formations. The major sill of the group has an average general thickness of about 1,500 to 2,000 feet and with minor interpretation can be traced for a strike distance of at least 120 miles. Minor sills of the group are as little as 175 feet thick but may, nevertheless, be continuous for strike distances of several miles. The entire belt of sills is generally less than 12 miles wide. It extends southeast of the present map-area for only about 10 miles and northward for an unknown distance.

Remnants of a parallel belt of ultramafic rocks may be present within the Younger Complex several miles northeast of the main belt. In the present map-area a lens of tremolite rock (map-unit 22b on the Wakuach Lake sheet) bordering Rivière à la Baleine is undoubtedly a metamorphic derivative of an ultramafic rock. Fahrig (1962, Fig. 2) showed two ultramafic lenses in the schist complex east of the Trough in the Griffis Lake area. Detailed mapping might reveal several more such lenses.

### *Lithology*

The writer's observations are confined to the northern half of the map-area, most notably the Ahr Lake area, and the following account reflects this limitation. However, Fahrig's special study (1953, 1962) on ultramafic rocks of the central Labrador Trough was performed on a sill of this belt only a few miles southeast of Retty Lake. Therefore, much of his data is undoubtedly applicable to ultramafic sills of this area and reference is made to this work.

#### *Megascopic Description*

The sills are composed mainly of chocolate-brown weathering, dense black serpentinite or serpentinized peridotite. Generally the rock is either massive or thinly sheeted parallel with the sill walls. In one place, near the base of a sill, compositional layering, also parallel with the sill walls, is distinct but not conspicuous. It dies out a short distance above this exposure and was not seen elsewhere.

Fahrig (1962, p. 4) noted a zonal arrangement in the ultramafic sills of the Griffis Lake area as follows:

1. Outer border zones — or actinolite zones
2. Intermediate zones — highly serpentinized
3. Central zone — containing fresh olivine

He also observed (1962, p. 2) that the ultramafic sills are typically overlain by a layer of meta-gabbro.

In the Ahr Lake area actinolite[1]-rich zones, ranging from 30 to more than 100 feet in thickness, are invariably present along the margins of the ultramafic sills. The interiors of the sills were not further subdivided, but rock-types representing the other two zones of Fahrig's were noted separately in places. Typically the basal actinolite-rich zone grades, in the manner of a chilled selvedge, from fine-grained adjoining the contact to gabbroic-textured a few feet into the sill.

[1] Actinolite is used here to denote the tremolite-ferrotremolite series and is not restricted to only the portion of the series that is demonstrably actinolite.

Farther from the margin it blends into massive serpentinite. The transition is readily apparent in outcrop where, even from a distance, the colour change from light greenish grey of the actinolite zone to deep brown of the serpentinite weathered surface is unmistakable. The upper marginal zones are not as well defined. At most of the exposed upper contacts the ultramafic sills are overlain by meta-gabbro and the dividing contact is extremely indistinct. Accordingly the actinolite-rich zone of the ultramafic sill is only distinguishable from meta-gabbro when appreciable serpentine is present.

Meta-gabbro commonly overlies ultramafic sills in Wakuach Lake map-area, but is only rarely found immediately below them. This feature is more prevalent than can be shown on the map of the area, for some of the meta-gabbro layers are too thin to be shown separately. Generally the ultramafic layer is thicker than the accompanying meta-gabbro layer and very commonly more than twice as thick. In places meta-gabbro only partly covers an ultramafic sill and the remainder of the sill is in direct contact with overlying country rock. The contact between ultramafic rock and overlying meta-gabbro, as mentioned above, is generally indistinct.

*Microscopic Description*

Typically, ultramafic rocks from the sill interiors are composed of pale green to colourless serpentine, which under crossed nicols is revealed as a closely packed array of pellets pseudomorphic after olivine. These pellets are generally $\frac{1}{2}$ to 1 mm in diameter. In a few places they contain eroded cores of olivine. Commonly ophitic augite crystals, 1 to 2 cm long, are present in varying stages of replacement to actinolite or serpentine. Magnetite generally forms 5 to 10 per cent of the rock. No orthopyroxene was observed by either the writer or Fahrig; if originally present it has been entirely serpentinized.

As the borders of the ultramafic sills are approached, the content of actinolite increases markedly; first as sparsely distributed, needle-like crystals that pierce the pseudomorphic pattern of the serpentine host at random, then as wide, shreddy-edged crystals that spread and intermesh until near the contact the original serpentine is eliminated, or nearly so. Albite, epidote–zoisite, and sphene are important constituents of the basal border zones at three widely separated places studied in the Ahr Lake area. They are confined to the outer 15 or 20 feet of the zones and constitute from about 15 to 40 per cent of the rock. Serpentine was not observed in any of the sections containing albite or epidote–zoisite. The relationship at the upper border zone is more complicated, for with the appearance of feldspar or its derivatives the rock is assumed to be part of the overlying meta-gabbro that is normally present.

## *Contacts*

All exposures of the lower contacts of ultramafic sills in the Ahr Lake area are at places where the sills adjoin argillaceous rocks of the Thompson Lake or Willbob Formations. At several places bleached zones a few inches wide containing scattered rosettes of actinolite crystals are immediately subjacent to the contact.

Farther away the sediments resume their black colour and actinolite is absent. These are interpreted as narrow metamorphic zones that have survived the effects of later regional metamorphism.

The problem associated with the upper contacts where they have generally been observed has already been discussed. Southeast of Chance Lake in the Ahr Lake area, meta-gabbro appears to be absent and the ultramafic sill is in direct contact with volcanic flows. However, the contact is indistinct and ultramafic rock passes into pillowed basalt across a poorly defined, recrystallized zone about 10 feet wide. Fahrig (1962, p. 7) noted that in the Griffis Lake area the lower contact of an ultramafic sill against volcanic rocks is remarkably gradational.

### *Origin*

Fahrig (1962, pp. 36–37) concluded that the Griffis Lake sill was intruded as a mush of olivine crystals in a gabbroic silicate liquid. Movement of the crystal mush during intrusion concentrated the olivine crystals in the centre of the fluid stream leaving a relatively higher proportion of the fluid phase adjoining the walls of the sill. Thus a chemical differentiation of the magma in a direction perpendicular to the fluid stream is brought about. This hypothesis admirably accounts for the various data assembled by Fahrig including:

1. the noteworthy concentration of Al, Ca, and Ti on the borders of the ultramafic masses relative to their interiors;

2. the preferred dimensional orientation of olivine crystals such that (010) tends to parallel the sill walls whereas enclosing augite ophites lack significant preferred orientation;

3. the calcic nature of the pyroxene, which implies derivation from a gabbroic liquid.

The presence of albite, epidote–zoisite, and sphene in the border zones of Ahr Lake ultramafic sills complements Fahrig's data on the distribution of Al, Ca, and Ti and can be similarly explained. Also, the 'chilled' borders noted at the lower contacts of the sills and the limited contact metamorphic effects observed in adjoining sediments are adequately accounted for by the hypothesis.

Fahrig (1962, p. 34) discussed the possibility that meta-gabbro overlying the ultrabasic sills is a result of differentiation in place, but rejected it for the following reasons:

1. Not all the ultramafic sills are overlain by gabbro.

2. The Cr content that is normally concentrated in pyroxene is abruptly reduced immediately above the ultramafic layer. If its distribution resulted from differentiation in place, it should have continued at a fairly high level in the original pyroxenes of the overlying gabbro.

3. The principal oxides show abrupt changes in value in passing from ultramafic rock to gabbro rather than a smooth transition. Some of the metal oxides such as $Na_2O$ are not represented in the ultramafic rock in pro-

portion to the amount of groundmass present, if the groundmass is assumed to have been a gabbroic fluid compositionally similar to overlying gabbro.

4. The gabbro is similar in composition to nearby meta-basalts and to other gabbro sills of the region.

The relationship of meta-gabbro and subjacent ultramafic rocks is explained instead by the assumption that the ultramafic material spread out beneath massive gabbro layers (Fahrig, 1962, p. 37).

The difficulties of applying the hypothesis of differentiation in place to the present problem, as pointed out by Fahrig, are formidable. Nevertheless the frequency with which meta-gabbro and ultramafic sills are found in the relationship described provides a strong incentive to explore further. The explanation offered by Fahrig is not entirely satisfactory as (a) thick successions of pillowed and massive flows should be as effective barriers as most gabbro sills, and (b) some of the overlying gabbro members are fairly thin and could be expected to possess little advantage as barriers in relation to other rocks.

Fahrig's (1962, p. 22) hypothesis of dynamic differentiation may offer a possible solution. Crystal-filled liquids, he pointed out, will have a higher viscosity than the same liquids, free of crystals and will be forced to migrate towards the centre of flow. Thus, the crystal fraction of a partly crystallized magma will be concentrated in the centre of the sill with the liquid phase at the boundaries.[1] However, the crystal-filled liquid will also have the greater density so that the maximum concentration of crystals can be expected somewhere below the centre line of the sill, leaving a wider zone of crystal-free liquid at the upper margin than at the lower. The position of the centre of flow, and therefore the relative thickness of upper and lower margins, would depend mainly on the velocity of flow and the difference in densities and viscosities of the two liquids[2]. Presumably a sill with an upper margin such as exists among the ultramafic sills of this area is quite possible.

The lack of an overlying gabbro member is not irreconcilable with this type of differentiation; in fact it may be a necessary adjunct. During intrusion of the crystal mush, liquid might be expected to be continually withdrawn from the system in the form of chilled margins until eventually the liquid phase is barely sufficient to act as a marginal lubricant. At this stage the upper margin would not likely be distinguishable as a separate unit.

Neither the abrupt change in the contents of Cr nor of the principal oxides in passing from ultramafic to gabbroic layers can be considered a serious challenge to this view of the differentiation. The boundary between crystal-rich and crystal-free liquids might be expected to be fairly abrupt and the resulting change in

[1] Note recent experimental work by Bhattacharji and Smith (1963, 1964).

[2] Experimental work by Bhattacharji during the winter of 1963–64 after this section was written demonstrated that this type of flow actually does take place. He found that a fluid charged with solids flowing through a sill-like conduit separates into three parts: a central region of liquid rich in solids and upper and lower bounding zones free of solids. When the flow velocity drops below a critical level the upper solid-free zone widens at the expense of the lower (Bhattacharji, personal communication).



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chemistry should reflect this. The Cr content of the ultramafic rock is present in both pyroxene and magnetite (Fahrig, 1962, pp. 24–25), neither of which continues into the overlying meta-gabbro. Magnetite appears to be represented by sphene, and pyroxene by actinolite. Thus the geochemical gradient between the two rocks, regardless of its original character, was probably steepened by metamorphism.

A similarity in chemical composition between meta-gabbro overlying the ultramafic rocks and other gabbros and basalts of the region suggests consanguinity of all early geosynclinal igneous rocks, but does not preclude the relationship between meta-gabbro and ultramafic rocks suggested.

Finally, the complete absence of feldspar within the ultramafic unit itself is a serious obstacle to any theory of ultramafic intrusion that depends upon a gabbroic liquid phase. The low NaO content of the ultramafic rock is a manifestation of this problem. If the interstitial liquid were gabbroic as suggested by Fahrig (1962, p. 23), and agreed with here by the writer, some plagioclase would be expected in the ultramafic layer. Its absence might be partly accounted for by the presence of augitic pyroxene but some difficulty remains. Hess (1960, pp. 109–113) postulated a process of diffusion to account for the lack or near lack of plagioclase in ultramafic layers, or alternately ferromagnesian minerals in anorthositic layers, of the Stillwater Complex. Interstitial liquid would crystallize a mineral of the same composition as the adjoining accumulative crystals, thus enriching the fluid in components not yet crystallized. The resulting chemical gradient between the interstitial fluid and the fluid overlying the crystal mush layer would act to remove the excess components in the interstitial fluid. The degree to which it succeeded would depend upon the thickness of the crystal mush layer and the rate of accumulation. A similar process might be suggested to explain the lack of plagioclase in the ultramafic layers of Labrador sills, although they are admittedly far thicker than the distances at which diffusion under these conditions might be considered effective (cf. Hess, 1960, pp. 146–148). A completely satisfactory explanation of this point, therefore, has not yet been reached but the possibility of arriving at such an explanation does not seem entirely remote.

## Glomeroporphyritic Gabbro

### *Name, Distribution, and Thickness of Sills*

The glomeroporphyritic gabbros are found throughout much of the length of the Labrador Trough and are variously known as blotchy gabbro, mottled gabbro, leopard rock, and anorthositic gabbro. They are reported from areas as far apart as the Menihek Lake area (Frarey, 1961) east of Schefferville and the Harveng Lake area (Bergeron, 1956) just south of Leaf Bay. In Wakuach Lake map-area, glomeroporphyritic gabbro is confined to a single belt of sills, 4 to 6 miles wide, on the west side of Murdoch Lake valley.

Glomeroporphyritic sills in the Ahr Lake area are as much as 1,500 feet thick (approximate). However, this thickness includes a substantial layer of non-porphyritic gabbro at the centre of the sill. The thickest continuous layer of g'omeroporphyritic gabbro found is less than 1,000 feet thick.

*Description*

Detailed examination of glomeroporphyritic gabbros was mainly confined to the Ahr Lake area from where most of the following data derives.

The rocks are coarse-grained feldspathic gabbros regularly spotted with roughly spherical aggregates of altered feldspar 6–15 cm across (Pl. IX). Generally one to six large aggregates occur per square yard of outcrop. The feldspar is highly altered labradorite and is set in a sparse groundmass of mainly dark amphibole, pyroxene, and chlorite.

At least three of the glomeroporphyritic sills in the Ahr Lake area appear to be composite; sinuous layers of non-porphyritic gabbro similar to the normal gabbros of the region are present at the centres of the sills or finger into them from the side. The best example is the westernmost glomeroporphyritic sill in the Ahr Lake area where the gabbro layer at its centre thins and thickens and swings from side to side along its length. In one place it lenses out, then reappears farther along. In the Hyland Lake area, Kavanagh (1954) also noted the symmetrical distribution of glomeroporphyritic gabbro on either side of normal gabbro and concluded that the sills there are composite.

The contacts between glomeroporphyritic and intra-sill normal gabbros are abrupt but not sharp. Glomeroporphyritic gabbro passes into normal gabbro at a surface that marks the limit of coarse feldspathic aggregates. Within the normal gabbro close to the contact are scattered clots and large phenocrysts of feldspar, which may have been wrenched from the adjoining rock. Isolated lenses of glomeroporphyritic gabbro roughly parallel with the sill margins are encountered within the interior gabbro; one such lens contains a vaguely defined gabbro dyke that appears continuous with surrounding gabbro.

The borders of glomeroporphyritic sills, chilled against adjoining rock, are dark basalts without visible phenocrysts. Within 2–10 feet of the contact the rock grades into dark, medium-grained gabbro with sparsely distributed feldspar phenocrysts and aggregates of phenocrysts. In a marginal zone 30–75 feet wide the rock is predominantly gabbro, but inward from here the number and size

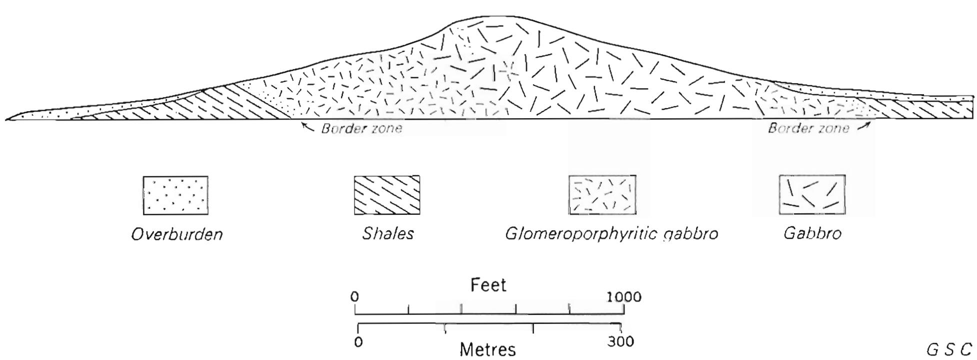



FIGURE 16. Cross-section of composite sill, glomeroporphyritic and normal gabbro. About 3 miles east of Hurst Lake.

of phenocrysts and aggregates increases sharply, until they coalesce to form typical glomeroporphyritic gabbro. In at least one place a laminar arrangement of phenocrysts occurs at the margin of the main glomeroporphyritic mass. A cross-section of a composite sill just northeast of Hurst Lake is shown on Figure 16.

In a few places glomeroporphyritic gabbro seems to be represented by one or more layers, a few inches wide, of closely packed feldspar phenocrysts and clusters of phenocrysts, or by attenuated alignments of single feldspar phenocrysts or clots in non-porphyritic gabbros. The layers are parallel with attitudes in adjoining rocks and are presumably the result of laminar flow. They are generally on strike with mappable units of massive glomeroporphyritic gabbro and may represent either (1) an incipient phase of its development, or (2) the debris of such a rock concentrated along flow layers of the invading, non-porphyritic gabbro.

#### *Age*

Glomeroporphyritic gabbros pre-date non-porphyritic gabbros with which they are associated in composite sills, but the lack of chilled contacts between the two rock types indicates that they were not separated by a significant time interval. Probably the non-porphyritic gabbros intruded the warm, semiconsolidated centres of glomeroporphyritic sills where resistance to emplacement would be minimal. These later gabbros of the composite sills are similar to other gabbros of the area and are probably part of the major gabbroic invasion, which has left the region pervaded with sills. However, the glomeroporphyritic gabbro does not necessarily pre-date all the normal gabbro sills. These may have been intruded over a sufficiently long period that they completely overlapped the time during which glomeroporphyritic sills were emplaced.

### Normal Gabbro

#### *Distribution and Thickness of Sills*

The normal or common gabbro of the region forms numerous, closely spaced sills, which intrude sedimentary rocks between Wakuach and Murdoch Lakes. Individual sills range from about 100 to about 2,000 feet in thickness. Thicknesses of 1,000 to 1,200 feet are not uncommon, but the average thickness is about 500 or 600 feet. Total aggregate thickness of normal gabbro sills is estimated to be about 15,000 to 20,000 feet. The length of individual sills is unknown, but some can be traced for at least 30 miles.

#### *Description*

Gabbro sills commonly dip at angles less than 45 degrees and so form cuesta-shaped ridges with escarpments characteristically ribbed by columnar jointing. Generally a set of pronounced planar joints intersect the columnar joints approximately at right angles, and provide a ready surface for measuring sill attitudes.

The gabbros are in general fairly unaltered, grey-green, medium-grained rocks. Within single sills they may range from sparsely porphyritic and ophitic rock types in the lower parts to coarsely pegmatitic gabbros in the upper. Veins of granite ranging from 2 inches to about 2½ feet in thickness are found in the upper parts of a few sills near Lac Cramolet. Three to 10 feet of chilled marginal phase have been observed at both upper and lower contacts of several sills. Calcite amygdules and argillite xenoliths were found in the upper chilled margin in a few places.

### Meta-gabbro

#### *Distribution and Thickness of Sills*

The meta-gabbro sills are found east of the Walsh Lake fault and throughout the width of the area from south to north boundary. The main belt of sills intrudes, and partly obscures, the Thompson Lake Formation. East of this a number of independent sills are found in the Willbob Formation where they generally intrude sedimentary interlayers. The thickest single meta-gabbro sill known is about 1,500 feet thick.

#### *Description*

The rock is mostly massive, medium-grained gabbro composed essentially of albite, epidote–zoisite, actinolite, and chlorite. A polkadot-like texture, formed of discrete, rounded masses of light to dark green actinolite in a chalky white matrix of epidote–zoisite and albite is characteristic. Elsewhere the amphibole forms irregular interconnecting patches. The main megascopic distinction between meta-gabbro and normal gabbro is the greater contrast between the dark and light components of the former. Plagioclase and pyroxene in normal gabbro are similar in colour tone whereas the corresponding light and dark minerals in meta-gabbro are sharply contrasted. The meta-gabbros are rarely foliated. Chilled zones are found at all exposed margins of meta-gabbro sills indicating that at least the gross igneous textures have survived. Two groups of meta-gabbro sills, viz. layered meta-gabbros and leucometa- and melanometa-gabbros, are of sufficient interest to require special description even though they are relatively unimportant quantitatively.

*Layered Meta-gabbros*

A sill intrusive into the Murdoch Formation about a mile northeast of Lac d'Argent contains a number of shallow to moderately east-dipping layers of very light coloured meta-gabbro in its upper part. These range from about 6 inches to a much greater thickness and appear to parallel the sill walls. The leucocratic layers are characterized by elongate blue-green amphibole, coarse irregular patches of garnet, and considerable quartz. A thin section contained an estimated 40 per cent quartz. A similar type of sill, also intrusive into the Murdoch Formation, on the east side of Walsh Lake has been described by Frarey (1952, p. 6) and Kavanagh (1954, pp. 89–94). According to Kavanagh it contains aplitic layers 5 to 10 feet thick along its northeastern (upper) side. These may be composed of up to nearly 50 per cent quartz and micropegmatite in addition to deep blue-green hornblende and chlorite. Most likely the acidic layers in both sills are pegmatitic schlieren formed during crystallization of the sills.

*Melanometa-gabbro and Leucometa-gabbro*

Two distinctive types of meta-gabbro termed melanometa-gabbro and leucometa-gabbro are found in the eastern part of the Ahr Lake area. Both are embodied in the same sills and also occur separately. Typical melanometa-gabbro consists of interlocking, equidimensional crystals of amphibole 0.5–3 cm in diameter. Few other minerals are apparent in hand specimen. Leucometa-gabbro in its most spectacular development is light cream to white, sparsely dotted with greenish blebs of amphibole. In other places it is bluish grey and inconspicuously marked with slightly darker amphibole crystals. Melanometa-gabbro occupies the lower part and leucometa-gabbro the upper part of two separate sills in the Ahr Lake area, one, just west of a string of ponds (Valley Ponds)[1] about 2½ miles northeast of Lac Harvut, and the other extending southeast from Keato Lake. In the former the upper member extends several miles beyond the lower member. The rock-types grade into one another across a narrow zone in the Keato Lake sill[2], and presumably the Valley Ponds sill. Thus they are essentially bipartite sills with a narrow transition zone.

Typical melanometa-gabbro is found without associated leucometa-gabbro in a faulted sill about 2 miles northeast of Lac Harvut. Faulting has complicated the boundaries; hence melanometa-gabbro may not necessarily have been emplaced separately. Leucometa-gabbro forms much of the sill about half a mile northeast of Lac Harvut, but does not seem to have a melanometa-gabbro base.

## The Younger Complex

The assemblage here termed the Younger Complex is composed of a variety of metamorphic and plutonic rocks that underlie the northeastern corner of the map-area. It is not part of the Labrador Trough as defined in this report, but may, in part, have belonged to the Labrador geosynclinal assemblage. The main justification for grouping these rocks under a single, unifying title is that they have shared a degree of metamorphism sufficient to obscure, if not entirely erase, their primary textures and structures. The relative ages of units within the complex are unknown and their contacts were not observed. They will therefore be taken up as they appear in succession eastward from the Trough.

### Quartz–Feldspar–Biotite Schist

#### *Lithology*

The quartz–feldspar–biotite schists are light grey, thinly laminated rocks that are remarkably uniform in appearance throughout the region in which they were observed. The laminations are most commonly 1 to 5 mm thick and are generally inconspicuous. Foliation, due to the orientation of biotite flakes parallel

[1] 'Valley Ponds' was a name given these lakes by the writer (1958) for purposes of reference. It is not a formally accepted geographic name. However, since the name Valley Ponds sill has already been used in the literature (Baragar, 1960) and no other reference point is close at hand, its use is continued informally in this memoir.

[2] Previously called Protecto Lake sill (Baragar, 1960).

with the lamination, is marked and gives the rock a ready fissility. Immediately east of the Labrador Trough, the rocks are readily recognizable as meta-greywackes and carbonaceous slates, commonly with graphite-coated, slickensided surfaces. Within one or two miles of the fault-boundary the metasediments assume the appearance typical of the quartz–feldspar–biotite schist of this unit.

The rocks in thin section are characteristically composed of a fine-grained mosaic of equidimensional quartz and untwinned plagioclase crystals with bedding or layering defined by a variation in the biotite content. Muscovite normally accompanies biotite but is subordinate to it. Accessory minerals are sphene, opaque iron oxide, apatite, zircon, and tourmaline; epidote is rare. Minor chlorite can generally be found partly replacing biotite. Modest amounts of garnet are found locally about 10 miles from the east boundary of the Trough. The grain size of plagioclase and quartz averages about 0.1 mm, and the length of biotite crystals, about 0.2 mm. Biotite and muscovite almost invariably show a strong preferred orientation parallel with the lamination, but in one specimen, two preferred orientations crossed each other at an angle of about 30 degrees. The minerals are mainly fresh and clear.

Estimated compositions of a number of specimens fall within these percentage ranges:

| | |
|---|---|
| quartz and plagioclase | 60–75 per cent |
| biotite (brown) | 20–35 per cent |
| muscovite | 1–15 per cent |
| iron oxide and sphene | minor–3 per cent |
| apatite, tourmaline, zircon | minor |

Quartz and feldspar are difficult to distinguish in unstained thin sections. Volumetric analyses of three stained thin sections are as follows:

| | BL9-99 | BL9-115 | BLH9-141 |
|---|---|---|---|
| quartz | 33 per cent | 30 per cent | 31 per cent |
| plagioclase | ($An_{15}$) 32 per cent | ($An_{20}$) 29 per cent | ($An_{23}$) 17 per cent |
| biotite | 31 per cent | 27 per cent | 34 per cent |
| muscovite | — | 13 per cent | 16 per cent |
| others | 4 per cent | 1 per cent | 2 per cent |
| Point counter, no. of points: | 1536 | 1114 | 1000 |

Specimen BL9-99 is from near the south tip of Lac Savalette about 1½ miles east of the Trough boundary; BL9-115 is from the shore of Rivière Savalette, about 8 miles from the Trough boundary; and BHL9-141 is from 2½ miles northeast of northern Lac Vannes, also about 8 miles from the Trough boundary. Potash feldspar is present in minor amounts in BLH9-141, but was not seen in the other two sections.

### *Internal Structural Relations*

Meta-greywackes and carbonaceous slates adjoining the eastern boundary of the Trough are more disturbed than the same unit farther east. Dips to the northeast range from 10 to 50 degrees, but are most commonly about 20 to 30 degrees. Southwest of Lac Savalette the rocks are involved in several tight folds that are steeply overturned to the west and plunge gently southward. Small drag-folds are common, and as seen in thin section the laminations are finely crumpled. The disturbed zone is presumed to be a manifestation of the major thrust fault thought to form the eastern boundary of the Trough.

From about 2 to 14 miles east of the Trough in the vicinity of Rivière Savalette the schists incline gently eastward in broad undulations. Dips range from 5 to 15 degrees. Eastward they steepen to 20 or 25 degrees as they become involved in a sharp downward flexure that is more severely marked in the rocks farther east.

### *Origin*

Recognizable meta-greywackes and slates immediately east of the Trough are presumably relics of the primary rock. Nothing in the composition and structure of the schists of this unit is incompatible with such an origin.

## Amphibolites

The amphibolites are a varied group of mafic rocks in which hornblende is a prominent constituent. They range from fine grained to coarsely phaneritic and from predominantly hornblendic to mainly feldspathic. They are most likely the metamorphic equivalents of mafic igneous rocks.

The western amphibolitic unit, about 15 miles northeast of the Trough, is a mixed assemblage of interlayered amphibolites, biotite schists, and quartz-rich, carbonaceous sediments. The amphibolite layers are massive to well foliated, dark green rocks ranging in grain size from several millimetres to nearly aphanitic. Leucocratic components are generally inconspicuous, but in places they form thin discontinuous seams that emphasize the foliation. Light to dark green hornblende composes between 40 and 85 per cent of the rock and plagioclase forms most of the remainder. Opaque iron oxides, iron sulphides, and sphene are ubiquitous accessory minerals, and minor garnet occurs locally. Sedimentary and schist layers are subordinate to the amphibolites but appear to be distributed throughout them. One variety is essentially a fine-grained carbonaceous quartzite which, locally at least, shows distinct crossbedding. Other layers are highly carbonaceous slate. The schist is virtually identical with quartz–feldspar–biotite schists farther west.

The eastern amphibolite unit is composed essentially of mafic gneisses with a notable leucocratic component. The grain size ranges from one to several millimetres, and the rocks vary from rudely foliated to well foliated. A variety that outcrops along the outlet stream from Lac Champdoré is a medium- to coarse-grained augen gneiss in which clots of feldspar, 5–10 mm across, are evenly distributed through a dioritic groundmass. Near the eastern side of the amphibolitic unit the mafic gneisses contain seams, masses, lenses, and veins of pink granitic material. The mafic gneisses are composed mainly of plagioclase, deep blue-green hornblende, varying amounts of (though generally subordinate) biotite, and epidote. Sphene and iron oxide are accessory minerals. The augen gneiss and some of the mafic gneisses on the east side of the member contain notable amounts of microcline and quartz.

The foliation in both belts of amphibolitic rocks dips moderately (40–50° mainly) to the northeast. No folds were observed.

## Garnet–Staurolite–Mica Schist

### *Lithology*

The unit is a layered assemblage of garnet–staurolite–mica schist, leucocratic quartz-rich metasediments, quartz–feldspar–biotite schists, and rarely hornblende schists. Garnet–staurolite–mica schist is the most abundant rock type, followed by the leucocratic metasediments. The layers range from half an inch to several feet in thickness.

Garnet–staurolite–mica schist is typically a crudely foliated, coarse-grained, nodular rock with a sufficiently high biotite content to give it a fairly dark appearance. Nodules are garnet crystals 2 to 20 mm in diameter. Well-formed, prismatic staurolite crystals range from less than a millimetre to 2 or more cm in length. In detail the rock is roughly laminated with discontinuous layers, a millimetre to several millimetres thick, alternately rich in biotite or quartz, feldspar, and muscovite. Typical specimens examined in thin section are seen to be dominated by mica. Muscovite forms from 30 to 60 per cent of the rock, and biotite 10 to 20 per cent. Other minerals present in approximate order of abundance are quartz, staurolite, garnet, and plagioclase. Kyanite is a minor constituent in two specimens. Accessory minerals are apatite, zircon, and an iron oxide mineral. The micas show a moderately preferred orientation parallel with the compositional layering. In some specimens bedding planes are marked by closely spaced crenulated strings of dusty opaque material and oriented slivers of muscovite that pass undisturbed through coarse plagioclase, staurolite, and biotite metacrysts. The foliation commonly sweeps around garnet metacrysts.

Leucocratic quartz-rich layers are characteristically even-grained rocks with weakly defined laminations. Biotite is a minor constituent. A typical specimen from the shore of Rivière Savalette 17½ miles northeast of the Trough yielded the following volumetric analysis:

| | |
|---|---|
| quartz | 73 per cent |
| plagioclase ($\pm An_{20}$) | 17 per cent |
| biotite | 5 per cent |
| garnet | 4 per cent |
| accessory minerals | 1 per cent |

In other localities the quartzite layers are similar, but commonly the feldspar content appears to be higher.

Quartz–feldspar–biotite schist interlayers are similar to the rocks of the same classification farther west. Hornblende schists are generally dark, fine-grained rocks with a pronounced alignment of hornblende needles. One thin section examined contained 80 to 85 per cent blue-green hornblende, 15–20 per cent quartz and untwinned plagioclase, and minor garnet, sphene, and iron oxide. The garnets in this specimen show clear evidence of rotation during growth. Garnet metacrysts contain S-shaped laminations that pass into the rock foliation on either end, and the crystals are partly or completely surrounded by quartz haloes presumably formed in the "strain shadows" during rotation.

#### *Origin*

The compositions of rocks of this unit and their pronounced layering make it evident that they were derived from a succession of sedimentary rocks, mainly argillites, quartzites, and arkoses. Beds of impure quartzite or greywacke are presumably responsible for the layers of quartz–feldspar–biotite schist found in the assemblage. The hornblende schists may have been derived from rare, interbedded volcanic rocks.

### Tremolite–Actinolite Rock

#### *Lithology*

Light grey, massive tremolite–actinolite rock outcrops intermittently across the regional strike for several hundred feet. It is shown on Wakuach Lake map-sheet as a lens-shaped body adjacent to Rivière à la Baleine, but its actual extent is unknown. The rock is fine grained except for sparsely scattered coarse tremolite crystals as much as 2 cm long. Commonly it has a rusty, pitted surface due to the weathering of iron carbonate.

The rock appears in thin section as a mass of randomly oriented, intermeshed, shreddy tremolite–actinolite crystals interspersed with irregular grains of carbonate and opaque iron oxide. The amphibole is colourless in thin section. The composition of the rock is estimated as follows; tremolite–actinolite, 85 per cent; carbonate, 10 per cent; iron oxide, 5 per cent.

#### *Origin*

Tremolite–actinolite rock is devoid of both feldspar and its metamorphic derivatives. Compositionally, then, it is an ultramafic rock; moreover it is almost identical with tremolite–actinolite rocks derived from known ultramafic rocks in the Labrador Trough. Fahrig (1962, Fig. 2) located two ultramafic bodies within the metamorphic complex east of the Trough. Thus, the presence of another ultramafic body in this environment is not unexpected and, together with those noted by Fahrig, may represent the remnants of a second ultramafic belt associated with the Labrador geosyncline.

## Granodiorite Gneiss

### *Lithology*

A varied group of gneisses occupies the northeastern corner of the map-area. The most characteristic type is a grey, medium-grained granitic rock with a distinct foliation marked by flat streaks and clots of biotite and locally hornblende. These are commonly aligned and give the rock a distinct lineation. In one place, where it was particularly noted, the lineation was approximately horizontal. Other types include pink granular rocks containing mainly quartz and feldspar with very little dark mineral and fine-grained, grey, massive rocks with modest to negligible amounts of biotite. Some of these show vaguely defined compositional layering. Irregular veins or dykes of pink to white, fine-grained aplite, rarely more than a few inches wide, cut the gneisses at a number of places, but are not plentiful.

In thin section the gneisses show allotriomorphic-granular texture with a moderate degree of preferred orientation in biotite and hornblende. Quartz and feldspar crystals are roughly equidimensional and generally range from half a millimetre to 2 mm in diameter. Microcline is almost invariably subordinate to plagioclase and is generally interstitial. The feldspars are well twinned and clear. Plagioclase compositions in various specimens range between about $An_{15}$ and $An_{30}$; thus the gneisses are mainly granodiorite and rarely quartz monzonite. Estimated compositions from a variety of specimens range as follows:

| | |
|---|---|
| quartz | 20–50 per cent |
| microcline | 1–25 per cent |
| plagioclase | 25–55 per cent |
| biotite, hornblende } mainly biotite | 5–20 per cent |

Biotite ranges from brown to olive and hornblende is generally deep blue-green. Accessory minerals are apatite, zircon, epidote, sphene, and iron oxide.

A belt of quartz–feldspar–biotite schist similar to that found along Rivière Savalette appears in the granitoid gneisses about 3½ miles northeast of Rivière à la Baleine. The layer of schist is a minimum of 500 feet wide; it was traced for a mile, and may extend at least an additional 3 miles along strike. It is friable and highly crenulated; locally granitic-like material appears along schist planes.

### *Internal Structural Relations*

Foliation in the gneisses parallels the strike of the Trough and dips are mainly 50 to 70 degrees northeast.

### *External Structural Relations*

The contact between granodiorite gneisses and amphibolite to the west is not exposed. Amphibolites just west of the contact seem to contain a higher content of aplites and interspersed quartz and feldspar than those farther west; conversely granodiorites just east of the contact appear to be more mafic than those farther east. These are impressions and should be interpreted with caution, but they suggest that the contact is gradational.

### *Age*

Granodiorite gneiss from a point 3½ miles northeast of Rivière à la Baleine yielded a K–Ar age of 1,635 m.y. (Lowden, 1961, p. 74). This is considerably younger than the age of 2,365 m.y. obtained from the Basement Complex (reported earlier in this memoir) west of the Labrador Trough.

### Discussion

The relationship of the Younger Complex to rocks of the Labrador Trough is unknown. Recent writers, notably Bergeron (1957, p. 104) and Fahrig (1957, pp. 115–116), have concluded that the metamorphic complex lying east of the Trough is equivalent in part to the Trough assemblage. Following are considerations pertinent to the problem in Wakuach Lake map-area.

1. A succession of metamorphic mineral assemblages representing increasing grades of metamorphism begins in the Trough and extends into the Younger Complex. Thus metamorphic effects in both sets of rocks appear to belong to the same period of metamorphism.

2. Primary rocks from which the Younger Complex was derived, as far as can be reasonably interpreted, were greywacke–shales, mafic volcanic rocks, ultramafic rocks, gabbros, argillites, quartzites, and arkoses. These are the types of rocks found in the Knob Lake Group, but also in most other geosynclinal assemblages.

3. The westernmost outcrops of the quartz–feldspar–biotite schist unit of the Younger Complex are readily recognizable greywacke and slate, little changed in appearance from that well within the Trough.

Thus evidence from this area strongly favours the views of Bergeron and Fahrig. An alternative explanation would be difficult to support. If the schists were assumed to be part of a pre-Kaniapiskau succession, which is the most likely alternative, the older group must have nearly coincided with the younger, for it seems to be continuous on the east side of the Trough, but is not identified on the west side. The work of Stockwell (1961) and the Geological Survey's age determination laboratories on rocks of the Canadian Shield has reduced the possibilities of postulating random occurrences of orogenic belts in space and time. A belt of schists distributed along the Labrador Trough cannot be part of the east-west trending orogenic belts of the preceding Superior Structural Province.

The granodiorite gneisses must be excluded from the arguments above for they are not demonstrably derived from supercrustal deposits. In fact, there is some evidence to suggest that they may belong to the Basement Complex. Basement gneisses underlie the eastern contact of the Labrador Trough in the Michikamau Lake area (Wynne-Edwards, 1960) and in the Brochant de Bonnard area (Bergeron, 1957) at both ends of the Trough. In effect, the Basement

Complex can be traced around or through the ends of the Trough only to disappear in the hinterland where deformation is most intense. It seems reasonable to postulate that the Basement Complex actually becomes involved in the deformation and loses its separate identity. Locally it may still be recognizable by special techniques. West of Fort Chimo, Beall and Sauvé (1960, p. 250) obtained a whole-rock, rubidium–strontium age of 2,470 million years from gneisses on the east side of the Trough. Potassium–argon ages of these gneisses and adjoining metamorphic rocks of the Trough ranged from 1,400 to 2,060 million years. The authors concluded that the gneisses are remetamorphosed basement rocks. Gradational contacts between the schists or amphibolites and the gneisses such as described by Fahrig (1957, p. 115) and noted above do not affect the argument. Wynne-Edwards (1960) described gradational contacts between Kaniapiskau and known basement rocks where deformation seems to be far less intense. The presence of a fairly thin member of quartz–feldspar–biotite schist evidently interlayered in granodiorite gneisses (mentioned on a preceding page) is readily explained as a remnant of the supercrustal group infolded into the basement. It is less readily accounted for by assuming granodiorite gneiss to be the result of either granitization or intrusion.

## Diabase

A single diabase dyke, about 12 feet wide, cuts iron-formation about 4½ miles directly east of Boundary Lake and just south of Leroy number 1 iron ore deposit. The dyke strikes slightly east of north and dips about 45 degrees west. It evidently belongs to the family of north-striking diabase dykes shown on Menihek Lakes map-sheet to the south (Frarey, 1961). The rock in thin section is dominated by a criss-crossing meshwork of plagioclase laths with subordinate interstitial olivine, pyroxene, and opaque iron minerals. Ten to 15 per cent olivine is present. Plagioclase is about $An_{70}$ and the pyroxene appears to be mainly or entirely titaniferous augite with a fairly small optic angle.

# *Chapter IV*

# PETROLOGY OF THE BASALTIC ROCKS

## Introduction

Basaltic rocks in the sense used in this chapter refer to rocks derived from basaltic magmas. Thus, the term covers gabbros as well as lavas. The petrography of the flows is fairly simple and has been dealt with adequately in the preceding chapter under the various formation headings. The petrography of the sills and the petrology of sills and flows together require more detailed examination.

The petrology of basaltic rocks in the Ahr Lake area was the subject of a previous paper (Baragar, 1960). Since the data for that work was collected the writer has had the opportunity to extend his observations to basaltic rocks throughout much of Wakuach Lake map-area. Fortunately the Ahr Lake study requires little revision as a result. Granitic veins, not found in the Ahr Lake sills, occur in the upper parts of sills elsewhere in the area, thus completing the full range of differentiation. Otherwise the characteristics of Ahr Lake sills appear to be generally applicable to sills throughout the area. There is no need to repeat here much of the detail carried in the previous publication. Nevertheless it will be necessary to present again the main features of that work so that old and new data can be worked into a coherent account based on the area as a whole.

## Normal Gabbros

### Petrographic Description

Most of the thicker gabbro sills are well differentiated. The resulting products found at corresponding levels from sill to sill throughout the area are remarkably similar to each other, but reasonably distinctive from those at other levels. Thus, differentiation of the normal gabbro sills in the area can be represented by a single, generalized diagram (Fig. 17). Similarly the differentiation products can be grouped into a few major types. These are described in the following sections.

#### *Porphyritic Gabbro*

Porphyritic gabbro is found at the base of a number of sills but is not widespread. The most notable occurrences seen by the writer are in the belt of sills immediately west of Murdoch Lake; elsewhere in the map-area the rock was noted in only three or four places. The rock is dark with sparse plagioclase pheno-

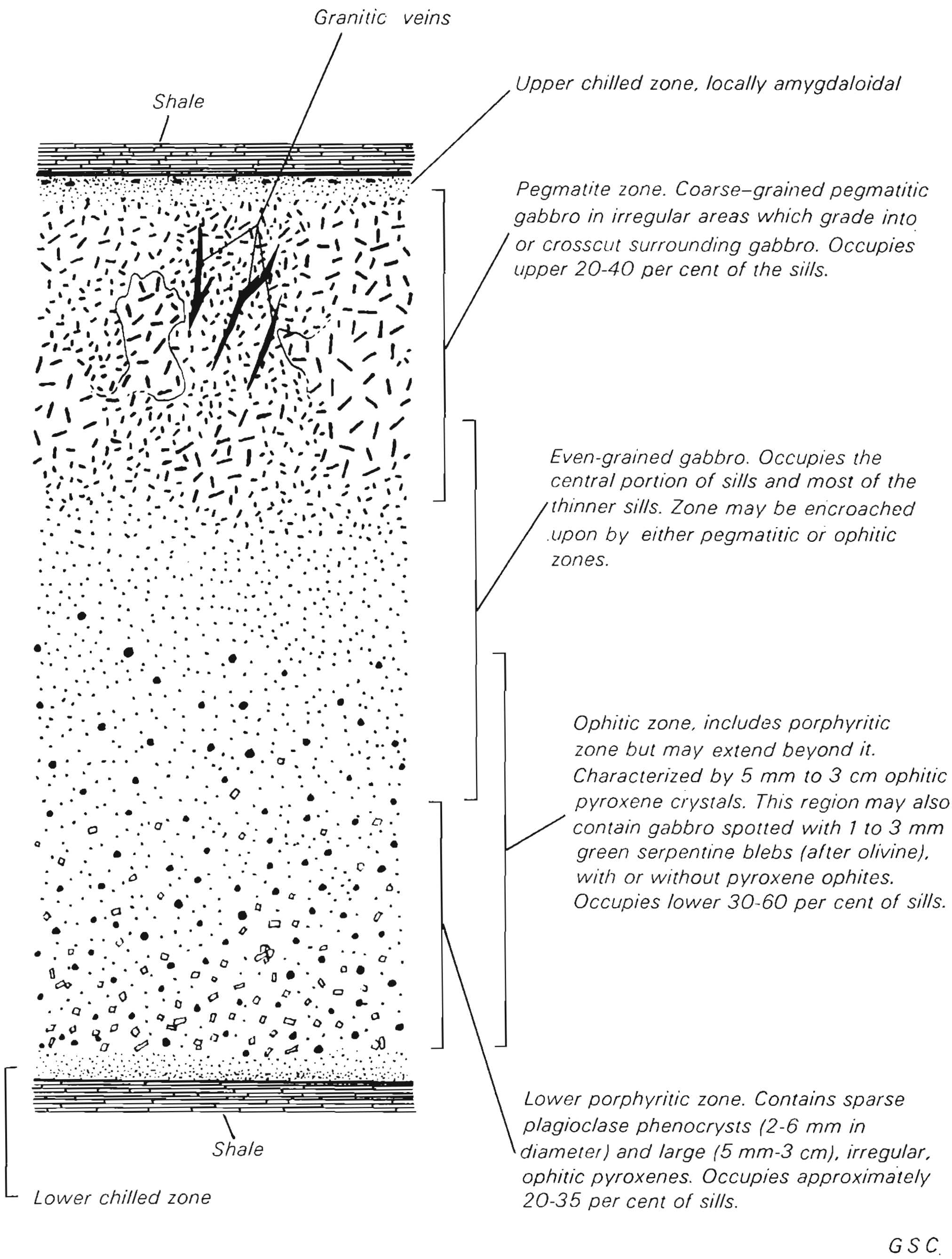



**FIGURE 17. Generalized cross-section of a differentiated gabbro sill.**

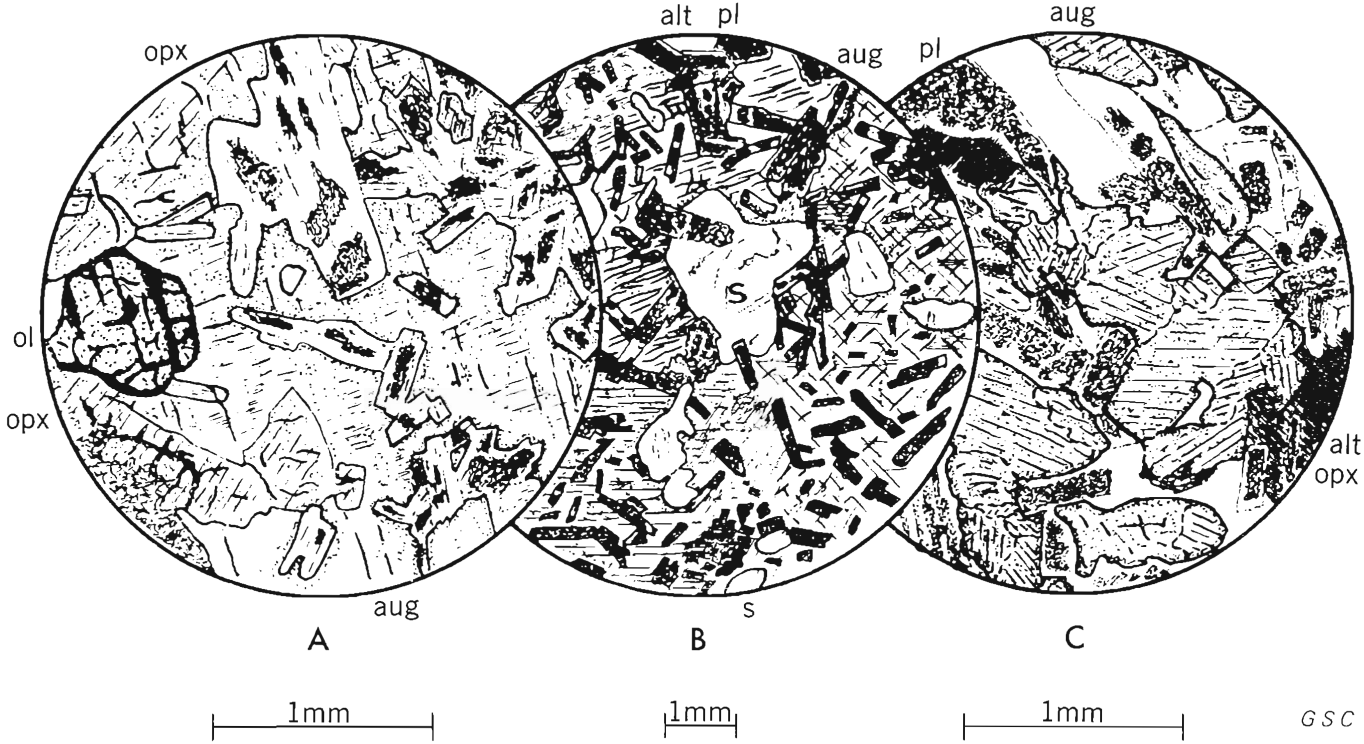



FIGURE 18. Porphyritic, ophitic, and even-grained gabbros.

A. Porphyritic gabbro showing olivine and plagioclase embedded in augite and orthopyroxene. Plagioclase is partly saussuritized. Specimen B184.

B. Ophitic gabbro containing serpentine pseudomorphs after olivine. Much of the field is occupied by large ophitic augite crystals that enclose numerous saussuritized plagioclase laths in addition to serpentine pseudomorphs. Specimen B295.

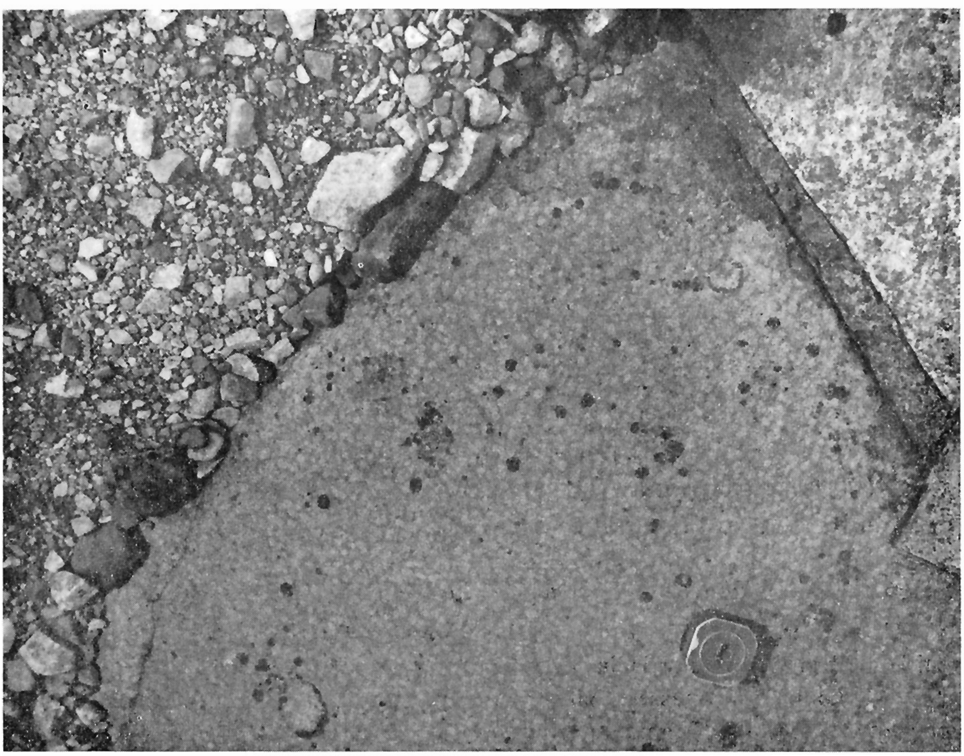
C. Even-grained gabbro showing subophitic augite and plagioclase. Minor orthopyroxene is present but severely altered. Specimen B295.

ol—olivine; opx—orthopyroxene; aug—augite; pl—plagioclase; alt—altered; s—serpentine crysts, 2–6 mm across, and ophitic pyroxenes as much as 3 cm in diameter. Two to a dozen phenocrysts may be visible in a hand specimen. Commonly the rock is spotted with serpentine blebs derived from olivine. Chemical analyses and norms are given in Table IX and volumetric analyses in Table IV.

Olivine varies from 4 to 18 per cent, combined pyroxenes from 20 to 30 per cent, and plagioclase from 50 to 60 per cent. 'Ores' are generally minor. Plagioclase phenocrysts, now mainly saussurite, constitute an estimated 1 to 10 per cent of the rock. Typical porphyritic gabbro is shown on Figure 18A. Many orthopyroxene crystals contain oriented exsolution streaks and blebs of augite, which may persist even when the crystal has been entirely replaced by serpentine. Augite with recognizable exsolved orthopyroxene has not been observed. 'Ores' are generally brownish masses of sphene containing blebs or triangularly arranged lamellae of ilmenite. Some masses reach 1 mm in diameter and are ophitic or subophitic.

### *Ophitic Gabbro*

Ophitic gabbros are widespread and appear to occupy the lower part of all sills that show any measure of differentiation whatsoever. The zone includes the porphyritic member at the base of the sill where this is present. In outcrop the



*W.R.A.B. 4-5-59*

PLATE X. Outcrop of ophitic gabbro showing mottled effect caused by thickly distributed augite ophites. The rock is near the base of a sill about 2 miles west of Lac de l'Axe.

rock has an irregular surface owing to differential weathering of plagioclase and ophitic pyroxenes and in places the ophitic texture is marked (Pl. X). The fresh surface is mottled greyish green, but the characteristic ophites are normally inconspicuous unless their cleavages reflect the light. Several ophitic pyroxenes up to 3 cm in diameter may occur in a hand specimen. Olivine, or its serpentine pseudomorph, is generally present but not as an essential constituent. In a few places the rocks are thickly spotted with brownish green serpentine blebs derived from olivine.

In thin sections the rock is seen to contain large, equidimensional augites packed with randomly oriented plagioclase laths and occasional serpentine patches after olivine. The large augite crystals occur singly and are separated by areas of even-grained, subophitic gabbro. A second pyroxene is rare, but evidence of its former existence in the form of serpentine patches, some with oriented augite lamellae, is not uncommon. Serpentine patches which pseudomorph olivine have a characteristic rounded shape and mesh-work structure; less regular patches are assumed to represent lime-poor pyroxene. Opaque minerals, principally ilmenite–sphene intergrowths, are again scarce. Figure 18B is of a typical thin section.

Volumetric analyses of ophitic gabbro specimens are given in Table IV and chemical analyses and norms in Table IX.

### *Even-grained Gabbro*

Gabbro with an even-grained texture is common. In differentiated sills it occupies the central and upper parts of the sill, and in thin, undifferentiated sills it seems to be the major rock type. In thin sections the transitional nature of this subdivision can be seen. The rocks range from finely ophitic with rare pseudomorphs after olivine to microscopically subpegmatitic marked by slightly elongated augite and minor interstitial iron-rich chlorite and quartz. The characteristic type is subophitic with generally equidimensional pyroxene and slightly elongated plagioclase crystals. The grain size is generally 1 to 3 mm. Pyroxene is predominantly augite, but relics of both orthopyroxene and pigeonite are found in places. Commonly the lime-poor pyroxenes are represented by pseudomorphs of serpentine, talc, or amphibole and many contain oriented lamellae of augite. Intergrowths of ilmenite and sphene are slightly more abundant than in the more basic differentiates. Figure 18C is a microscopic view of a specimen of even-grained gabbro, and volumetric data are given in Table IV.

### *Subpegmatitic Gabbro and Gabbro Pegmatite*

The upper 20 to 50 per cent of differentiated sills contain abundant pegmatitic gabbro in irregular isolated bodies a few inches to a few feet across, or in large interconnecting masses that penetrate through, and occupy much of, the upper parts of sills. Pegmatite grades into or crosscuts interassociated, even-grained gabbro. Crosscutting boundaries are abrupt but not knife-edged; individual crystals are continuous across the boundary but change character in

109927

PLATE XI. Thin section showing the contact between pegmatitic and even-grained gabbro. X 4 magnification

passing from one rock to the other (Pl. XI). Commonly the transition zone is marked by a 'fence' of closely spaced, elongated, mafic crystals oriented at right angles to the boundary. Figure 17 illustrates the patchy character of pegmatite distribution.

The pegmatites are divided into gabbro pegmatite or subpegmatitic gabbro depending on the presence or absence respectively of micropegmatite or potash feldspar. Otherwise they are similar to gabbro pegmatites found elsewhere in differentiated basic rocks (Walker, 1953; Tomkeieff, 1929, pp. 102–109; Broderick, 1935, pp. 510–511; Cornwall, 1951, p. 163). Of the two types subpegmatitic gabbro is by far the more abundant.

Both types of pegmatite are characterized by greatly elongated, locally bent and twisted, tabular augites. Commonly these are twinned along the medial plane (100) of the elongated crystals so the (001) striations form herringbone patterns. They are as much as 6 cm long, 2 cm wide, and 2 mm thick, but are characteristically 8 to 10 mm long. In many places tabular plagioclase is interleaved with augite crystals. Not uncommonly amphibole partly or completely replaces augite. Knots of quartz and 'ores' up to several millimetres in diameter can commonly be seen in hand specimens. The rocks have generally a high iron content and are dark in contrast with the light greenish grey colour of more mafic gabbros. Leucocratic pegmatites marked by pink feldspars, an abundance of quartz, and minor potash feldspar are found in the upper part of a sill immediately east of Lac Pelegrin. These are the most acidic of the pegmatitic gabbros found in the area and partly bridge the gap between gabbro pegmatites and granitic veins. Plate XII illustrates pegmatitic gabbros in outcrop.

*W.R.A.B. 4-10-59*

PLATE XII. Pegmatitic gabbro in the upper part of a sill on the west side of Lac Ribero.

In thin sections the pegmatites are generally seen to be dominated by a framework of elongated augite or amphibole and plagioclase crystals (Fig. 19). The interstices are filled with quartz, 'ores', and generally one or more of epidote, a high-birefringent, iron-rich chlorite, apatite, and zircon. Gabbro pegmatites contain in addition abundant micropegmatite or minor potash feldspar. The latter has not been seen as separate crystals, but only as irregular patches in plagioclase. Potash feldspar constitutes about 10 per cent of the total feldspar in pegmatite of the Lac Pelegrin sill. Augite is the only pyroxene found in the pegmatites, but enclosed serpentine or chlorite patches in some crystals may represent altered pigeonite. Walker (1953, p. 46) found that pigeonite is the normal lime-poor pyroxene of gabbro pegmatites and commonly forms altered cores in augite. Volumetric analyses of subpegmatitic gabbro and gabbro pegmatite are given in Table IV and chemical analyses and norms in Table IX.

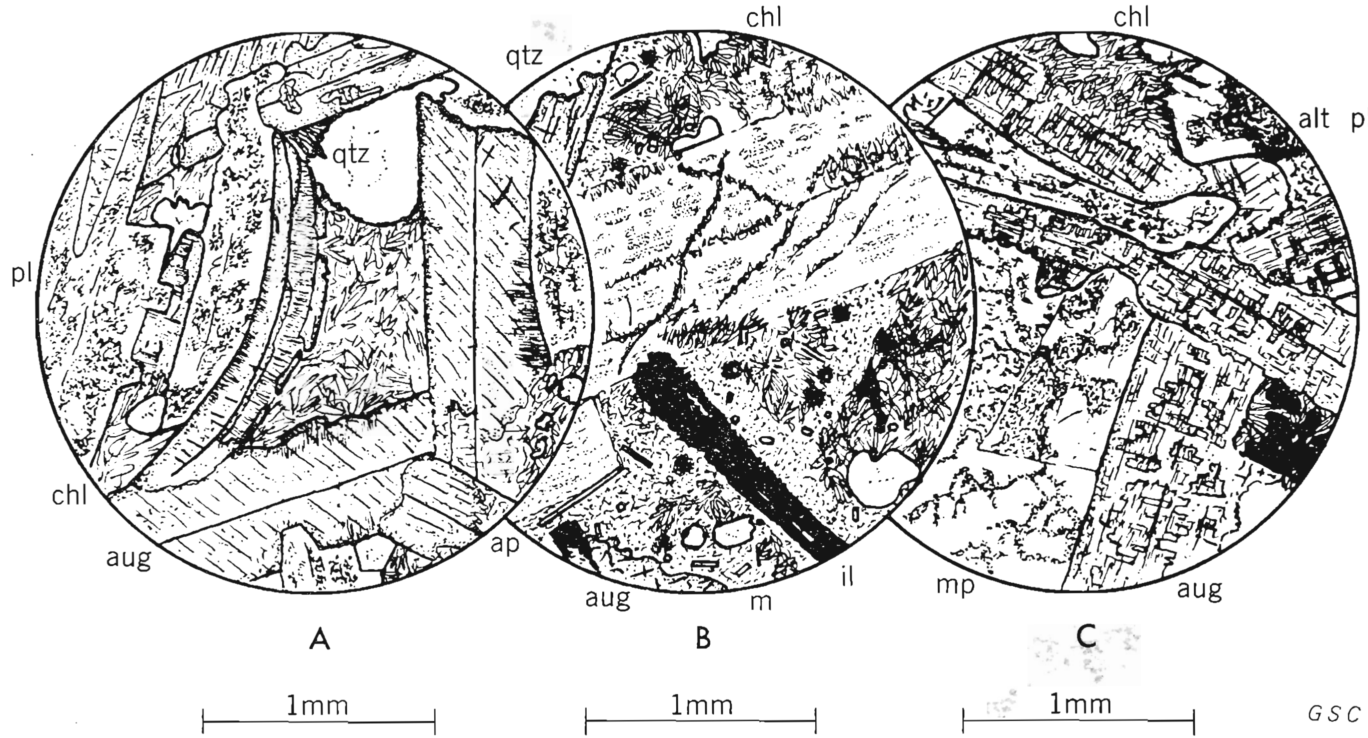



FIGURE 19. Subpegmatitic gabbro, albite gabbro pegmatite, and gabbro pegmatite.

A. Subpegmatitic gabbro showing the characteristic elongated, twinned, and, in places, bent augite crystals. Interstices contain high birefringent chlorite and quartz. Specimen B286.

B. Albite gabbro pegmatite. Shows a slightly clouded albite crystal veined and partly replaced by chlorite. The surrounding interstitial space is filled with high birefringent chlorite, unidentified mesostasis, apatite, quartz, and iron oxide. Platy ilmenite in the lower part of the field is typical. Specimen R9.

C. Gabbro pegmatite. Elongated and twinned augite crystals (partly altered) are the dominant feature. Micropegmatite, chlorite, and iron oxide minerals occupy the interstices. Specimen R554.

pl—plagioclase; qtz—quartz; chl—chlorite; aug—augite; mp—micropegmatite; il—ilmenite; m—mesostasis; ap—apatite; alt—altered

### *Granitic Veins*

Granitic veins ranging from 2 inches to about 2½ feet in thickness are found in the upper parts of sills near Lac Cramolet. The most notable occurrences are at the north edge of the large central peninsula of the lake and a mile to 2 miles north-northwest of the northern tip of the lake. Samples for chemical analyses (Table IX) were taken from veins at both localities.

*W.R.A.B. 2-21-59*

PLATE XIII. Granitic vein in a gabbro sill, central peninsula, Lac Cramolet. Note the matching walls.

Veins on the central peninsula of Lac Cramolet are typically 2 to 14 inches wide, 10 to about 100 feet long, and bounded by sharp, matching walls (Pl. XIII). They are composed of fine- to medium-grained, light cream rock containing randomly oriented, pale green slivers of actinolite and scattered rusty weathering clots of finely disseminated sulphide and oxide minerals. At least some of the sulphide mineral is chalcopyrite. The vein boundaries are not chilled and although sharp they are not a discontinuity. Crystals of each rock-type appear to be mutually intergrown across the contact. Vein matter is composed predominantly of submicrographically intergrown quartz and albite with scattered, shreddy actinolite, epidote, and 'ores' (Table IV, BL9-206). The texture ranges from an irregular but intimate intergrowth of quartz and feldspar to that which approaches a graphic intergrowth. Potash feldspar was not observed. The albite is slightly clouded with finely scattered epidote and an unidentified, dusty mineral

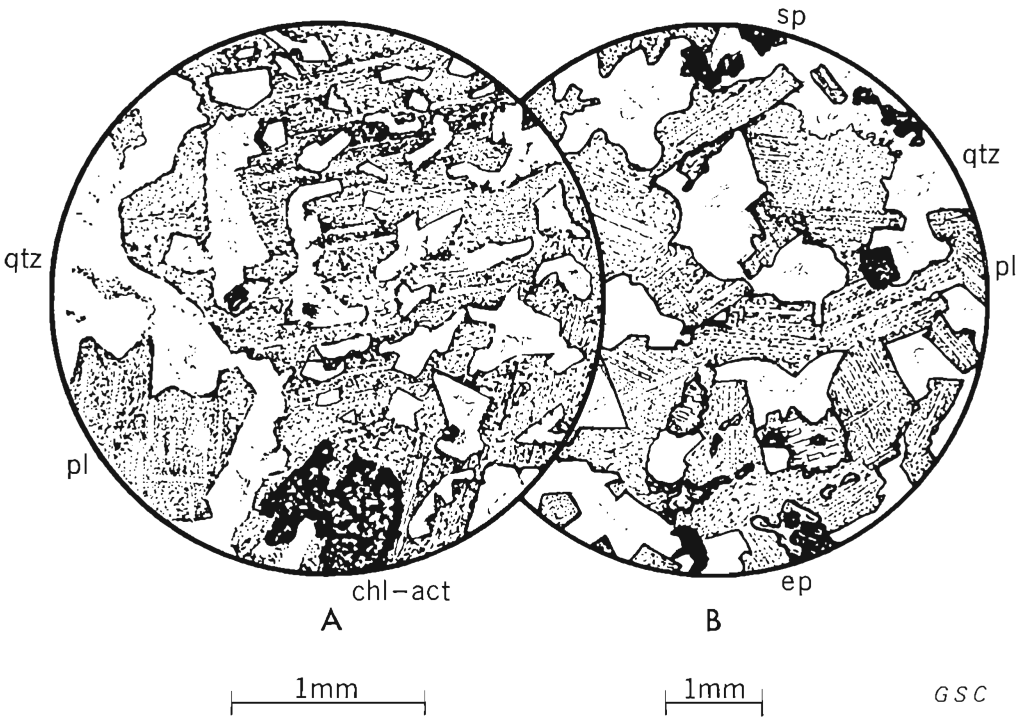



FIGURE 20. Granitic vein rocks.

A. Shows submicrographic intergrowth of quartz and slightly clouded albite. A patch of chlorite and actinolite appears at the bottom of the field and minor actinolite is embedded in some of the quartz. Specimen BL9-206.

B. The field is dominated by stout crystals of slightly clouded albite with quartz flling the interstices. Epidote, chlorite, sphene, and apatite are accessory minerals. Specimen BL9-227.

qtz—quartz; pl—plagioclase; chl—chlorite; act—actinolite; ep—epidote; sp—sphene

Figure 20A is typical of the rock in thin section.

The host rock is dark, even-grained gabbro with some subpegmatitic characteristics such as intermittent elongated amphiboles and interstitial quartz, epidote, and 'ores'. In addition the rock contains pods, 5–15 mm in diameter, composed of a core of carbonate surrounded by radiating crystals of epidote and quartz.

Northwest of Lac Cramolet a number of granitic veins were found for about 1½ miles along a ridge-summit formed by a west-dipping gabbro sill. These range from about 6 inches to 2½ feet in thickness, but unlike the veins described above the contacts with surrounding gabbro are generally gradational. Vein matter ranges from white or light cream coloured rocks with minor mafic patches to grey, intermediate rocks with a substantial mafic content. One of the lighter rocks was selected for chemical analyses (Table IX, BL9-227). The texture of the rock in thin section is dominated by randomly oriented, thick, euhedral laths of albite. Quartz tends to fill the interstices, but locally appears to corrode and replace the albite. Chlorite, tremolite, sphene, and epidote are minor constituents, but do not seem to be influenced in their growth by the major minerals. Apatite and carbonate are the principal accessory minerals. No micrographic texture is apparent and potash feldspar was not observed. Figure 20B is a typical microscopic view.

The host rock of the analyzed granitic vein is a dark, even-grained gabbro. It is little altered and contains both augite and pigeonite. The plagioclase is $An_{75}$. Elsewhere in the same sill granitic veins cut typical pegmatitic gabbro. Volumetric analyses of the two granitic vein samples are given in Table IV and chemical analyses in Table IX.

### *Albite Gabbro Pegmatite*

This variety is found only in the Ahr Lake area and all known occurrences can be related to the same sill. It forms the upper part of a folded sill immediately west and north of Hurst Lake, and grades down into altered, even-grained gabbro similar to that in other sills of the region. Albite gabbro pegmatite is a late differentiate of the gabbro and is considered equivalent to gabbro pegmatite, but of special derivation (Baragar, 1960, pp. 1629–1630). The rock is black or dark brown and contains conspicuously elongated (up to 2 cm), randomly oriented, white, light brown, or colourless plagioclase crystals. In a few places hexagonal plates of ilmenite up to 0.5 cm in diameter are visible.

Thin sections are seen to be dominated by stout albite laths and long, thin plates of ilmenite, set in a heterogeneous mixture of pyroxene, altered iron ores, chlorite, and other fine-grained minerals (Fig. 19B). Pyroxene is less conspicuous than in subpegmatitic gabbro and gabbro pegmatite and is extensively chloritized. Albite is generally slightly clouded and may be partly veined by chlorite. Locally it shows minor alteration to pumpellyite and prehnite. Pumpellyite also occurs in interstices and is locally an important minor constituent of the rock. Chlorite with subordinate quartz is a common interstitial filling. It is partly the high-birefringent variety found in some of the other pegmatitic phases and may show extreme pleochroism from brilliant orange-brown to green. Apatite, commonly with pleochroic haloes, is ubiquitous and is mainly embedded in chlorite and quartz. In addition to platy ilmenite, iron ores may occur as altered masses of sphene with triangularly arranged lamellae of ilmenite. A mesostasis of fine-grained chlorite, unidentified colourless material of low birefringence, and minute amphibole needles, all heavily charged with dusty, fibrous particles, is present in many specimens (Fig. 19B). Commonly the mesostasis contains irregular quartz crystals and abundant rods of apatite; it is interpreted as devitrified glass. Quartz is a minor constituent in most specimens of albite gabbro pegmatite, but in one locality micropegmatite is fairly abundant. Volumetric analyses of albite gabbro pegmatites are given below and chemical analyses and norms in Table IX.

| | R151 (per cent) | R94 (per cent) |
|---|---|---|
| plagioclase | 41 | 34 |
| augite | 25 | 10 |
| iron ores | 11 | 11 |
| chlorite (iron-rich) | 18 | 20 |
| quartz | 2 | 4 |
| apatite | n.d. | 2 |
| pumpellyite | 4 | None |
| mesostasis | None | 19 |

### *Feldspathic Gabbro*

Subdivisions described above are sufficient to classify most of the normal gabbro variants. One type that does not seem to belong to this sequence is a leucocratic, feldspar-rich gabbro found in two successive sills three-quarters of a mile and 1½ miles east of Lac de l'Axe.

In the western sill, within a stratigraphic distance of about 150 feet, ophitic gabbro grades upward through a zone of well-developed gabbro pegmatites to chalky white, feldspathic gabbro. The latter occupies a zone a few tens of feet wide and is overlain by dark, even-grained gabbro with pegmatitic phases. Outcrop is intermittent so the relationship of feldspathic gabbro to the remaining sill rock is obscure. The feldspathic rock contains 75 to 85 per cent saussuritized plagioclase; remaining constituents are mainly altered augite, and minor sphene.

In the eastern sill, lenses, streaks, and bulbous masses of white feldspathic material, generally parallel with the sill walls, are found within ophitic gabbro about 200 feet above the base. They range mainly from a few inches to several feet in length and are commonly from an inch to 8 inches thick. The feldspathic rock is entirely saussuritized plagioclase except for very minor sphene, carbonate, and chlorite. The specimen examined has a grain size between one-half and one mm.

### *Chilled Phases*

The chilled phases typically exhibit a very fine grained, indistinct groundmass containing scattered clusters of microphenocrysts. Plagioclase, augite, and serpentine pseudomorphs after olivine are the microphenocrysts generally present, but not uncommonly all three are pseudomorphs in serpentine or chlorite. One volumetric analysis of a chilled contact specimen is:

| | |
|---|---|
| groundmass | 83 per cent |
| plagioclase | 12 per cent |
| augite | 1 per cent |
| serpentine (after olivine) | 4 per cent |

TABLE IV

*Volumetric Analyses of Normal Gabbros*

| | Porphyritic Gabbro | | | | Ophitic Gabbro | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | B343 | B184 | B279 | B154 | B345 | R90C | B295 | R99 | R126 | B183 |
| Plagioclase | 51 | 53 | 62 | 49 | 56 | 46 | 49 | 63 | 50 | 60 |
| Olivine (+ pseudomorphs) | 4 | 13 | 15 | 18 | 4 | 3 | 13 | 13 | 2 | — |
| Orthopyroxene (+ pseudomorphs) | 34 | 11 | 11 | 26 | — | — | — | — | — | — |
| Augite | | 18 | 9 | | 22 | 39 | 29 | 19 | 24 | 27 |
| Pigeonite (+ pseudomorphs) | | | | | | 3 | | | | |
| Serpentine (uncertain origin) | 5 | — | — | — | 15 | 5 | 8 | 2 | 19 | 6 |
| Hornblende, biotite, and chlorite | 5 | 4 | 2 | 5 | 2 | 2 | — | — | 3 | 4 |
| 'Ores' | 1 | 1 | 1 | 2 | 1 | 2 | 1 | 3 | 2 | 3 |

| | Even-grained Gabbro | | | | Subpegmatitic Gabbro | | | | Gabbro Pegmatite | |
|---|---|---|---|---|---|---|---|---|---|---|
| | R34 | R64 | R100 | R153 | B286 | R23 | R129 | B152[1] | R554 | B155 |
| Plagioclase | 51 | 48 | 63 | 58 | 46 | 41 | 41 | 53 | 35 | 39 |
| Olivine pseudomorphs | — | — | 2 | — | — | — | — | — | — | — |
| Augite | 33 | 33 | 27 | 26 | 28 | 39 | 21 | 26 | 18 | 27 |
| Serpentine (after pigeonite?) | 13 | — | — | 7 | 7 | — | 11 | — | — | — |
| Serpentine (uncertain origin) | — | 17 | 5 | — | — | — | — | — | — | — |
| Hornblende, biotite, and chlorite | — | — | — | 6 | — | — | — | — | — | — |
| 'Ores' | 3 | 2 | 3 | 2 | 4 | 5 | 5 | 5 | 4 | 8 |
| Chlorite | — | — | — | — | 8 | 9 | 10 | 15 | 15 | 13 |
| Quartz | — | — | — | 1 | 7 | 6 | 12 | 1 | 28 | 13 |
| Micropegmatite | | | | | | | | | (32) | (9) |

| | Granitic Veins in Gabbro | |
|---|---|---|
| | BL9-206 | BL9-227 |
| Plagioclase | 52 | 52 |
| Quartz | 39 | 27 |
| Tremolite–actinolite + chlorite | 5 | 8 |
| 'Ores' | 1 | 3 |
| Epidote | 2 | 3 |
| Accessory minerals (mainly carbonate and apatite) | 1 | 5 |

| | Averages | | | | | |
|---|---|---|---|---|---|---|
| | Porphyritic Gabbro | Ophitic Gabbro | Even-grained Gabbro | Subpeg-matitic Gabbro | Gabbro Pegmatite | Granitic Veins |
| Plagioclase | 54 | 55 | 55 | 44 | 38 | 52 |
| Pyroxene[2] | 33 | 37 | 41 | 34 | 22 | 6 |
| Olivine[2] | 12 | 6 | 0.5 | — | — | — |
| 'Ores' | 1 | 2 | 3 | 5 | 6 | 2 |
| Quartz | — | — | 0.5 | 8 | 20 | 34 |
| Iron chlorite | — | — | — | 9 | 14 | — |
| Accessory minerals[3] (mainly carbonate, epidote, and apatite) | — | — | — | Tr. | Tr. | 6 |

[1]Transitional between even-grained and subpegmatitic gabbros.
[2]Pyroxene includes serpentine (other than that pseudomorphous after olivine) and other mafic accessories. Olivine comprises olivine and olivine pseudomorphs.
[3]Epidote in discrete crystals, not obviously derived from plagioclase.

### *Alteration*

The normal gabbros have all been altered to some extent, partly by regional metamorphism and partly by autometamorphism common to most differentiated sills or flows (Emmons, 1927, pp. 75–76; Cornwall, 1951, pp. 162–163; Hotz, 1953, p. 683). The effects due to each cannot be clearly separated; indeed, for some minerals they may overlap. Broadly, the effects of autometamorphism are evident in the general increase of alteration from the early to late crystallizing fractions of the sills. This presumably reflects the action of ascending late magmatic fluids on earlier formed crystals. Thus, fresh olivine, although rare, is found only in porphyritic gabbros; lime-poor pyroxenes persist into the even-grained gabbros, but are more commonly fresh in the basic members; and plagioclase shows steady deterioration from porphyritic to pegmatitic gabbros.

Alteration effects on the various minerals are as follows. (a) Plagioclase is typically saussuritized to a fine-grained, greyish brown mass. In many places pumpellyite and possibly prehnite are present, but they are considered to be products of regional metamorphism and will be discussed later. (b) Olivine is generally replaced by a meshwork of serpentine or chlorite and less commonly talc, bowlingite, tremolite, and even quartz. Magnetite is commonly released by the replacement. (c) Orthopyroxene and pigeonite alter to serpentine, talc, chlorite, and amphibole. (d) Augite resists alteration to a marked degree but in places shows a brownish discoloration that blurs the optical properties and reduces the birefringence. More rarely it is partly replaced by tremolite–actinolite or chlorite. Striations parallel with (001) are fairly common and appear to be due to lamellae of chlorite or serpentine along this plane. However, these may result from alteration of pigeonite exsolution lamellae on the (001) planes of augite rather than from the alteration of augite. Pigeonite lamellae can be expected to have exsolved from augite, but have not been found in Wakuach Lake map-area. Since pigeonite is normally readily altered, exsolution lamellae are not likely to survive even a modest degree of alteration. (e) Most of the iron 'ores' are replaced by a murky, brownish, translucent mass that is mainly sphene. However, ilmenite, whether as lamellae in the sphene masses or as separate crystals, is little altered.

## Normal Gabbro Series

Repetition in sills throughout the area of the succession from porphyritic or ophitic gabbro to gabbro pegmatite establishes this as the normal differentiation sequence. Granitic veins are an expectable end product and are included with the others in what is here termed the normal gabbro series. Albite gabbro pegmatite is considered an alternative or branch product of the same parent magma formed under somewhat different conditions. On Figure 21 the variation in average mineral content throughout the series is graphed using figures supplied in Table IV. Albite gabbro pegmatites are excluded as they are atypical of the differentiation sequence found in this area.

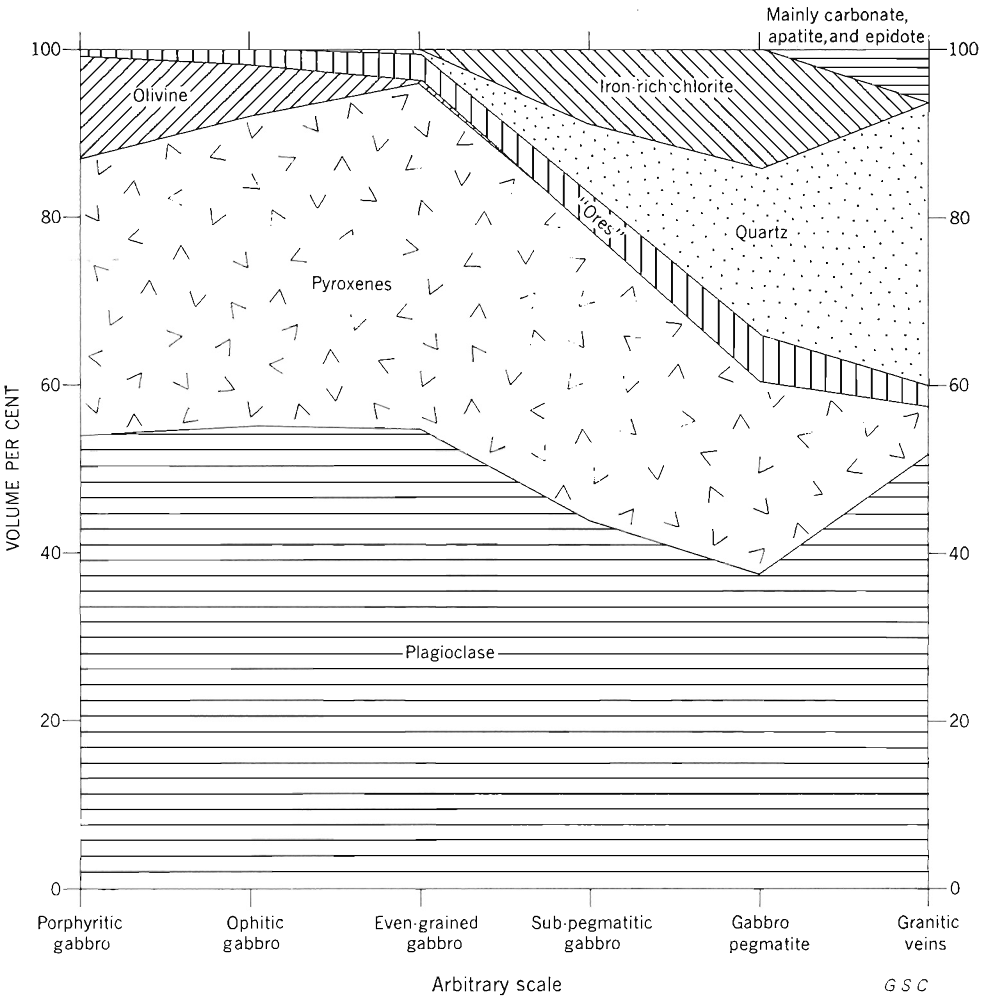



FIGURE 21. Mineral variation in normal gabbro series. Cumulative diagram based on average volumetric analyses of each type of gabbro (Table IV).

## Mineralogy

Much of the data that follows is summarized from the previous study of Ahr Lake basaltic rocks (Baragar, 1960, pp. 1609–1616). Details may be obtained from the original publication.

### *Olivine*

Fresh olivine is rare and found only in the most basic rocks. The olivine from three porphyritic gabbro specimens determined optically ranges from $Fa_{32}$ to $Fa_{35}$. In these the olivine is somewhat richer in iron than co-existing orthopyroxene.

### *Orthopyroxene and Pigeonite*

Both orthopyroxene and pigeonite are highly susceptible to alteration. Accordingly they are more commonly represented in pseudomorphous form than in primary form. Fresh lime-poor pyroxene is limited to porphyritic, ophitic, and rarely even-grained gabbros. Orthopyroxenes determined optically are bronzites ranging from $Fe_{24}$ to $Fe_{29}$. Pigeonites range from $Fe_{34}$ to $Fe_{36}$, if one assumes a maximum calcium content (9 per cent).

Both orthopyroxene and pigeonite generally exhibit augite exsolution lamellae and are commonly intimately intergrown with augite. Thus a large augite crystal may contain several irregular, dispersed patches of orthopyroxene or pigeonite with a common orientation. In addition each patch may contain augite exsolution lamellae, that show optical continuity with the augite host. The b and c axes of both minerals in the intergrowth coincide. It appears that augite and lime-poor pyroxene have exsolved and partly separated from a common solid solution and that each in turn has further exsolved remaining components of the other as the magma cooled.

The original identity of a lime-poor pyroxene pseudomorph can commonly be determined by the orientation of augite exsolution lamellae that still survive. The methods follow from the conclusions of Poldervaart and Hess (1951, p. 481). Two sets of lamellae indicate that the original pyroxene was orthopyroxene inverted from pigeonite. One set of lamellae may mean either original orthopyroxene with augite exsolved on (100) or original pigeonite with augite exsolved on (001). Commonly the (001) lamellae may be recognized by reference to the orientation of intergrown augite or by their herringbone pattern, which marks the position of a (100) twin in the original pigeonite.

Generally orthopyroxene seems to retain the b and c axes of pigeonite upon inversion. This is most readily apparent in the coincidence of (100) with the composition plane of a former pigeonite twin, now marked by (001) augite lamellae, or by reference to the orientation of intergrown augite. In a few widely separated places in the area it is evident that orthopyroxene did not retain the b and c axes of pigeonite upon inversion, for (100) of orthopyroxene bears no predictable relationship with (001) lamellae of original pigeonite. Brown (1957, pp. 542–543) first observed such a relationship in the Skaergaard pyroxenes.

Lime-poor pyroxene seems to respond to differentiation in the following ways. Orthopyroxene predominates in the early differentiates but gives way to pigeonite in the middle stages. Pigeonite may persist into the pegmatite phase but is not preserved and contains no augite lamellae. Lime-poor pyroxene is intimately intergrown with augite in the mafic members of the series but seems to be more distinctly separate in the middle members.

### *Augite*

The compositions of augites from sills in the Ahr Lake area are summarized on Figure 22 where the compositional trend normal for tholeiites (Hess, 1941, p. 585) is also shown. The compositions are based on determinations of refractive

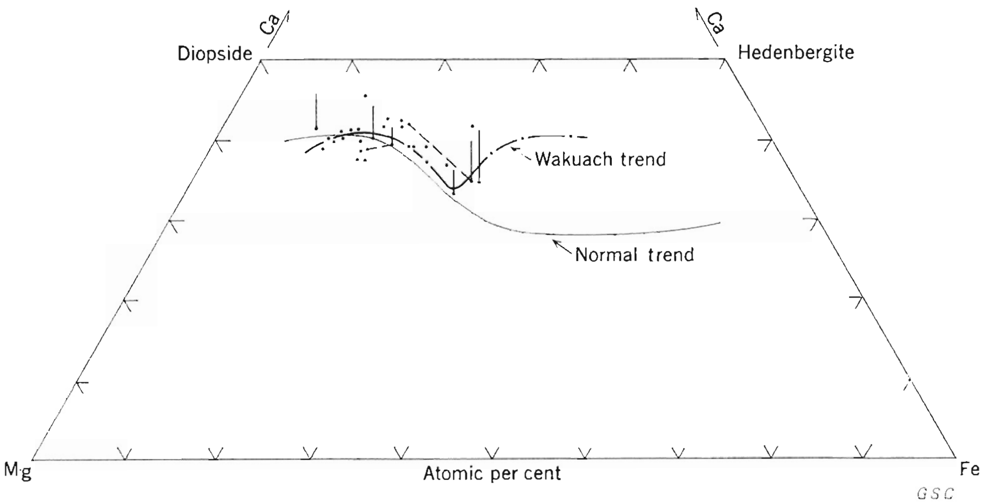



FIGURE 22. Augite compositions (33) in Wakuach Gabbros in comparison with the normal trend. All compositions are based on optical determinations.

index and 2V using Hess' correlation curves (1949, p. 634). It should be noted in comparing trends that Hess used no pegmatitic augites in establishing the normal trend, whereas these augites are largely responsible for the iron-rich end of the trend in Wakuach gabbros. Some of the augites, possibly because of alteration, show a variation of 2V within a single crystal without a detectable variation in refractive index. These are represented on the diagram by a line parallel with the index curve. The base of each such line should probably be assumed to represent the original composition.

The range of compositions shown on Figure 22 is probably nearly a maximum for this area. The pyroxene at the iron-rich end of the trend line is from a pegmatite representing the iron-enriched apex of the chemical differentiation trend. More acid rocks, such as the granitic veins and the pink potash feldspar-bearing gabbro pegmatites of Lac Pelegrin, contain amphibole but no augite. Presumably this is the approximate point in differentiation where the vapour pressure exceeds the one at which augite is stable for the temperature prevailing.

The most marked response of augite to the changing magmatic environment during differentiation is manifested by its form. In the porphyritic and ophitic stages augite is present as roughly equidimensional, ophitic crystals. In even-grained gabbros, it may be equidimensional entirely or a small percentage of the crystals may be slightly elongated. In subpegmatitic gabbro and gabbro pegmatite, the augite consists of characteristically elongated, tabular plates almost invariably twinned along the (100) plane.

### *Plagioclase*

The plagioclase composition ranges generally from $An_{80}$ to $An_{60}$ in porphyritic and ophitic gabbros, from $An_{70}$ to $An_{50}$ in even-grained gabbros, from $An_{60}$ to about $An_5$ in pegmatitic gabbros. Generally the more sodic plagioclases are in pegmatites of evident high alteration. Most subpegmatitic gabbros and gabbro pegmatites showing modest alteration effects contain plagioclase of $An_{55}$ to $An_{35}$. The plagioclase is mainly zoned so that compositions are approximate. In both albite gabbro pegmatite and the granitic veins the plagioclase is albite and is slightly clouded.

### *Opaque Minerals*

Sulphide minerals are generally minor, but appear to be slightly more abundant in both early and late differentiates. In porphyritic and ophitic gabbros pyrrhotite predominates and chalcopyrite and pyrite are subordinate. In the granitic veins chalcopyrite and pyrite appear to be the main sulphide minerals when any are present.

Iron ores are most commonly ophitic masses of turbid, granular sphene enclosing triangularly arranged lamellae of ilmenite. The sphene is similar to what is normally called leucoxene, but in X-ray powder photographs only sphene and ilmenite could be identified. Elongated, skeletal ilmenite crystals (Fig. 19B) are mainly restricted to pegmatitic gabbros, particularly albite–gabbro pegmatites.

Sphene masses with oriented lamellae of ilmenite are interpreted as altered titaniferous magnetite containing ilmenite lamellae on the (111) planes. Recent work by Lindsley (1962, pp. 100–106) provided convincing evidence that ilmenite lamellae in 'magnetite' resulted from oxidation of a magnetite–ulvöspinel solid solution rather than from exsolution. Original magnetite–ulvöspinel solid solutions may, depending upon their composition and the oxygen fugacity, either react with the environment to form iron–titanium spinel with ilmenite–hematite solid solution lamellae or, in the absence of excess ilmenite, remain stable. The composition of the iron–titanium spinel remaining after reaction is a function of the reaction temperature and oxygen fugacity, but for basalt temperatures it should probably be in the vicinity of 50 per cent of each of magnetite and ulvöspinel (cf. Lindsley, 1962, Fig. 35). Upon cooling, ulvöspinel- and magnetite-rich components mutually exsolve and form a fine-grained intergrowth (Vincent and Phillips, 1954; Lindsley, 1962, p. 105). Replacement of the iron–titanium spinels by sphene while ilmenite remains unaltered may be partly due to the susceptibility to replacement of such an intergrowth.

The oxide minerals form a guide to the state of oxidation in the magma (Buddington, *et al.,* 1955), which in turn influences the differentiation trend (Kennedy, 1955, p. 501; Wilkinson, 1957, p. 452). For much of the course of crystallization in the Wakuach Gabbros, magnetite–ulvöspinel solid solutions evidently separated from the magma. At the pegmatitic stage ilmenite appears to be more abundant both as lamellae in sphene masses and as separate crystals. The oxidation potential in the magma might be interpreted from this as generally fairly low, but increasing with differentiation to at least the pegmatitic stages.

The identification of primary oxide minerals in the granitic veins is uncertain. Some goethite appears with the sulphide minerals, but it may be an alteration product.

### *Chlorite*

A variety of chlorite peculiar to many of the pegmatitic gabbros appears to be a late-crystallizing, primary mineral. It has brown to green pleochroism and a fairly high birefringence, both of which increase in intensity with an increasing iron to magnesium ratio in the rock. An extreme variety from an albite gabbro pegmatite specimen (R94) has the following properties:

| | |
|---|---|
| Pleochroism: | x = deep reddish brown, |
| | y = z = grass green |
| Refractive index: | $\gamma = 1.658 \pm 0.002$ |
| | $\alpha = 1.641 \pm 0.002$ |
| | $\gamma - \alpha = 0.017$ |
| 2V: | Essentially zero (−) |

The high-birefringent chlorite inhabits the interstices of pegmatitic gabbros together with such typically late-crystallizing minerals as quartz and apatite. Normal, low-birefringent chlorites appear to be mainly alteration products of mafic minerals but some, particularly in the granitic veins where they may be in discrete patches, are possibly primary.

### *Epidote*

Coarse, discrete epidote crystals and clusters of crystals that occur in the interstices of some of the pegmatitic gabbros and granitic veins are considered primary, late-crystallizing minerals. These are distinct from the ubiquitous, brown, fine-grained epidote–zoisite masses that blight much of the plagioclase throughout the gabbros and are obviously derived from it.

### *Potash Feldspar, Apatite, and Zircon*

These three minerals occur only in the late differentiates, but are by no means universal in these rocks. Potash feldspar was recognized, and confirmed by staining, in only three of several dozen thin sections of pegmatitic gabbros, but might be present in two or three more. It is absent from the granitic veins. Characteristically, it appears as irregular patches in albite crystals and does not form crystals of its own. Zircon was recognized in a number of gabbro pegmatites and may be present in others even though it was not identified. Spectrographic analyses (Table X) show a high content of zirconium in some pegmatitic specimens lacking recognizable zircon. Zircon crystals are generally accompanied by pleochroic haloes. Euhedral rods of apatite are very common accessory minerals in pegmatitic gabbros and granitic veins. Generally they are finely dispersed through the interstices of the rock, but in places reach lengths up to 2 mm and penetrate all surrounding minerals. Some of the apatite crystals are accompanied by pleochroic haloes, which may or may not extend along the complete length of the rod.

## Chemistry

Two analyses of granitic veins (BL9-206, BL9-227) from the Lac Cramolet region have been added to those of the normal gabbro series previously reported (Baragar, 1960, pp. 1634–1635), and all are presented in Tables IX and X. Basalt analyses also given in these tables represent: (1) lavas of the Menihek Formation in the Ahr Lake area (composite sample B54A); and (2) a composite sample of Willbob Lake lavas from the Griffis Lake area (Fahrig, 1962, p. 4). These are included in the normal gabbro series with a view to comparing the chemistry of lavas and sills.

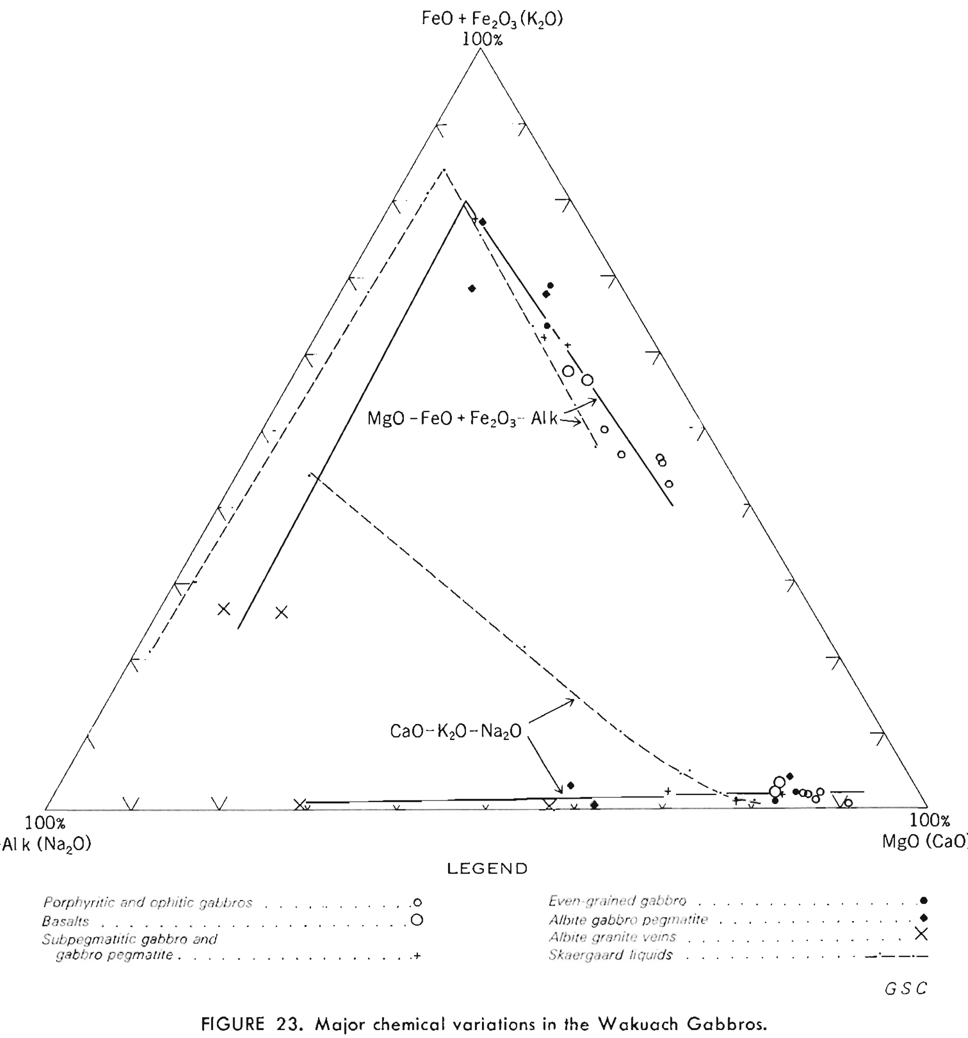



FIGURE 23. Major chemical variations in the Wakuach Gabbros.

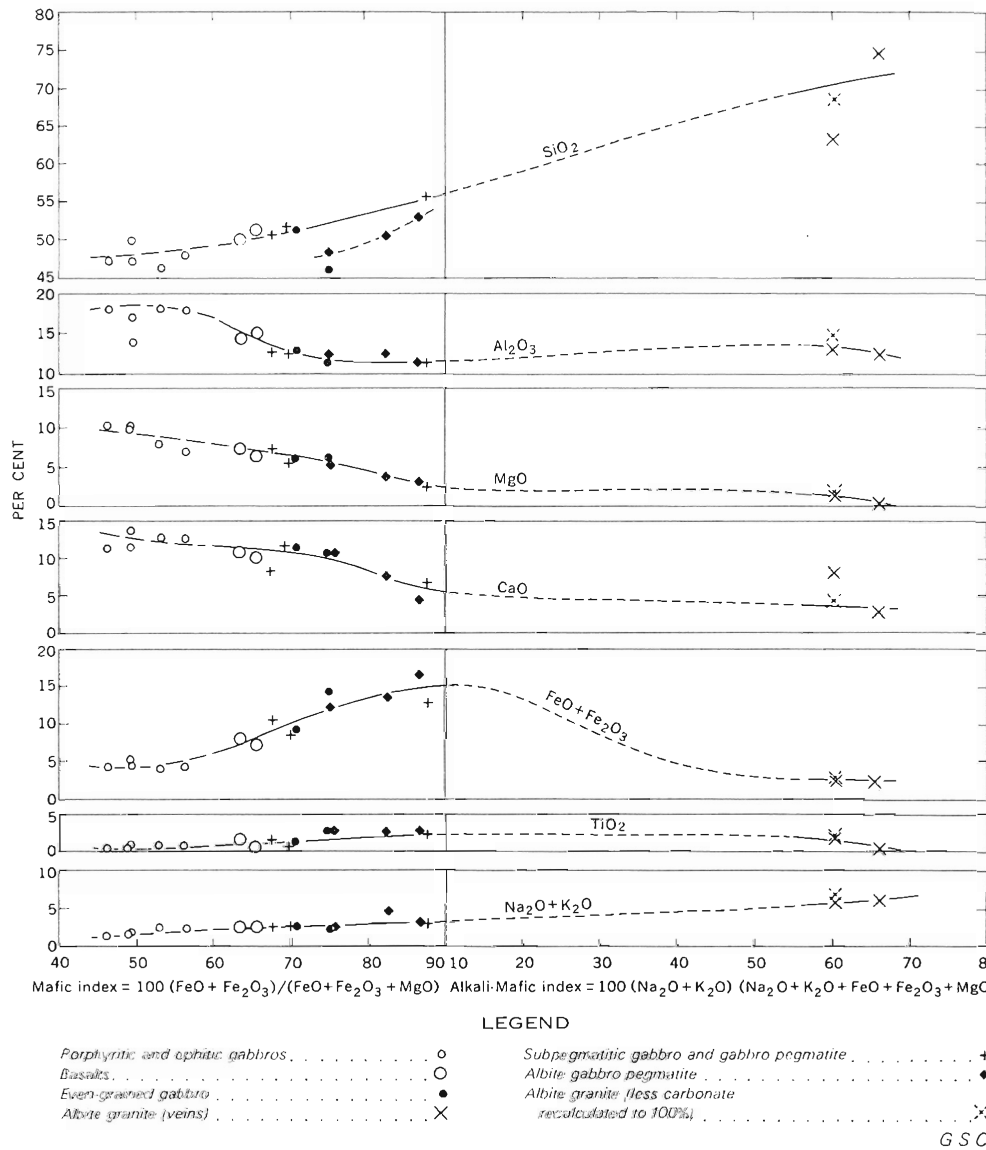



FIGURE 24. Variations of oxides with mafic and alkali-mafic indices in the normal gabbro series.

## *Major Elements*

Variations in the major elements of the normal gabbro series are shown on Figures 23 and 24. Figure 23 shows the principal characteristics of differentiation in the Wakuach Gabbros: a marked enrichment in iron for the major part of the series, with late development of sodic granite and an exceptionally low potash content throughout. The similarity of the Wakuach trend with that of the Skaergaard liquids (shown for comparison) is striking except for lack of potash enrichment in the former.

Figure 24 illustrates the variation of each major oxide during differentiation. The measures of differentiation used are the mafic and alkali-mafic indices (Simpson, 1954, p. 238) for the iron-enrichment and sodium-enrichment parts of the trend, respectively. These assume that the stage reached in a differentiating magma is indicated by the degree of enrichment of either iron or alkalis. Iron-enrichment is most marked during the major stage of crystallization in a mafic magma, but alkali-enrichment is dominant in the late-crystallizing fraction. It

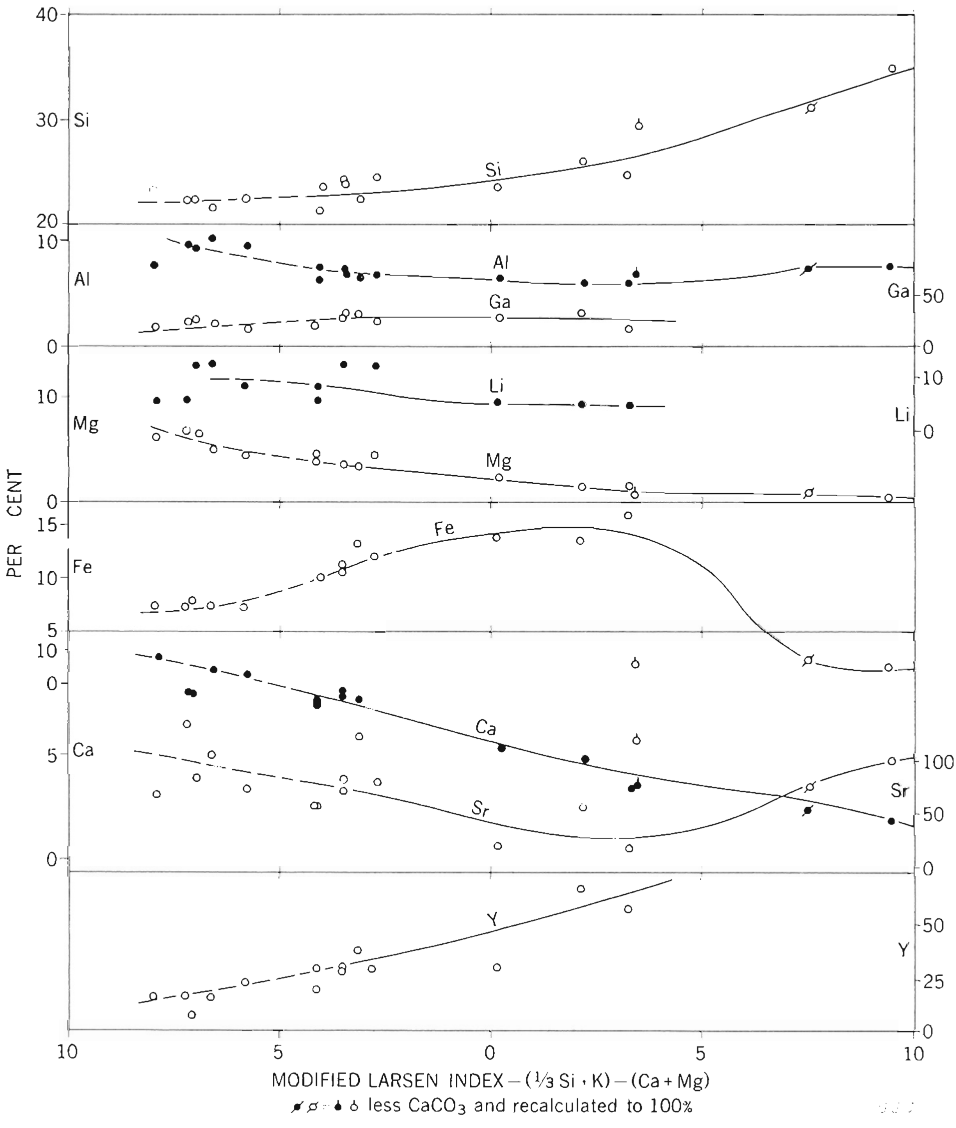
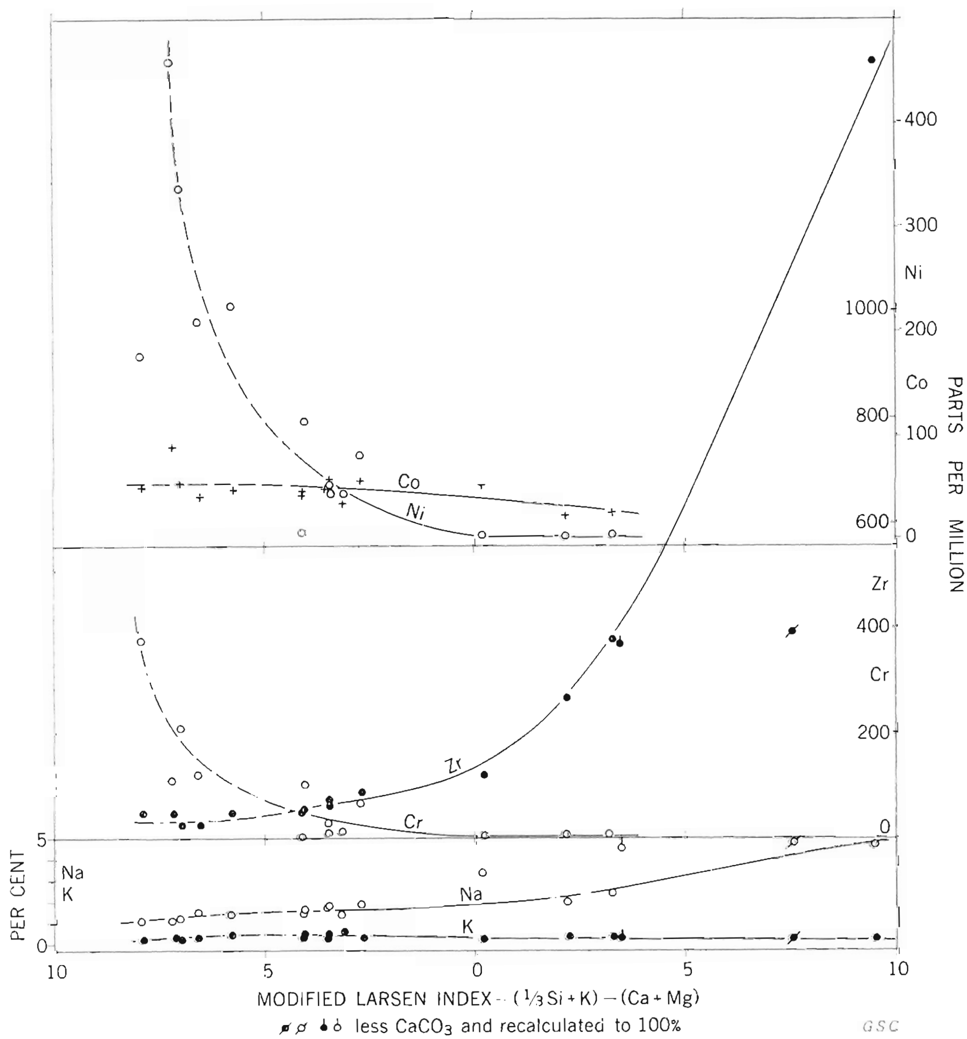



FIGURE 25. Variation of major and minor elements with

is therefore desirable to represent the course of differentiation in each phase by the index most suitable. For the Wakuach Gabbros where each phase is clearly separated this is obviously a satisfactory and reliable method. Features of note in this diagram are: the steady rise of silica content throughout the series; the marked decline of iron in the late phase; and the decline of aluminum, magnesium, and calcium oxides during the main period of crystallization.

*Minor Elements*

Variations in major and minor elements of the series are shown on Figure 25. The curves of minor elements are placed adjoining the curves of major elements with which they are normally most closely associated. The diagram is the modified




modified Larsen index in the normal gabbro series.

Larsen diagram used by Nockolds and Allen (1953; 1954; 1956) in which the abscissa is the function ($\frac{1}{3}$ Si + K) — (Ca + Mg) and the ordinate the metal in parts per million or per cent. One of the granitic samples (BL9-227) contains considerable carbonate, which is represented in the analyses by 3.18 per cent $CO_2$. It is doubtful whether this is representative of the granitic fraction as the mobility of $CO_2$ in the magma would likely favour its local concentration. Accordingly, values from both the original analysis and one from which the calcite has been eliminated are shown on the diagram. For most elements the modified values seem to fit the curves better.

Notable features in the behaviour of the minor elements during differentiation are given in the following paragraphs. Chromium and nickel are both abundant in early members of the series, but decline rapidly to the vanishing point in the middle members. This is the characteristic behaviour for both metals in the tholeiitic series elsewhere (Nockolds and Allen, 1956, pp. 47, 51–55). Nickel generally replaces magnesium or ferrous iron in olivines (Wager and Mitchell, 1951) and can be expected to decline as the abundance of olivines diminishes; chromium substitutes for ferric iron in magnetite and augite (Wager and Mitchell, 1951).

Gallium remains nearly constant or increases slightly. According to Wager and Mitchell (1951, p. 182) and Rankama and Sahama (1950, pp. 720–721), it substitutes for aluminum in silicates and for ferric iron in magnetite. In Nockolds' and Allen's (1956) diagrams gallium remains constant or declines slightly.

Cobalt is constant throughout most of the mafic part of the series but declines slightly near the end. The cobalt:nickel ratio increases. This behaviour is similar to some of the tholeiitic series studied by Nockolds and Allen (1956, p. 54), particularly the Dillsburg sill. According to Wager's and Mitchell's (1951, p. 162) data, cobalt is distributed between pyroxene, ilmenite, magnetite, and olivine in proportion to their abundance in the rock. Like nickel, it substitutes for magnesium and ferrous iron.

Zirconium increases at a continually rising rate throughout the series. According to Rankama and Sahama (1950, pp. 565–566) and Mason (1952, pp. 115–116), zirconium is present in igneous rocks principally as zircon, but Wager and Mitchell (1951, p. 192) found zirconium in pyroxene and apatite, where it is reported to substitute for calcium. In the Wakuach Gabbros, zircon was recognized in some gabbro pegmatites and granitic veins, but it is doubtful whether any could be found in the earlier members of the series. Presumably, zirconium resides in pyroxene or other calcium minerals until it is sufficiently concentrated in the late stages of differentiation to crystallize its own mineral. Its behaviour here is characteristic for tholeiitic series according to the data of Nockolds and Allen (1956).

Yttrium increases steadily throughout the mafic part of the series. This behaviour accords with data by Nockolds and Allen (1956, p. 45). It occurs in apatite, sphene, and pyroxene (Wager and Mitchell, 1951, p. 163; Rankama and Sahama, 1950, p. 524), where it replaces calcium. In the Wakuach basaltic rocks it probably occurs in augite and in later stages also in apatite.

Strontium analyses show a particularly erratic distribution. Generally they define a curve that decreases through the mafic members of the series, but rises in the granitic phase. This differs from its normal behaviour in tholeiitic series (Nockolds and Allen, 1956, pp. 61–65), where the strontium content remains nearly constant or increases slightly with rising values of the function. However, in common with its behaviour elsewhere, the ratio of strontium to calcium increases with differentiation throughout the series. Strontium normally follows calcium, but, according to Turekian and Kulp (1956, p. 265), in basaltic rocks its concentration is independent of the amount of calcium, and its variation is mainly a function of magmatic differentiation or initial regional differences. The limited spread in strontium values caused by differentiation in any one province is generally outweighed by marked differences in average values between provinces (Turekian and Kulp, 1956, pp. 267–269). Thus the average strontium content of basaltic rocks should be characteristic for a province. The average of eleven determinations for the Wakuach Gabbros is 84 ppm, which is lower than for thirty provinces listed by Turekian and Kulp (1956, p. 268).

The barium content is generally low except for two erratically high values (Table X). Since barium follows potassium, its behaviour reflects the low potassium content of this province.

### *Comparison of Lavas and Gabbros*

The analysis of Menihek basalt (B54A, Table IX) is remarkably similar to that of the composite sample of Willbob Lake basalts, and both conform closely with the trend lines and variation curves of the normal gabbro series (Fig. 23). Characteristics peculiar to the normal gabbros, such as unusually low contents of potassium and strontium, are reflected in the lava. There can be little doubt that both lavas and gabbros derive from a common source.

## Glomeroporphyritic Gabbro

### Petrographic Descriptions

Sills of glomeroporphyritic gabbro consist of marginal zones, as much as 75 feet wide, and coarse-grained interiors. Normal gabbros found within some of the sills are evidently slightly later intrusions; thus the sills are considered to be composite.

The marginal rocks are similar to even-grained gabbro. Labradorite laths and equidimensional augite are ophitic or subophitic. Rare serpentine blebs are pseudomorphs after olivine. Iron ores form altered masses in interstices or show subophitic texture with plagioclase. Augite is partly replaced by hornblende and plagioclase is saussuritized. The first feldspar clots, 2–10 feet inward from the chilled margin, are generally small phenocrysts 2–5 mm in diameter, or clusters of two or three phenocrysts. Farther from the contact the size of the phenocrysts and the number in a cluster increase until the clots have a diameter of 2–3 cm. As marginal gabbro passes into glomeroporphyritic gabbro the number of feldspathic clots increases rapidly, crowding the gabbroic groundmass into interstitial spaces, and finally eliminating it as a recognizable phase. The phenocrysts throughout are heavily saussuritized plagioclase. Volumetric data of a fine-grained specimen ($\pm$ 0.2 mm) 2 feet from the contact are:

| | |
|---|---|
| plagioclase phenocrysts | 2 per cent |
| plagioclase | 47 per cent |
| pyroxene | 30 per cent |
| hornblende | 14 per cent |
| chlorite | 1 per cent |
| iron ores | 5 per cent |
| serpentine (after olivine) | 1 per cent |

Typical glomeroporphyritic gabbro in the interior of the sills contains scattered large aggregates of altered plagioclase, 6–15 cm in diameter, set in a feldspar-rich groundmass of 0.5–3 cm grain size. It consists of loosely packed feldspar clots and interstitial pyroxene, hornblende, quartz, biotite, chlorite, and apatite. Clots are single, nearly equidimensional plagioclase crystals ($\pm$ 1 cm) or clusters of several crystals that are extensively altered to saussurite, sericite, and locally prehnite. In many places the marginal zones of crystals are unaltered; more rarely whole crystals remain unaltered. Plagioclase shows strong progressive zoning with an abrupt outer rim. Small, little altered plagioclase laths (0.5 mm) are found in interstices and may be ophitically enclosed by pyroxene or hornblende. Augite crystals up to 2 cm across fill areas between feldspar clots. Hornblende forms large single crystals or masses of minute, randomly oriented acicular crystals. At least part of the hornblende is derived from pyroxene. Altered iron ores ranging up to 3 mm in diameter enclose small interstitial plagioclase crystals ophitically. Brown biotite forms fringes around the ores. Quartz is interstitial and apparently crystallized last. Two sections show minor amounts of a light brownish mesostasis composed of vaguely defined, finely crystalline quartz charged with minute, thread-like crystals and dusty material. Locally the mesostasis surrounds and embays larger crystals of clear quartz and corrodes plagioclase. Possibly the material is devitrified glass. Interstitial, fine-grained mixtures of amphibole, quartz, epidote, and unidentified dusty material are seen in other thin sections and may be of similar origin. Apatite is generally embedded in mesostasis.

Plagioclase ranges from $An_{70}$ in the interior of crystals in the coarse-grained phase to $An_{24}$ at the outer rim. In a marginal rock it was determined as $An_{60}$.

Normative plagioclase from the chemical analysis of a feldspathic clot (Table IX, R181) is $An_{72}$ whereas that from the chemical analysis of the coarse-grained groundmass is $An_{63}$.

## Chemistry

### *Major Elements*

Chemical analyses of glomeroporphyritic gabbros given in Table IX are for specimens of a chilled contact (R161), a marginal rock 2 feet from the contact (R163), the groundmass of fully developed glomeroporphyritic gabbro (R177), and one of the larger feldspathic aggregates (R181). The chilled contact specimen contains abnormally high amounts of potash feldspar and little lime in comparison with the other analyses and it might reasonably be concluded that it has been slightly metasomatized. The norm of the feldspathic aggregate (R181) contains about 6 per cent nepheline, but if muscovite is substituted for orthoclase and zoisite for anorthite the norm has minor quartz. The chemical analysis of typical glomeroporphyritic gabbro resembles analyses of anorthositic gabbros given by Buddington (1939, p. 36), except for its lower silica.

### *Minor Elements*

Spectrographic analyses of a set of marginal specimens (R163-172) taken intermittently from the contact into the glomeroporphyritic gabbro interior are given in Table X. Feldspathic clots were excluded from all but R172. Significant features are as follows: (1) The minor element content of the four marginal rocks is similar and differs from the glomeroporphyritic phase. This supports petrographic observations that gabbro of the margin is uniform throughout except for addition of feldspathic clots. (2) Strontium values in the marginal rock are similar to those in normal gabbros and exceptionally low compared with other basaltic provinces (Turekian and Kulp, 1965, p. 268).

### *Relationship of Glomeroporphyritic Gabbro to Normal Gabbro Series*

Chemical analyses (recalculated water-free to 100 per cent) of the chilled contact and marginal zone specimens of glomeroporphyritic gabbro are compared with the basalt analysis (B54A) in Table V. The basalt is characteristic of the normal gabbro series, as will be shown later in this report. If the effects of metasomatism (high potassium, low calcium) are neglected, the chilled contact is similar to the basalt regarded as parent magma of the normal gabbros. A general relationship between normal gabbros and the border phase of leopard rock is also indicated by the low strontium content. The border phase passes into leopard rock by a marked increase in number and size of feldspathic clots. The conclusion indicated is that leopard rock is derived from a magma similar to the parent of the normal gabbro series by addition of plagioclase.

TABLE V

*Analyses of Glomeroporphyritic Gabbro, Marginal and Contact Specimens, and Normal Gabbro Series Basalt*

*(Recalculated water-free, to 100 per cent)*

| | B54A[1] | R161[2] | R163[3] |
|---|---|---|---|
| $SiO_2$ | 50.07 | 50.28 | 50.43 |
| $TiO_2$ | 1.32 | 1.65 | 1.61 |
| $Al_2O_3$ | 14.79 | 15.50 | 16.95 |
| $Fe_2O_3$ | 1.44 | 0.88 | 0.82 |
| FeO | 11.69 | 12.80 | 13.09 |
| MnO | 0.22 | 0.14 | 0.19 |
| MgO | 7.55 | 7.48 | 3.93 |
| CaO | 10.52 | 6.12 | 10.09 |
| $Na_2O$ | 2.06 | 2.17 | 2.32 |
| $K_2O$ | 0.25 | 2.74 | 0.48 |
| Rest | 0.09 | 0.24 | 0.09 |

[1]Basalt, presumed parent of normal gabbro series.
[2]Chilled contact, leopard rock sill.
[3]Border zone gabbro, leopard rock sill 2 feet from contact.

## Meta-gabbros

Meta-gabbro sills found in sedimentary and volcanic rocks on the eastern side of the Labrador Trough seem to be uniformly metamorphosed to the greenschist facies and present generally a remarkably similar appearance. Two varieties of sills with distinctive features are: (1) leucometa-gabbro–melanometa-gabbro compound sills; and (2) layered sills containing acidic schlieren or layers. The limited data obtained on the layered sills was presented in Chapter III and they will not be considered further.

### General Type

#### *Petrographic Description*

The meta-gabbros are normally composed of albite, epidote (may include clinozoisite and zoisite), chlorite, actinolite, and sphene. Textures are metamorphic, but locally shapes of former plagioclase laths are apparent in dusty aggregates of epidote and albite. Actinolite forms shreddy-edged, equidimensional crystals up to 4 mm in diameter, or small acicular crystals; some are bent or broken. Chlorite is associated with actinolite or is in separate patches. Sphene

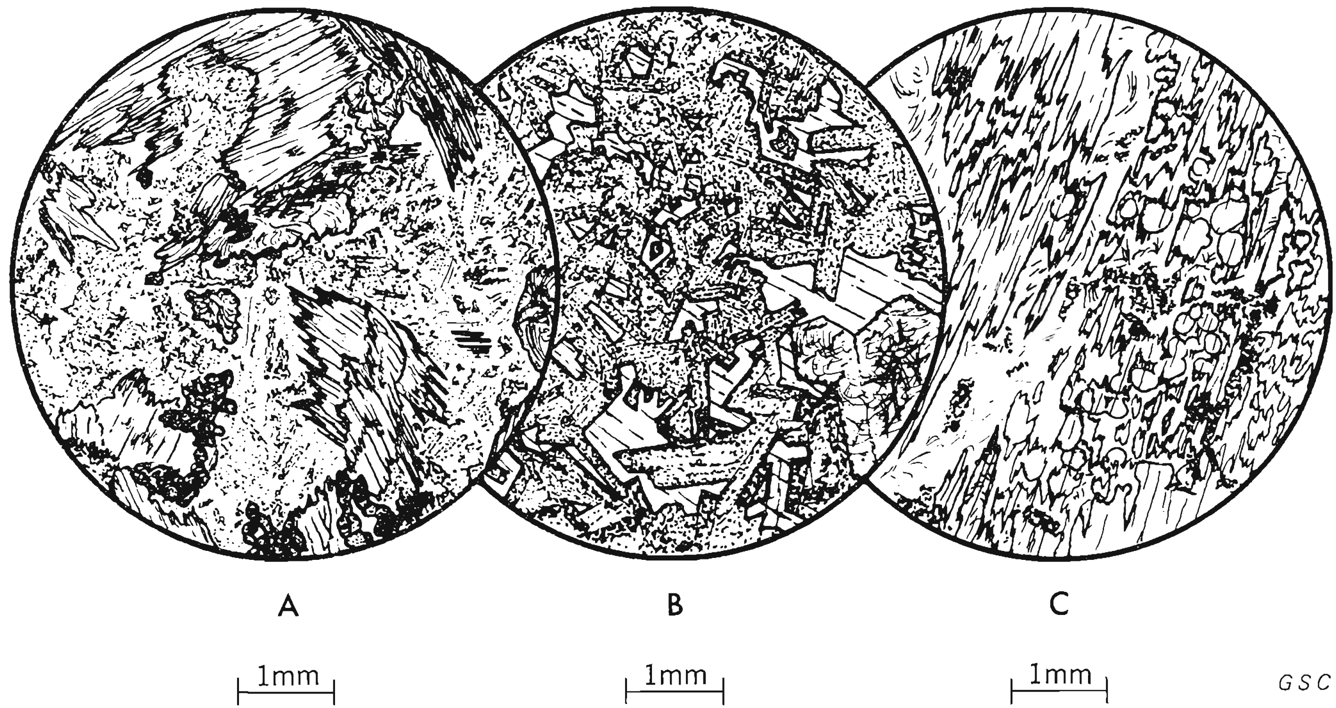



FIGURE 26. Meta-gabbro, leucometa-gabbro, and melanometa-gabbro.

A. Meta-gabbro showing shreddy-edged crystals of actinolite in a groundmass of saussuritized plagioclase. Chlorite patches contain finely acicular actinolite, and clusters of sphene crystals show relict cores of opaque iron oxide. Specimen B71.

B. Leucometa-gabbro showing the dominance of heavily saussuritized plagioclase in the rock. Tremolite forms large ophitic crystals of widely separated parts. One patch of chlorite contains numerous acicular crystals of tremolite. Specimen R336.

C. Melanometa-gabbro showing circular structures in a large actinolite crystal. This is presumed to be a pseudomorph of ophitic pyroxene with its contained olivine. Chlorite and scattered patches of saussuritized plagioclase form the areas of lower relief. Specimen R337.

forms small granular patches, some of which have relic cores of opaque iron oxide. Volumetric analysis of one specimen (B71) is:

| | |
|---|---|
| albite | 22 per cent |
| epidote | 21 per cent |
| actinolite | 48 per cent |
| chlorite | 5 per cent |
| sphene | 4 per cent |

Figure 26A is a typical microscopic view of meta-gabbro.

*Chemistry*

Chemical analyses of meta-gabbros are given in Table IX. Three of the analyses (B-71, 121-50, 46-50 + 6-50)[1] are of the general type of meta-gabbro and one (202-50 + 159-50)[1] is of porphyritic or blotchy meta-gabbro associated with porphyritic basalts of the Willbob Lake Formation. An average of the three analyses of the general type of meta-gabbro (46-50 + 6-50, B71, 121-50) is compared in Table VI with the average of fourteen analyses of the normal gabbro series (including the basalt B54A). The latter does not include the analyses of granitic veins (BL9-206, BL9-227), which are quantitatively negligible. The most significant difference between the averages is a substantially lower iron-magnesium ratio in meta-gabbro, despite higher silica and alkalis.

TABLE VI

*Comparison of Normal and Meta-gabbros*

*(Recalculated water-free, to 100 per cent)*

| | 1 | 2 |
|---|---|---|
| $SiO_2$ | 52.12 | 49.74 |
| $TiO_2$ | 0.67 | 1.56 |
| $Al_2O_3$ | 12.83 | 14.47 |
| $Fe_2O_3$ | 2.45 | 1.90 |
| FeO | 9.75 | 12.45 |
| MnO | 0.14 | 0.25 |
| MgO | 8.76 | 6.86 |
| CaO | 9.67 | 10.20 |
| $Na_2O$ | 2.86 | 2.20 |
| $K_2O$ | 0.42 | 0.22 |
| $P_2O_5$ | 0.14 | 0.15 |
| Rest | 0.19 | |

1. Average meta-gabbro (3 analyses).
2. Average normal gabbro (14 analyses).

[1] Analyses of samples 121-50, 46-50 + 6-50, and 202-50 + 159-50, from the adjoining Griffis Lake area, were generously provided by W. F. Fahrig.

Minor elements of typical meta-gabbro must be represented by the single analysis B71 (Table X). Little similarity is to be seen with analyses of normal gabbro except that strontium is extremely low. Otherwise the minor element values fit poorly the curves for normal gabbros.

### *Relation to Normal Gabbro Series*

Meta-gabbro sills are interpreted as metamorphic equivalents of the normal gabbro sills for the following reasons. (1) All sills were emplaced in the same tectonic environment (Labrador geosyncline) during the same tectonic phase of development (pre-folding). Together the sills form a continuous swarm that extends for 350 miles along the Labrador Trough. It is hardly credible that those sills that now fall within the greenschist facies zone should also have had a separate origin. (2) Chemical data from the meta-gabbros is meagre; nevertheless they show some similarity with those of the normal gabbro series, particularly in regard to the major elements and strontium. (3) Gabbros emplaced stratigraphically below the Denault Formation appear on both sides of a metamorphic boundary (coincident with a fault) that passes through Lac Low. On the north side of the boundary the sills are typical meta-gabbros, on the south they are typical normal gabbros.

## Melanometa-gabbro and Leucometa-gabbro

### *Petrographic Description*

Melanometa-gabbro consists essentially of large (0.5–3 cm), interlocking, equidimensional actinolite crystals crowded with randomly oriented, dusty albite–epidote laths and scattered irregular patches of colourless to pale green chlorite. The amphibole is colourless to light green and in a single crystal may change sharply from one to the other. Some crystals contain curious circular structures marked by slight discontinuities between amphibole within and outside the structure (Fig. 26C). The two are in optical continuity, but may show a slight colour difference. Where circular structures are in contact, the amphibole passes from one structure to the next without interruption and areas between the structures contain shreddy amphibole or chlorite wedges. These textures are interpreted as pseudomorphs of augite poikilitically enclosing olivine, which was apparently abundant. Volumetric analyses of melanometa-gabbro are included in Table VII. In view of the coarse grain size these analyses are estimates.

Leucometa-gabbro owes its light colour to the chalky white saussuritized plagioclase and pale to colourless actinolite of which it is predominantly formed. In thin section, it is seen to be a confused intergrowth of amphibole, epidote, albite, chlorite, and sphene. Primary forms of plagioclase crystals are partly preserved locally by epidote–albite intergrowths rimmed with clear albite, but textures of former ferromagnesian minerals have been obliterated. Portions of some thin sections consist of closely packed pseudomorphs of plagioclase almost to the exclusion of ferromagnesian minerals. Actinolite is colourless to pale green and commonly forms large ophitic crystals of widely separated parts. Chlorite is colourless in thin section and occurs in scattered patches. Table VII contains two volumetric analyses of leucometa-gabbro and Figure 26B shows part of a typical thin section.

The gradation from melanometa-gabbro to leucometa-gabbro in the Keato Lake sill is marked by an increase in plagioclase and a reduction in both colour and quantity of mafic minerals. Volumetric analyses (Table VII, R471–R468) of specimens taken across the narrow gradational zone, representing a stratigraphic width of about 200 feet, show the change in mineralogy. Between specimens R470 and R471 the major part of the change takes place, but the change in colour is more evenly spread over the four specimens. The colour change is evidently due to variation in the iron content of the amphibole and possibly chlorite. Refractive index and specific gravity determinations on amphiboles in the set of specimens from the transition zone (Baragar, 1960, p. 1618) indicates a reduction in iron content corresponding to diminution in colour. The compositional span ranges from ferrotremolite 38 in melanometa-gabbro to ferrotremolite 20 in leucometa-gabbro.

TABLE VII

*Volumetric Analyses of Melanometa-gabbro and Leucometa-gabbro*

| | R337[1] | R471[2] | R470[2] | R469[2] | R468[2] | R293[3] |
|---|---|---|---|---|---|---|
| Albite | 13 | 3 | 9 | 4 | 5 | 6 |
| Epidote and albite | 20 | 13 | 53 | 52 | 56 | 63 |
| Amphibole | 53 | 52 | 21 | 26 | 34 | 18 |
| Iron ores | 2 | 1 | 1 | 1 | 2 | 1 |
| Chlorite | 12 | 31 | 16 | 17 | 3 | 12 |

[1]Melanometa-gabbro, lower part of Valley Ponds sill.

[2]Specimens from lower to upper zone of Keato Lake sill: R471, melanometa-gabbro; R470 and R469, transitional rocks; R468, leucometa-gabbro.

[3]Leucometa-gabbro, upper part of Valley Ponds sill.

### *Chemistry*

Leucometa-gabbro and melanometa-gabbro have unusual compositions. In Table VIII they are compared with similar analyses found in the literature. Melanometa-gabbro closely resembles an olivine-rich basalt from Hawaii; the closest match found for the leucometa-gabbro is a saussurite–smaragdite gabbro from the Swiss

Alps. Descriptions of the latter by Bonney (1892) could to a large extent apply to the leucometa-gabbro of this area. Norms of melanometa- and leucometa-gabbro (Table IX) confirm that the former was rich in olivine and the latter in plagioclase.

TABLE VIII

*Melanometa-gabbro, Leucometa-gabbro, and Other Analyses*

*(Recalculated water-free, to 100 per cent)*

| | 1 | 2 | 3 | 4 |
|---|---|---|---|---|
| $SiO_2$ | 47.90 | 47.50 | 48.05 | 45.89 |
| $TiO_2$ | 0.39 | 0.70 | 0.19 | 1.81 |
| $Al_2O_3$ | 23.52 | 14.56 | 24.60 | 15.36 |
| $Fe_2O_3$ | 1.45 | 1.16 | 2.30 | 1.22 |
| FeO | 4.32 | 9.14 | 1.80 | 8.12 |
| MnO | 0.07 | 0.18 | 0.05 | 0.08 |
| MgO | 6.75 | 15.21 | 8.71 | 13.22 |
| CaO | 12.78 | 9.19 | 11.17 | 10.71 |
| $Na_2O$ | 2.72 | 2.04 | 2.80 | 2.30 |
| $K_2O$ | 0.10 | 0.06 | 0.28 | 0.67 |
| $P_2O_5$ | tr. | 0.26 | 0.05 | 0.62 |

1. Leucometa-gabbro R293, Wakuach Lake map-area.
2. Melanometa-gabbro R337, Wakuach Lake map-area.
3. Saussurite–smaragdite gabbro, moraine of Allalin glacier near Zermat, Switzerland (Burri and Niggli, 1945, p. 597).
4. Olivine basalt lava north slope Hualalai, Hawaii (Washington, 1923, p. 102).

The minor elements show an anomalous behaviour in melanometa-gabbro–leucometa-gabbro sills. In a series of specimens from melanometa-gabbro to leucometa-gabbro in the Keato Lake sill (Table X, R-472–R468), chromium increases steadily whereas normally it is concentrated in the ferromagnesian-rich fractions and diminishes towards the felsic fraction. Nickel behaves normally, with high values in melanometa-gabbro decreasing towards leucometa-gabbro, but vanadium varies anomalously from high values in leucometa-gabbro to low values in melanometa-gabbro. The behaviour of other trace elements is compatible with the changes in mineralogy. Strontium and gallium are both higher in leucometa-gabbro, whereas lithium is lower. Variation of trace elements from melanometa-gabbro (R337) to leucometa-gabbro (R293) in the Valley Ponds sill confirms the pattern found in the Keato Lake sill, except that chromium is about the same in mafic and salic portions. Barium, zirconium, and scandium all show marked increases in the uppermost specimen of the Keato Lake sill, but are approximately the same in upper and lower parts of the Valley Ponds sill.

# TABLE IX

*Chemical Analyses and Norms of Wakuach Gabbros*

| Specimen No. | Normal Gabbros | | | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | B279 | B295 | R90C | R189 | B345 | R129 | B286 | R127 | R195 | R93 | R151 | R94 | R554 | BL9-227 | BL9-206 |
| $SiO_2$ | 46.06 | 45.67 | 49.12 | 44.50 | 46.95 | 48.99 | 49.97 | 50.10 | 44.48 | 46.73 | 48.82 | 50.12 | 54.02 | 63.09 | 74.16 |
| $TiO_2$ | 0.58 | 0.73 | 0.75 | 0.77 | 0.80 | 1.39 | 0.76 | 1.31 | 2.81 | 2.86 | 2.33 | 2.59 | 2.08 | 1.76 | 0.27 |
| $Al_2O_3$ | 17.86 | 16.68 | 13.77 | 18.90 | 17.92 | 12.42 | 13.14 | 12.83 | 11.44 | 12.05 | 12.25 | 10.83 | 11.15 | 13.13 | 12.47 |
| $Fe_2O_3$ | 1.29 | 1.40 | 1.17 | 1.26 | 1.48 | 2.61 | 2.00 | 2.07 | 1.19 | 2.25 | 4.01 | 1.96 | 1.44 | 0.49 | 1.09 |
| $FeO$ | 7.81 | 8.55 | 8.33 | 7.74 | 7.77 | 12.74 | 11.42 | 12.17 | 17.46 | 14.66 | 13.88 | 18.16 | 16.19 | 2.01 | 1.32 |
| $MnO$ | 0.17 | 0.18 | 0.20 | 0.18 | 0.20 | 0.25 | 0.26 | 0.23 | 0.26 | 0.28 | 0.26 | 0.24 | 0.36 | 0.06 | 0.08 |
| $MgO$ | 10.58 | 10.24 | 9.71 | 7.99 | 7.15 | 7.41 | 5.74 | 5.93 | 6.16 | 5.57 | 3.76 | 2.97 | 2.36 | 1.36 | 0.66 |
| $CaO$ | 10.68 | 10.67 | 13.18 | 12.17 | 12.15 | 8.55 | 10.94 | 10.75 | 10.37 | 10.39 | 7.19 | 4.26 | 6.59 | 7.90 | 2.50 |
| $Na_2O$ | 1.37 | 1.44 | 1.34 | 1.80 | 1.86 | 2.34 | 2.19 | 2.30 | 1.71 | 1.83 | 4.26 | 2.76 | 2.57 | 5.82 | 5.97 |
| $K_2O$ | 0.23 | 0.10 | 0.05 | 0.26 | 0.32 | 0.12 | 0.28 | 0.12 | 0.31 | 0.48 | 0.01 | 0.19 | 0.18 | 0.04 | 0.04 |
| $P_2O_5$ | 0.23 | 0.05 | 0.05 | 0.03 | 0.12 | 0.00 | 0.11 | 0.00 | 0.06 | 0.30 | 0.10 | 0.59 | 0.25 | 0.21 | 0.04 |
| $Cr_2O_3$ | 0.02 | tr. | 0.13 | 0.16 | tr. | n.d. | tr. | n.d. | tr. | n.d. | n.d. | tr. | n.d. | | |
| $H_2O+$ | 3.26 | 3.99 | 2.20 | 3.76 | 3.09 | 3.73 | 2.82 | 2.57 | 3.03 | 2.62 | 3.34 | 4.08 | 3.02 | 0.68 | 0.54 |
| $H_2O-$ | 0.25 | 0.36 | 0.16 | 0.25 | 0.20 | 0.09 | 0.12 | 0.09 | 0.24 | 0.11 | 0.43 | 0.35 | 0.00 | 0.04 | 0.14 |
| $CO_2$ | 0.01 | 0.08 | 0.05 | 0.01 | 0.01 | 0.00 | 0.27 | 0.00 | 0.04 | 0.00 | n.d. | 0.22 | 0.00 | 3.18 | 0.38 |
| S | 0.01 | 0.07 | 0.06 | 0.02 | 0.03 | n.d. | 0.15 | n.d. | 0.18 | n.d. | n.d. | 0.17 | n.d. | n.d. | n.d. |
| Total | 100.40 | 100.21 | 100.27 | 99.70 | 100.05 | 100.64 | 100.17 | 100.47 | 99.74 | 100.13 | 100.64 | 99.65 | 100.21 | 99.82 | 99.86 |
| Less O = S | 0.01 | 0.03 | 0.03 | 0.01 | 0.01 | | 0.06 | | 0.08 | | | 0.07 | | | |
| Net Total | 100.39 | 100.18 | 100.24 | 99.69 | 100.04 | 100.64 | 100.11 | 100.47 | 99.66 | 100.13 | 100.64 | 99.58 | 100.21 | | |
| Sp. gr. | 2.98 | 3.01 | 3.06 | 2.99 | 3.02 | 2.98 | 3.04 | 3.02 | 3.09 | 3.05 | 2.99 | 3.08 | 2.97 | | |
| Mafic Index | 46.2 | 49.3 | 49.4 | 53.0 | 56.4 | 67.5 | 70.0 | 70.7 | 75.0 | 75.2 | 82.7 | 87.0 | 88.0 | 64.7 | 78.5 |
| Alkali-Mafic Index | | | | | | | | | | | | | | 60.4 | 66.2 |

| Specimen No. | Norms | | | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | Normal Gabbros | | | | | | | | | | | | | | |
| | B279 | B295 | R90C | R189 | B345 | R129 | B286 | R127 | R195 | R93 | R151 | R94 | R554 | BL9-227 | BL9-206 |
| Qu | — | — | — | — | — | 0.90 | 2.66 | 2.95 | — | 0.73 | — | 8.64 | 12.46 | 21.06 | 34.14 |
| Or | 1.34 | 0.61 | 0.28 | 1.56 | 1.89 | 0.72 | 1.67 | 0.72 | 1.84 | 2.84 | 0.06 | 1.11 | 1.06 | — | — |
| Ab | 11.59 | 12.17 | 11.33 | 15.21 | 15.74 | 19.77 | 18.52 | 19.46 | 14.48 | 15.47 | 36.03 | 23.34 | 21.77 | 49.25 | 50.83 |
| An | 41.93 | 38.76 | 31.44 | 42.74 | 39.62 | 23.04 | 25.21 | 24.32 | 22.62 | 23.26 | 14.27 | 16.61 | 18.36 | 9.45 | 7.23 |
| Di Wo | 3.97 | 5.56 | 13.89 | 7.27 | 8.33 | 8.10 | 11.12 | 12.12 | 11.83 | 11.00 | 8.67 | 1.43 | 5.29 | 3.48 | 1.04 |
| Di En | 2.52 | 3.43 | 8.48 | 4.30 | 4.75 | 3.90 | 4.88 | 5.31 | 4.41 | 4.44 | 2.96 | 0.32 | 1.05 | 2.70 | 0.60 |
| Di Fs | 1.20 | 1.81 | 4.63 | 2.61 | 3.22 | 4.09 | 6.21 | 6.78 | 7.64 | 6.66 | 5.95 | 1.20 | 4.62 | 0.40 | 0.40 |
| Hy En | 12.95 | 12.29 | 15.45 | 0.49 | 7.20 | 14.57 | 9.42 | 9.46 | 5.42 | 9.43 | 5.35 | 7.08 | 4.82 | 0.70 | 1.00 |
| Hy Of | 6.19 | 6.48 | 8.46 | 0.30 | 4.87 | 15.32 | 12.02 | 12.10 | 9.37 | 14.18 | 10.73 | 26.32 | 21.15 | 0.13 | 0.66 |
| Ol Fo | 7.62 | 6.85 | 0.17 | 10.59 | 4.10 | — | — | — | 3.86 | — | 0.74 | — | — | — | — |
| Ol Fa | 4.00 | 3.99 | 0.10 | 7.12 | 3.05 | — | — | — | 7.37 | — | 1.64 | — | — | — | — |
| Il | 1.11 | 1.38 | 1.43 | 1.46 | 1.52 | 2.64 | 1.44 | 2.49 | 5.34 | 5.43 | 4.43 | 4.92 | 3.97 | 3.34 | 0.61 |
| Mt | 1.88 | 2.04 | 1.69 | 1.83 | 2.15 | 3.77 | 2.89 | 3.01 | 1.74 | 3.27 | 5.81 | 2.85 | 2.08 | 0.70 | 1.62 |
| Ap | 0.53 | 0.13 | 0.13 | 0.07 | 0.26 | — | 0.26 | — | 0.13 | 0.69 | 0.23 | 1.38 | 0.59 | 0.50 | 0.01 |
| Pyr | 0.04 | 0.20 | 0.14 | 0.04 | 0.05 | — | 0.33 | — | 0.38 | — | — | 0.38 | — | — | — |
| Calc | 0.02 | 0.18 | 0.11 | 0.02 | 0.02 | — | 0.61 | — | 0.09 | — | — | 0.50 | — | 7.20 | 0.90 |
| Chrom | — | — | 0.20 | 0.09 | — | — | — | — | — | — | — | — | — | — | — |
| $H_2O$ | 3.51 | 4.35 | 2.36 | 4.01 | 3.29 | 3.82 | 2.94 | 2.66 | 3.27 | 2.73 | 3.77 | 4.43 | 3.02 | 0.72 | 0.68 |
| Total | 100.40 | 100.23 | 100.29 | 99.71 | 100.06 | 100.64 | 100.18 | 100.48 | 99.79 | 100.13 | 100.64 | 100.51 | 100.24 | 99.63 | 99.72 |

Analyst: J. A. Maxwell, Geol. Survey of Canada: B295, B286; D. E. McKay, Geol. Survey of Canada: B279, R90C, R189, R195, R94; S. Courville, Geol. Survey of Canada: B345, BL9-227, BL9-206; H. B. Wiik, Helsinki, Finland: R129, R127, R93; W. R. A. Baragar: R151, R554.

TABLE IX (*conc.*)

*Chemical Analyses and Norms of Wakuach Gabbros*

| | Basalt | | Leopard Rock | | | | Meta-gabbros | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | B54A | Composite meta-basalt | R161 | R163 | R177 | R181 | R337 | R293 | 46–50 6–50 | B71 | 121–50 | 202–50 159–50 |
| $SiO_2$ | 48.52 | 49.86 | 48.03 | 48.92 | 47.43 | 45.71 | 44.88 | 46.77 | 50.60 | 52.61 | 49.20 | 46.90 |
| $TiO_2$ | 1.28 | 0.53 | 1.58 | 1.56 | 0.91 | 0.25 | 0.67 | 0.38 | 0.28 | 0.93 | 0.74 | 0.84 |
| $Al_2O_3$ | 14.36 | 15.07 | 14.87 | 16.35 | 20.57 | 29.04 | 13.73 | 23.09 | 9.78 | 13.57 | 14.15 | 18.19 |
| $Fe_2O_3$ | 1.40 | 2.39 | 0.84 | 0.81 | 1.33 | 2.11 | 1.10 | 1.42 | 1.08 | 0.98 | 5.12 | 3.58 |
| FeO | 11.34 | 9.63 | 12.27 | 12.60 | 7.44 | 0.74 | 8.61 | 4.24 | 9.68 | 7.84 | 10.95 | 7.71 |
| MnO | 0.22 | 0.37 | 0.14 | 0.19 | 0.15 | 0.03 | 0.17 | 0.07 | 0.12 | 0.21 | 0.09 | 0.07 |
| MgO | 7.33 | 6.34 | 7.15 | 3.82 | 4.08 | 1.38 | 14.34 | 6.64 | 11.81 | 8.82 | 4.99 | 6.20 |
| CaO | 10.21 | 9.64 | 5.88 | 9.75 | 11.32 | 14.22 | 8.66 | 12.55 | 11.73 | 7.80 | 8.73 | 11.23 |
| $Na_2O$ | 2.05 | 1.76 | 2.05 | 2.26 | 2.80 | 2.63 | 1.95 | 2.66 | 1.88 | 3.90 | 2.56 | 1.60 |
| $K_2O$ | 0.24 | 0.43 | 2.62 | 0.46 | 0.57 | 1.23 | 0.06 | 0.10 | 0.34 | 0.14 | 0.75 | 0.53 |
| $P_2O_5$ | 0.09 | 0.20 | 0.21 | 0.08 | 0.19 | 0.03 | 0.25 | tr. | 0.08 | 0.09 | 0.24 | 0.23 |
| $Cr_2O_3$ | tr. | tr. | 0.02 | n.d. | tr. | n.d. | 0.07 | n.d. | 0.09 | 0.03 | n.d. | n.d. |
| $H_2O$ | 2.88 | 3.18 | 4.02 | 2.63 | 2.85 | 2.56 | 4.91 | 1.92 | 1.89 | 2.62 | 1.46 | 2.24 |
| $H_2O$ | 0.21 | 0.34 | 0.27 | 0.08 | 0.19 | 0.17 | 0.11 | 0.00 | 0.11 | 0.13 | 0.31 | 0.18 |
| $CO_2$ | 0.00 | tr. | 0.00 | 0.05 | 0.00 | 0.00 | 0.09 | n.d. | 0.00 | 0.00 | 0.36 | 0.27 |
| S | 0.06 | 0.15 | 0.03 | n.d. | 0.03 | n.d. | 0.06 | n.d. | 0.10 | 0.00 | 0.18 | 0.16 |
| Total | 100.19 | 99.89 | 99.98 | 99.56 | 99.77 | 100.10 | 99.66 | 99.84 | 99.57 | 99.67 | 99.91 | 99.93 |
| Less O≡S | 0.03 | 0.06 | 0.01 | | 0.01 | | 0.03 | | 0.04 | | 0.07 | 0.06 |
| Net Total | 100.16 | 99.83 | 99.97 | 99.56 | 99.76 | 100.10 | 99.63 | 99.84 | 99.53 | 99.67 | 99.84 | 99.87 |
| Sp. gr. | 3.02 | 3.02 | | | 2.99 | | 2.95 | 2.96 | | 2.94 | | |
| Mafic Index | 63.7 | 65.5 | 64.7 | 76.8 | 68.3 | | 40.3 | 46.0 | 47.7 | 50.0 | 76.3 | 64.5 |

| | Norms | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | Basalt | | Leopard Rock | | | | Meta-gabbros | | | | | |
| | B54A | Composite meta-basalt | R161 | R163 | R177 | R181 | R337 | R293 | 46–50 6–50 | B71 | 121–50 | 202–50 159–50 |
| Qu | — | 4.64 | — | 1.73 | — | — | — | — | — | — | 3.50 | 1.80 |
| Or | 1.39 | 2.56 | 15.48 | 2.73 | 3.40 | 7.29 | 0.33 | 0.61 | 2.00 | 0.84 | 4.45 | 3.12 |
| Ab | 17.36 | 14.90 | 17.36 | 19.14 | 23.71 | 11.28 | 16.52 | 20.09 | 15.89 | 32.99 | 21.66 | 13.53 |
| An | 29.27 | 31.94 | 23.62 | 33.11 | 41.90 | 63.80 | 28.55 | 50.78 | 17.25 | 19.11 | 24.90 | 40.90 |
| Ne | — | — | — | — | — | 5.94 | — | 1.31 | — | — | — | — |
| Di {Wo | 8.70 | 6.09 | 1.74 | 5.94 | 5.69 | 2.75 | 5.09 | 4.80 | 16.87 | 7.95 | 6.08 | 4.81 |
| Di {En | 4.32 | 3.03 | 0.82 | 1.96 | 2.67 | 2.38 | 3.39 | 3.23 | 10.21 | 4.80 | 2.78 | 2.82 |
| Di {Fs | 4.21 | 2.93 | 0.90 | 4.17 | 2.96 | — | 1.32 | 1.20 | 5.74 | 2.72 | 3.25 | 1.76 |
| Hy {En | 12.97 | 12.75 | 8.71 | 7.55 | 1.47 | — | 8.18 | — | 12.35 | 13.00 | 9.65 | 12.62 |
| Hy {Of | 12.67 | 12.28 | 9.50 | 16.07 | 1.64 | — | 3.19 | — | 6.94 | 7.35 | 11.22 | 7.86 |
| Ol {Fo | 0.68 | — | 5.80 | — | 4.22 | 0.74 | 16.91 | 9.32 | 4.80 | 2.92 | — | — |
| Ol {Fa | 0.73 | — | 6.97 | — | 5.16 | — | 7.25 | 3.80 | 2.98 | 1.82 | — | — |
| Il | 2.43 | 1.00 | 3.01 | 2.96 | 1.73 | 0.49 | 1.28 | 0.71 | 0.53 | 1.76 | 1.40 | 1.59 |
| Mt | 2.04 | 3.47 | 1.23 | 1.18 | 1.92 | 1.71 | 1.60 | 2.06 | 1.58 | 1.41 | 7.43 | 5.19 |
| Ap | 0.20 | 0.46 | 0.49 | 0.26 | 0.23 | 0.07 | 0.59 | — | 0.20 | 0.20 | 0.56 | 0.53 |
| Pyr | 0.14 | 0.31 | 0.05 | — | 0.07 | — | 0.17 | — | 0.03 | — | 0.38 | 0.34 |
| Calc | — | — | — | 0.11 | — | — | 0.20 | — | — | — | 0.82 | 0.61 |
| Chrom | — | — | 0.02 | — | — | — | 0.11 | — | 0.13 | 0.05 | — | — |
| Hem | — | — | — | — | — | 0.93 | — | — | — | — | — | — |
| $H_2O$ | 3.09 | 3.52 | 4.29 | 2.71 | 3.04 | 2.73 | 5.02 | 1.92 | 2.00 | 2.75 | 1.77 | 2.42 |
| Totals | 100.20 | 99.88 | 99.99 | 99.62 | 99.81 | 100.11 | 99.70 | 99.83 | 99.50 | 99.67 | 99.85 | 99.90 |

Analyst: J. A. Maxwell, Geol. Survey Canada: B54A, R337; D. E. McKay, Geol. Survey Canada: R161, R177; S. Courville, Geol. Survey Canada: B71; R. G. Fabry, Geol. Survey Canada: Composite metabasalt (Fahrig, 1953, p. 8); 46-50, 6–50; 121–50; 202–50, 159–50; unpublished analyses provided by W. F. Fahrig; W. R. A. Baragar: R163, R181. R293.

# TABLE X

*Geochemical Data (Expressed as parts per million)*

| Specimen No. | Normal Gabbro Series | | | | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | R90C | B279 | B295 | R189 | B345 | R195 | B54A | B286 | R127 | R93 | R129 | R151 | R554 | R94 | BL9-227[2] | BL9-206[2] |
| Si(x $10^{-3}$) | 234 | 222 | 223 | 217 | 226 | 214 | 234 | 242 | 240 | 225 | 247 | 234 | 260 | 246 | 297 | 349 |
| Al(x $10^{-3}$) | 76 | 98 | 92 | 104 | 98 | 63 | 78 | 72 | 70 | 66 | 68 | 67 | 61 | 60 | 70 | 67 |
| Fe'''(x $10^{-3}$) | 8 | 10 | 10 | 9 | 11 | 9 | 10 | 15 | 16 | 17 | 20 | 30 | 10 | 14 | 3 | 8 |
| Fe''(x $10^{-3}$) | 66 | 63 | 70 | 63 | 62 | 142 | 91 | 92 | 98 | 118 | 103 | 112 | 129 | 148 | 16 | 10 |
| Ca(x $10^{-3}$) | 97 | 79 | 80 | 91 | 90 | 77 | 75 | 81 | 79 | 76 | 64 | 53 | 49 | 32 | 57 | 18 |
| Mg(x $10^{-3}$) | 60 | 67 | 65 | 49 | 44 | 39 | 46 | 36 | 37 | 34 | 46 | 24 | 15 | 19 | 8 | 4 |
| Na(x $10^{-3}$) | 10 | 10 | 11 | 14 | 14 | 13 | 15 | 17 | 18 | 14 | 18 | 33 | 20 | 22 | 44 | 45 |
| K(x $10^{-3}$) | 0.5 | 2 | 1 | 2 | 3 | 3 | 2 | 2 | 1 | 4 | 1 | 0.1 | 2 | 2 | 0.3 | 0.3 |
| Ti(x $10^{-3}$) | 5 | 4 | 5 | 5 | 5 | 18 | 8 | 5 | 8 | 18 | 9 | 14 | 13 | 16 | 10.5 | 1.6 |
| P | 228 | 1000 | 228 | 131 | 524 | 262 | 383 | 480 | 0 | 1310 | 0 | 436 | 1140 | 2700 | 916 | 175 |
| Mn | 1800 | 1410 | 1490 | 1490 | 1800 | 2120 | 1800 | 2120 | 1880 | 2280 | 2040 | 2120 | 2900 | 1960 | 465 | 620 |
| Ga | 18 | 13 | 15 | 10 | 7 | 10 | 10 | 17 | 21 | 20 | 15 | 17 | 24 | 7 | | |
| Cr | 750 | 220 | 425 | 235 | 21 | 5 | 180 | 49 | 7 | 20 | 130 | - - | — | - - | | |
| Cr[1] | 1000 | 145 | 550 | 390 | 410 | 29 | 320 | 76 | | | | | | 17 | | |
| V | 245 | 195 | 250 | 220 | 225 | 1000 | 450 | 410 | 450 | 350 | 450 | 625 | — | — | — | — |
| Li | 5 | 5 | 12 | 12 | 8 | 10 | 5 | 12 | 12 | | 12 | 5 | 5 | 5 | | |
| Ni | 170 | 500 | 400 | 205 | 220 | 5 | 110 | 47 | 40 | 40 | 77 | - - | 3 | 4 | | |
| Ni[1] | 177 | 335 | 480 | 184 | 330 | 28 | 112 | 81 | | | | | | 15 | | |
| Sn[1] | 40 | 24 | 25 | 27 | 15 | 11 | 15 | 8 | | | | | | 14 | | |
| Cu[1] | 130 | 104 | 130 | 71 | 150 | 44 | 145 | 128 | | | | | | 120 | | |
| Co | 48 | 85 | 50 | 32 | 48 | 40 | 45 | 48 | 58 | 32 | 52 | 52 | 22 | 26 | | |
| Sc | 46 | | 30 | 12 | 12 | 47 | 55 | 46 | 54 | 26 | 54 | 35 | 24 | — | >100 | - - |
| Zr | 45 | 48 | 25 | 23 | 48 | 52 | 57 | 75 | 62 | 170 | 92 | 120 | 267 | 380 | 370 | 1480 |
| Y | 10 | 10 | — | 8 | 16 | 12 | 24 | 24 | 22 | 32 | 23 | 23 | 62 | 52 | | |
| Sr | 70 | 150 | 80 | 105 | 72 | 57 | 57 | 83 | 70 | 125 | 82 | 18 | 57 | 24 | 74 | 100 |
| Sr[1] | 53 | 108 | 96 | 113 | 99 | 64 | 56 | 101 | | | | | | 61 | | |
| Ba | 20 | 36 | 50 | 50 | 22 | 23 | 25 | 107 | 27 | 160 | 35 | 105 | 22 | 49 | 25 | 37 |
| Rb | - | 10 | — | 10 | 10 | 10 | — | — | — | | — | - - | — | - - | | |
| Modified Larsen function | −7.9 | −7.2 | −7.0 | −6.6 | −5.8 | −4.1 | −4.1 | −3.5 | −3.5 | −3.16 | −2.7 | +2.7 | +2.2 | +3.3 | +3.4 | +9.5 |

| Specimen No. | Leopard Rock | | | | | Meta-gabbros | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | R163 | R165 | R169 | R172 | R175 | B71 | R235 | R293 | R337 | R472 | R471 | R470 | R469 | R468 |
| Si(x $10^{-3}$) | 234 | | | | | 253 | | 223 | 222 | | | | | |
| Al(x $10^{-3}$) | 89 | | | | | 74 | | 124 | 73 | | | | | |
| Fe'''(x $10^{-3}$) | 6 | | | | | 7 | | 10 | 8 | | | | | |
| Fe''(x $10^{-3}$) | 101 | | | | | 63 | | 34 | 71 | | | | 12 | |
| Ca(x $10^{-3}$) | 72 | | | | | 57 | | 91 | 65 | | | | | |
| Mg(x $10^{-3}$) | 24 | | | | | 55 | | 41 | 92 | | | | | |
| Na(x $10^{-3}$) | 17 | | | | | 30 | | 22 | 15 | | | | | |
| K(x $10^{-3}$) | 4 | | | | | 1 | | 1 | 0.5 | | | | | |
| Ti(x $10^{-3}$) | 10 | | | | | 6 | | 2 | 4 | | | | | |
| P | 350 | | | | | 393 | | tr. | 114 | | | | | |
| Mn | 1570 | | | | | 1720 | | 548 | 1410 | | | | | |
| Ga | 15 | 12 | 15 | 15 | 24 | 5 | 15 | 17 | 5 | 5 | 5 | 12 | 12 | 17 |
| Cr | 120 | 105 | 117 | 200 | 115 | 245 | 190 | 457 | 510 | 160 | 180 | 250 | 505 | 550 |
| Cr[1] | 197 | 145 | 200 | 227 | 1158 | 385 | | | 820 | | | | | |
| V | 500 | 500 | 450 | 450 | 375 | 220 | 185 | 120 | 90 | 67 | 50 | 120 | 90 | 220 |
| Li | 12 | 25 | 12 | 8 | 12 | 12 | 12 | 8 | 15 | 12 | 12 | 5 | 5 | 5 |
| Ni | 170 | 220 | 110 | 210 | 122 | 190 | 30 | 190 | 550 | 435 | 435 | 210 | 240 | 70 |
| Ni[1] | 100 | 175 | 125 | 183 | 92 | 163 | | | 570 | | | | | |
| Sn[1] | 13 | 14 | 15 | 12 | 19 | 18 | | | 33 | | | | | |
| Sn | | | | | | | 15 | 15 | | 10 | 10 | 10 | 10 | 22 |
| Cu[1] | 142 | 165 | 105 | 124 | 112 | 16 | | | 45 | | | | | |
| Co | 52 | 68 | 45 | 50 | 49 | 45 | 29 | 32 | 53 | 49 | 62 | 40 | 52 | 30 |
| Sc | 55 | 48 | 33 | 40 | 28 | 46 | 46 | 14 | 12 | — | — | 12 | 10 | 47 |
| Zr | 100 | 92 | 100 | 87 | 51 | 52 | 42 | 26 | 24 | 26 | 27 | 28 | 26 | 51 |
| Y | 22 | 25 | 21 | 21 | 16 | 15 | 14 | — | — | — | — | — | — | 12 |
| Sr | 72 | 92 | 93 | 70 | 115 | 27 | 380 | 425 | 58 | 155 | 75 | 540 | 310 | 500 |
| Sr[1] | 80 | 108 | 118 | 89 | 155 | 44 | | | 103 | | | | | |
| Ba | 50 | 47 | 36 | 28 | 47 | 11 | 338 | 12 | 12 | 28 | 35 | 23 | 17 | 125 |
| Rb | 25 | 32 | 14 | — | 18 | — | 32 | — | — | — | — | — | — | — |

[1] Minor elements analyzed at spectrographic laboratories as follows: [1] Lamont Geological Observatory; [2] Geol. Survey of Canada; remainder at Cambridge University, England.

## Petrology and Petrogenesis

### Normal Gabbro Series

The normal gabbro series includes normal gabbros and basalts. Meta-gabbros, which are probably the metamorphosed counterpart of the normal gabbros, are less firmly related to the series on the basis of the meagre chemical data available, and are therefore excluded from the following discussion. Glomeroporphyritic gabbros and leucometa- and melanometa-gabbros are considered separately.

#### *Parent Magma*

The parent magma of the normal gabbro series is the one from which all members of the series were derived. Two lines of reasoning that point to the lavas as representatives of the unmodified parent magma are as follows:

(1) They stand at or near the beginning of the liquid line of descent (Fig. 24). Most porphyritic and ophitic gabbros that precede them are to some degree enriched in early-formed plagioclase, olivine, or pyroxene, and the rocks that immediately succeed them are subpegmatitic gabbros. It is evident that the lavas are bracketed by rock types enriched in opposing constituents from the standpoint of differentiation and therefore must be close to the initial composition.

(2) The true average of a differentiated series should be the same as the parent magma. It is impossible to compute the true average without knowing the quantities of each rock type involved, but the arithmetic average of the normal gabbro series analyses (excluding the quantitatively unimportant granitic veins) is probably a close approximation. The average analysis is compared with the lava analyses in Table XI where it may be noted that the similarity is striking. As the Menihek lava analysis is closer to the begin-

TABLE XI

*Average Normal Gabbro and Basalt Analyses*
*(Recalculated water-free, to 100 per cent)*

| | 1 | 2 | 3 |
|---|---|---|---|
| $SiO_2$ | 50.07 | 49.74 | 51.84 |
| $TiO_2$ | 1.32 | 1.56 | 0.55 |
| $Al_2O_3$ | 14.79 | 14.47 | 15.68 |
| $Fe_2O_3$ | 1.44 | 1.90 | 2.48 |
| FeO | 11.69 | 12.45 | 9.99 |
| MnO | 0.22 | 0.25 | 0.38 |
| MgO | 7.55 | 6.86 | 6.57 |
| CaO | 10.52 | 10.20 | 10.00 |
| $Na_2O$ | 2.06 | 2.20 | 1.82 |
| $K_2O$ | 0.25 | 0.22 | 0.45 |
| $P_2O_5$ | 0.09 | 0.15 | 0.24 |
| | 100.00 | 100.00 | 100.00 |

1. Basalt B54A, west of Murdoch Lake.
2. Average normal gabbro (14 analyses).
3. Composite sample of Doublet meta-basalt, Griffis Lake area (W. F. Fahrig, 1962, p. 4).

ning of the liquid line of descent, it will be assumed to represent the parent magma. Differentiation of the gabbros was evidently entirely in situ.

### *Trend of Differentiation*

The iron-enrichment trend characteristic of the normal gabbros is probably a result of a low initial water content in the magma. Kennedy (1955, pp. 500–502) argued that in "wet" magma iron is oxidized to a greater degree than in "dry" magma because of free oxygen available from dissociation of water. In a "wet" magma, iron crystallizes early as a spinel and the liquid changes in composition toward diorite. In a water-deficient magma, iron remains predominantly as ferrous iron and is concentrated in the liquid phase; hence the magma becomes progressively enriched in iron. These views were confirmed by work on the system $MgO–FeO–Fe_2O_3–SiO_2$ (Osborn, 1958, p. 1626). Osborn suggested that basaltic magmas of orogenic belts contain enough water to differentiate along the calc-alkaline trend, whereas basaltic magmas of sills and extrusives are poor in water and follow a trend of iron-enrichment. Few basaltic provinces show as extreme a trend as the Wakuach Gabbros (cf., Walker and Poldervaart, 1949, pp. 657–661); hence the writer concludes that the Wakuach magma was exceptionally "dry". Nevertheless, most of the sills contained sufficient water to produce coarse-grained pegmatitic phases in their upper portions.

### *Contrasting Differentiation Products*

Differentiation in the Wakuach Gabbros led generally to the development of pegmatitic gabbros and ultimately, in some cases, to granitic veins. One sill developed albite gabbro pegmatite, a distinctive rock chemically similar to gabbro pegmatite. Presumably albite gabbro pegmatite is the product of a parallel differentiation series derived from essentially the same parent magma, for it conforms closely with the trend curves of the normal gabbro series (Figs. 23 and 24). Its development must have depended upon special physical conditions or the presence of a substance not now apparent. In this connection the fine-grained, indeterminate mesostasis, so characteristic of the rock, is of interest. It is interpreted as a devitrified glass. Support of this view is found in Walker's reference to pegmatites in Tasmanian dolerite sills (1956, pp. 440–441), which carry devitrified glass in place of micropegmatite. Of the several conditions likely to lead to glass formation, escape of volatiles seems the most applicable to the present case. Thus the albite gabbro pegmatites may be the result of differentiation of an exceptionally dry phase of the magma.

### *Magma Type*

The parental basalt of the normal gabbro series, as represented by the analysis of Menihek lava (B54A), is compared in Table XII with the average analyses of parental basalts of the four major magma types: tholeiitic, alkaline, calc-alkaline, and high-alumina basalts. The resemblance between the parent magma of the Wakuach Gabbros and average parental tholeiite is striking. Parental alkali basalt has lower silica and higher alkalis, whereas the parent of the calc-alkaline series shows little similarity. Alumina in the Wakuach parent is too low to classify it as high-alumina basalt.

The mineralogy of the Wakuach Gabbros is also consistent with that of tholeiites. Olivine is present only in the early differentiated members, and quartz appears in members formed in the late stages of differentiation. Lime-poor pyroxene and augite formed throughout much of the crystallization period. In alkali basalts, on the other hand, olivine forms throughout the crystallization period, and augite is generally the sole pyroxene. In the late stages alkali feldspar, analcite, or a feldspathoid appears in place of quartz.

The bulk chemistry of the parental basalt shows little to distinguish it from other tholeiitic provinces but three features of the series are distinctive: (1) a differentiation trend showing extreme iron enrichment, (2) an extraordinary low potassium content throughout, and (3) a low strontium content throughout.

TABLE XII

*Wakuach Magma and Parental Tholeiitic, Alkali Basaltic, Calc-alkaline, and High-alumina Basaltic Magmas*

*(Calculated water-free, to 100 per cent)*

| | 1 | 2 | 3 | 4 | 5 |
|---|---|---|---|---|---|
| $SiO_2$ | 50.07 | 51.5 | 46.6 | 54.0 | 48.10 |
| $TiO_2$ | 1.32 | 1.5 | 2.9 | 0.9 | 0.89 |
| $Al_2O_3$ | 14.79 | 14.8 | 15.7 | 18.1 | 18.22 |
| $Fe_2O_3$ | 1.44 | 1.6 | 3.9 | 2.5 | 1.04 |
| $FeO$ | 11.69 | 9.9 | 8.8 | 5.8 | 8.31 |
| $MnO$ | 0.22 | - | 0.2 | 0.1 | 0.17 |
| $MgO$ | 7.55 | 7.2 | 7.7 | 5.5 | 8.96 |
| $CaO$ | 10.52 | 11.0 | 10.0 | 8.4 | 11.30 |
| $Na_2O$ | 2.06 | 2.0 | 2.9 | 3.4 | 2.80 |
| $K_2O$ | 0.25 | 0.5 | 0.9 | 1.1 | 0.14 |
| $P_2O_5$ | 0.09 | — | 0.4 | 0.2 | 0.07 |

1. Menihek basalt specimen B54A.
2. Average parental tholeiite, average of 6 parental tholeiites (Nockolds and Allen, 1956, p. 76).
3. Average parental alkali basalt. Average of 8 parental alkali basalts (Nockolds and Allen, 1954, p. 282).
4. Average parental calc-alkali magma. Average of 10 parental magmas (Nockolds and Allen, 1953, p. 139).
5. High-alumina basalt, Medicine Lake Highlands, California (Yoder and Tilley, 1957, p. 157).

## Glomeroporphyritic Gabbro

### *Genesis*

Glomeroporphyritic magmas appear to have been an integral part of the Labrador magmatic province. In the form of sills they are intimately associated in space and time with normal gabbros for a strike length of at least 300 miles along the Labrador Trough. As flows they are interlayered with Willbob Lake lavas for a distance that may be in excess of 50 miles. The marginal zones of glomeroporphyritic gabbro sills possess chemical idiosyncrasies (low potassium, low strontium) characteristic of the normal gabbros. It is difficult to escape the conclusion that glomeroporphyritic magma was simply a phase of normal magma development.

Conditions under which magmas may emerge enriched in feldspathic components have been studied by Verhoogen (1954, pp. 87–88), Yoder (1955, pp. 106–107), Buddington (1956, p. 100), and Yoder and Tilley (1962, pp. 448–453, 461–462). Most of these were discussed in a previous paper (Baragar, 1960, pp. 1632–1633), but Yoder's and Tilley's recent work provides much pertinent data not then available. Basalt, they find, is generally in the form of amphibolite when crystallized in the presence of excess water. Conversely if the amphibolite is melted under water pressure plagioclase is the first component to liquefy. If the liquid could be separated at this stage it would form a plagioclase-rich rock. Glomeroporphyritic gabbros, which contain a normative 70 per cent plagioclase (Table IX), would presumably emerge at a somewhat higher temperature when a greater proportion of the rock was liquid and as a consequence would be more easily separated. The composition of the plagioclase that results from early separation of the liquid phase would likely be andesine or oligoclase according to Yoder and Tilley. At a higher temperature labradorite, such as exists in the glomeroporphyritic gabbro, could be the expected plagioclase. Thus partial melting of a primary amphibolite under water pressure is evidently a condition that can give rise to magmas of the composition of glomeroporphyritic gabbros. Complete melting could yield material similar to the normal gabbro series magma.

### *Emplacement*

Regardless of the origin of the glomeroporphyritic gabbro it was evidently emplaced as a liquid heavily charged with plagioclase phenocrysts and aggregates of phenocrysts.

The flow characteristics of liquids charged with solid particles have been studied by Robertson and Mason (1956), Baines (1956), and Bhattacharji (Bhattacharji and Smith, 1963, 1964; and personal communication). Baines (1956, p. 30) discovered that in the flow of a dilute suspension of sulphite pulp in a smooth pipe the pulp concentrated in a central plug when the velocity increased beyond a critical value. The plug moved with a uniform velocity throughout and was surrounded by an annulus of water in laminar motion. Further increases in velocity eventually produced turbulent flow. Bhattacharji demonstrated in models of dykes and sills that flow produced an inward migration of solid particles with their eventual concentration in the centre. A plug was not obtained as in the other experiments, but with higher concentrations of solid particles it is possible that 'plug flow' may ultimately result.

Flow differentiation was postulated as the mechanism that produced ultramafic layers in the ultramafic–meta-gabbro sills of the Labrador Trough in an earlier section of this report. It may also be applied in the development of glomeroporphyritic gabbro. The 'plug flow' described by Baines admirably fits the picture of a glomeroporphyritic sill with its uniform core of feldspathic aggregates and marginal gabbro zones. However, in a sill the 'plug' would be tabular. When the movement ceased the magma would probably be sufficiently viscous to preserve the divisions of flow structure. The laminar structure at the boundary between glomeroporphyritic gabbro and the border phase in one outcrop previously described indicates that laminar flow existed in the border region. Elsewhere, the structure may have been obscured by slight settling of particles after flow had stopped.

This type of flow should exaggerate the concentration of plagioclase phenocrysts in a magma. As the flow proceeds the liquid phase should be continually withdrawn from the magma and fixed as chilled contacts. Thus inevitably the concentration of solid phase must increase.

### Leucometa-gabbro and Melanometa-gabbro

Sills grading stratigraphically upwards from melanometa-gabbro to leucometa-gabbro are not the product of normal differentiation for the following reasons. (1) Despite differences in colour, melanometa-gabbro has only a slightly lower mafic index (40) than leucometa-gabbro (46). (2) Normative plagioclase is nearly the same in both rocks ($An_{68}$ for leucometa-gabbro, $An_{61}$ for melanometa-gabbro). (3) Leucometa-gabbro is only slightly enriched in alkalis and silica relative to melanometa-gabbro. (4) Trace elements such as chromium and vanadium do not show the behaviour characteristic of a normal differentiation series.

Both rocks are obviously mafic, but one contains a preponderance of the lighter constituents of mafic rocks, and the other the heavier constituents. It is tempting to interpret the division of these sills into heavier and lighter fractions as due to gravity separation. However, even if it could be shown that the original magma were dense enough to float plagioclase, it is doubtful that so complete a separation could be obtained by simple floating and sinking. Possibly, intrusion of an extensively crystallized mafic magma would produce gravity separation by action analogous to that of a jig used in commercial mineral separation. Jostling of crystals against one another during magma movements would allow heavier crystals to work to the bottom, displacing lighter crystals upward. The liquid medium need not be heavier than the plagioclase crystals. Either heavy or light fractions may be sufficiently in excess of the other to continue beyond its termination in apparently independent sills. Such a method of origin is highly speculative. The only positive conclusion that can be reached is that leucometa-gabbro and melanometa-gabbro are not normal differentiation products.

## The Labrador Magmatic Province

### *Regional Distribution*

Petrological and petrochemical data supplied in this report are considered to be representative of geosynclinal igneous rocks in the southern part of the Labrador Trough. Data and observations of Sauvé (1957) from a corresponding igneous assemblage in the northern part of the Labrador Trough (Gerido Lake) are remarkably similar. Sauvé's assemblage includes: a profusion of differentiated gabbro sills, sills of blotchy gabbro (glomeroporphyritic gabbro), and pillowed and massive basalts with local blotchy (porphyritic) flows. Differentiation in the sills is marked by extreme iron enrichment similar to the Skaergaard trend, but with notable paucity in the potash content throughout (Sauvé, 1957, pp. 137–142). In detail the points of similarity between igneous assemblages in the two areas are extraordinarily numerous. There can be little doubt that the same magmatic province is common to the entire Labrador geosyncline. In summary, its principal distinguishing features are: (1) a differentiation trend marked by extreme enrichment in iron; (2) extraordinary deficiency in potassium and probably strontium throughout the series; and (3) the presence of a phase greatly enriched in plagioclase (glomeroporphyritic gabbro).

### *Tectonic Significance*

Magmas of the Labrador magmatic province constitute part of the geosynclinal assemblage. Whether they occur as sills or lavas has little significance in this context, for presumably a favourable variation in load and lateral pressures would convert one to the other. As geosynclinal magmas they have a well-recognized position in orogenesis (Benson, 1926; Harpum, 1960; and many others) but a less well-defined character. Many authors agreed that such magmas are basaltic or andesitic, commonly spilitic, and almost invariably accompanied by ultramafic intrusions (Benson, 1926; Harpum, 1960; Rittman, 1962, pp. 164–165). However, in most orogenic belts they are highly deformed and metamorphosed; accordingly their petrological and structural relationships are largely obscured. The mafic rocks are commonly indeterminate greenschists and the ultramafic rocks, serpentines. The term ophiolites (Steinmann, 1905) has been generally applied to such assemblages.

In the Labrador Trough what must be essentially the ophiolite association has partly survived the ravages of orogenesis with only minor alteration. As a consequence several interesting relationships have been brought to light.

(1) The magma was a common tholeiitic basalt most closely related chemically to such non-orogenic magmas as the Skaergaard (Wager and Deer, 1939) and Elephants Head–New Amalfi intrusions of the Karroo province (Poldervaart, 1944)[1]. No spilitic rocks have been found in the Labrador magmatic province to date.

(2) The gabbro or dolerite sills are petrologically similar to typical sills and swarms of sills of continental or non-orogenic association such as the Karroo dolerites (Walker and Poldervaart, 1949), the Palisade diabase (Walker, 1940), the Tasmanian dolerites (Edwards, 1942), and the Whin sill (Tomkeieff, 1929).

[1] These are compared with the Labrador magma in Baragar (1960, pp. 1636-1639).

(3) The ultramafic intrusions in the Labrador Trough occur as continuous sills at least 120 miles long in contrast with the swarms of lens-like ultramafic masses common to most orogenic belts (Turner and Verhoogen, 1960, p. 308). On the other hand, ultramafic bodies in the schists east of the Labrador Trough, where the deformation has been more intense, appear to be lens-like as in the other orogenic belts. Thus Rittman's (1962, p. 165) proposal that serpentine masses are sheared and broken by the intense folding that usually follows their emplacement tends to be borne out by the two sets of occurrences found in the Labrador orogenic belt.

## *Chapter V*

# METAMORPHISM

The effects of regional metamorphism on rocks of the Labrador Trough completely overshadow the all but negligible effects of thermal metamorphism adjoining the mafic intrusions. The former have penetrated to some degree the bulk of the Trough deposits whereas the latter are rarely more than a skin of minor alteration on the contact of sills. In at least one place a pod of thermally metamorphic rocks is conspicuous, but generally thermal metamorphism constitutes more of a problem in its apparent modesty than in its severity.

## Regional Metamorphism

Metamorphic zones of increasing grade succeed one another eastward across the Labrador Trough and into the schists and gneisses of uncertain origin beyond. The isograds roughly parallel the structure of the Trough in its wide central region, but north of Koksoak River (Bergeron, 1957, p. 106; Gross, 1961, p. 5) they cross it and involve even the westernmost deposits in high-ranking metamorphism. At the south end of the Trough they abut the metamorphic front of the Grenville Structural Province. In Wakuach Lake map-area, where the Labrador Trough is at its widest, the isograds are well separated and the metamorphic zones correspondingly broad.

On Figure 27 and in Table XIII the major features of regional metamorphism in this area are summarized. The western two-thirds of the Labrador Trough is underlain by a broad zone exhibiting weak, uneven metamorphic effects. This is called the subgreenschist facies zone in recognition of the incipient nature of the metamorphism. It is followed on the east by the greenschist facies zone, which overlaps the eastern boundary of the Trough, and in turn by the almandine amphibolite facies zone. A fault provides an abrupt boundary between the subgreenschist facies and greenschist facies zones in the northern half of the area. South of this the metamorphic boundary evidently gives way to a broad transitional zone of mixed metamorphic effects (Frarey, 1961, personal communication). The eastern boundary of the greenschist facies zone is placed between plagioclases of albite and oligoclase compositions. It is therefore a boundary of definition only; the rocks immediately on either side are lithologically indistinguishable. Bounding the almandine amphibolite facies zone on the east are granodiorite gneisses.

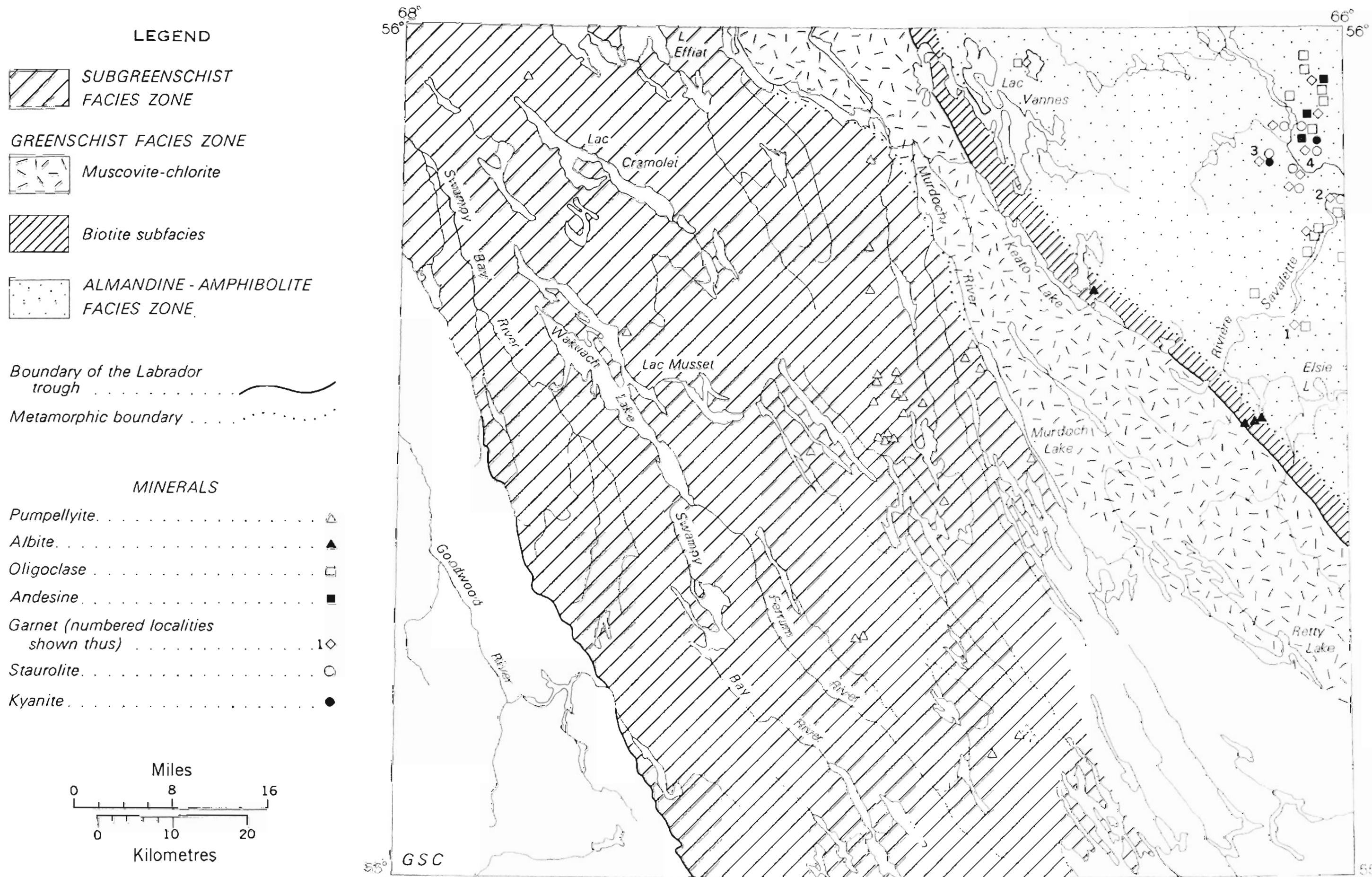



FIGURE 27. Regional metamorphic zones in Wakuach Lake map-area.

TABLE XIII

*Characteristic Features of Metamorphic Zones in Wakuach Lake Map-Area*

| Rocks | Subgreenschist Zone | Greenschist Facies Zone | | Almandine Amphibolite Facies |
|---|---|---|---|---|
| Mafic igneous rocks | Characteristic minerals: pumpellyite prehnite?<br><br>Texture: mainly primary | Assemblages: albite–epidote–actinolite–chlorite–sphene<br><br>Texture: completely recrystallized<br><br>Amphibole: colourless to pale green, rarely bluish green | | Assemblages: oligoclase–hornblende ± almandine; oligoclase–hornblende–epidote; andesine–hornblende–biotite–epidote<br><br>Texture: completely recrystallized<br><br>Amphibole: green |
| | | Muscovite–chlorite Subfacies | Biotite Subfacies | |
| Shales and greywackes | Minerals: no new minerals<br><br>Texture: primary, no evident recrystallization | Assemblages: quartz–albite–muscovite–chlorite; quartz–albite–muscovite–chlorite–epidote ± actinolite<br><br>Texture: completely recrystallized | Assemblages: quartz–albite–biotite; quartz–albite–muscovite–biotite<br><br>Texture: completely recrystallized | Assemblages: quartz–oligoclase–biotite ± almandine; quartz–oligoclase–biotite–muscovite; quartz–muscovite–biotite–staurolite–almandine (plagioclase); quartz–muscovite–biotite–staurolite–kyanite (plagioclase and almandine)<br><br>Texture: coarsely recrystallized |
| Iron-formation | Minerals (metamorphic? diagenetic?): minnesotaite, riebeckite, stilpnomelane<br><br>Texture: minutely seriate; "chert" grain size: 0.001–0.1 mm; metallic grain size: 0.005–0.05 mm | Minerals: characteristic minerals: acmite, actinolite, riebeckite, stilpnomelane<br><br>Texture: completely recrystallized; "chert" grain size: 0.01–0.05 mm; metallic grain size: 0.05–0.1 mm | | |

## Subgreenschist Facies Zone

Notable metamorphic effects in rocks of this zone are all but confined to iron-formation, carbonate rocks, gabbros, and basalts. The remaining rocks, largely quartzites, greywackes, and shales, show little recognizable evidence of regional metamorphism. Even among the former, the effects of diagenesis and deuteric alteration cannot be separated with certainty from those of regional metamorphism. Some of the difficulties were discussed under alteration of gabbros (p. 124). It was noted that the general pattern of alteration, an increase in intensity through succeeding stages of differentiation, is characteristic of deuteric alteration in differentiated sills. Yet not all the alteration could be attributed to this effect, for the earliest member of the series, porphyritic gabbro, as well as undifferentiated lavas, exhibits variable amounts of similar types of alteration. Another influence that is difficult to evaluate is the effect during intrusion of closely-spaced gabbro sills on one another and on adjoining rocks. Water, from sedimentary host rocks desiccated by the heat of the intrusion, might be expected to migrate into nearby earlier gabbros and lavas causing notable alteration. It is, therefore, with considerable uncertainty that features marking the subgreenschist facies zone of metamorphism are discussed. The least equivocal of these are considered in the following sections.

### *Pumpellyite*

Pumpellyite is widespread in basic igneous rocks in that part of the map-area underlain by rocks of the subgreenschist facies (Fig. 27), but is by no means ubiquitous. It is generally a secondary product in plagioclase and is found at all differentiation stages in gabbro as well as in basaltic lavas. In a few gabbro pegmatites, particularly albite gabbro pegmatites, it occurs in the interstices of the rock. Pumpellyite may accompany the saussurite typical of plagioclase alteration, but not uncommonly is alone or in association with finely crystalline, colourless minerals, one of which is tentatively identified as prehnite.

The pumpellyite is highly pleochroic from deep blue-green (Y) to pale brownish yellow (X and Z). Commonly it appears as sheaf-like clusters marked by fanning of (100) cleavages or thin crystal tablets. Seki and Kuriyagawa (1962, pp. 538–539) determined the optical properties of pumpellyite in a few of the writer's samples. These ranged as follows:

$\beta$ refractive index = 1.694–1.705, most 1.703–1.705

$2V = (+)\ 9^\circ$ to $44^\circ$, most $(+)\ 36^\circ$ to $44^\circ$

$a \wedge X = 19.5^\circ$ to $26^\circ$, most $24^\circ$ to $26^\circ$

The indices indicate an extraordinarily high content of ferric iron, according to the data of Coombs (1953) and Seki (1961).

### *Iron-Formation*

The metamorphism of the Sokoman Formation, previously described (p. 56), is characterized by heterogeneous recrystallization of what was originally chert and jasper. The texture is seriate, such that finely crystalline quartz with an average grain size of less than 0.001 mm is interspersed with microscopic patches in which the grain size may reach a maximum of 0.3 mm. According to James (1955, p. 1475), the grain size of recrystallized chert is one of the criteria of metamorphic grade in iron-formation. In the Sokoman Formation it is apparent that recrystallization is incomplete; grain sizes consistent with diagenesis are intimately associated with grain sizes characteristic of biotite grade. This probably denotes a metamorphic environment of sufficient energy to initiate, but not complete, recrystallization of chert.

The significance of iron-silicate minerals such as minnesotaite (see discussion p. 57), stilpnomelane, and riebeckite is unknown, since their full fields of stability have not yet been defined.

### *Carbonate Rocks*

Rocks of the various carbonate formations in the subgreenschist facies zone show a heterogeneity of grain sizes similar to the iron-formation. Generally the grain size is less than 0.05 mm, but in patches it ranges up to 0.3 mm. No new minerals were observed.

## Greenschist Facies Zone

All rocks in the greenschist facies zone have been completely recrystallized. No primary minerals remain and detailed textural relationships, such as particle and original crystal shapes, have been all but obliterated. The typical mineral assemblage of meta-gabbros and meta-basalts is albite–epidote–actinolite–chlorite–sphene. Chlorite is subordinate to actinolite, minor biotite is present, and quartz is rare in most of the rocks. The actinolite is mainly colourless to very pale green in thin section. No plagioclase more calcic than albite was found. One meta-basalt assemblage near the north boundary of the map-area comprises blue-green amphibole, carbonate, chloritoid, and sphene.

Sedimentary rocks in this zone fall into two groups: (1) those interbedded with volcanic rocks in the Labrador Trough, and (2) quartz–feldspar–biotite schists just east of the Trough. Mineral assemblages of the first group are mainly quartz–albite–muscovite–chlorite and quartz–albite–muscovite–chlorite–epidote ± actinolite. Epidote and actinolite presumably appear where the sediments are tuffaceous. Biotite occurs with chlorite and muscovite locally, but in all such places appears to be partly replaced by chlorite. The biotite may be a relic from earlier thermal metamorphism adjoining the sills. Characteristic mineral assemblages of the second group are quartz–albite–biotite, and quartz–albite–muscovite–biotite. Chlorite is either a minor replacement of biotite or is absent in each assemblage. Thus the two groups appear to define the quartz–albite–muscovite–chlorite and quartz–albite–epidote–biotite subfacies (Fyfe, *et al.*, 1958, pp. 219–223) of the greenschist facies. Their distribution is shown on Figure 27.

The metamorphism of the Doublet Group iron-formation has already been discussed (p. 74). In a comparison of it with the Sokoman Formation of the subgreenschist facies zone, the major changes attributable to metamorphism were noted to be: (1) an increase in the grain sizes of quartz, metallic minerals, and riebeckite; and (2) the development of acmite and tremolite–actinolite. Minnesotaite is not present and is presumably not stable, but there can be no certainty that it would have been an original mineral in this iron-formation in any event. The principal mineral assemblages are: (1) quartz–magnetite–hematite–amphibole (actinolite–riebeckite); (2) quartz–magnetite–hematite–acmite ± riebeckite; (3) quartz–ankerite–stilpnomelane; and (4) quartz–magnetite–hematite.

### Almandine Amphibolite Facies Zone

A plagioclase composition more calcic than albite is essential to the definition of this zone (Fyfe, *et al.*, 1958, p. 218). Accordingly, the western boundary of the zone was placed on the basis of plagioclase compositions determined in a succession of specimens taken eastward from the east boundary of the Trough. The locations and generalized compositions are shown on Figure 27.

The three major groups of rocks contained within this zone are: (1) quartz–feldspar–biotite schists; (2) amphibolites; and (3) garnet–staurolite–mica schists. Typical assemblages of each are, respectively, as follows: (1) quartz–oligoclase–biotite ± almandine and quartz–oligoclase–biotite–muscovite; (2) oligoclase–hornblende ± almandine, oligoclase–hornblende–epidote, and andesine–biotite–hornblende–epidote; and (3) quartz–oligoclase (andesine)–biotite–almandine, quartz–muscovite–biotite–staurolite–almandine (plagioclase), quartz–muscovite–biotite–staurolite–kyanite (plagioclase and almandine). These assemblages are characteristic of the staurolite–quartz subfacies, the lowest subdivision of the almandine amphibolite facies (Fyfe, *et al.*, 1958, p. 229).

Observations and data on some of the minerals of the above assemblages follow. Plagioclase compositions shown on Figure 27 range from $An_{17}$ to $An_{35}$ and may, to some extent, be influenced by the composition of the host rock. Those in mafic rocks appear to be more calcic. Hornblende is invariably deep green as opposed to the colourless to pale green amphibole of the greenschist facies zone. Four garnets determined by refractive indices and unit-cell edges are almandines with up to about 25 per cent of other garnet components. The data in order of increasing distance from the Labrador Trough are: (1) a=11.579, R.I.=1.805; (2) a=11.550, R.I.=1.813; (3) a=11.539, R.I.=1.807; (4) a=11.553, R.I.=1.800. The localities of these garnets are indicated by number on Figure 27.

### Discussion

Regional metamorphism in the Labrador orogenic belt is of the classical Dalradian type first described by Barrow (1893, 1912). Succeeding zones of increasing metamorphic grade, classified here on the facies principle, are nevertheless characterized by the same critical minerals (chlorite, biotite, almandine, staurolite–kyanite[1]) that define the Scottish zones. Discrete zones of kyanite and sillimanite

[1] In the pelitic rocks.

have not yet been found in the Labrador belt and at the opposite end of the succession a broad pumpellyite zone is present. Miyashiro (1961, pp. 278–283) classified regional metamorphic successions into three standard series—andalusite–sillimanite, kyanite–sillimanite, and jadeite–glaucophane—representing environments of successively higher pressures. The Labrador series, like the Dalradian type, belongs to the kyanite–sillimanite group of intermediate pressures.

The subgreenschist facies zone is of special interest in view of recent work on low temperature metamorphism (Coombs, 1954; Fyfe, *et al.*, 1958; Coombs, *et al.*, 1959; Packham and Crook, 1960; Coombs, 1960; 1961; Seki, 1961). Two new facies have been defined to cover the region between diagenesis and the greenschist facies: the lower grade zeolite facies (Turner in Fyfe, *et al.*, 1958; Coombs, *et al.*, 1959) and the higher grade prehnite–pumpellyite meta-greywacke facies (Coombs, 1960). The zeolite facies is subdivided into a low temperature stage characterized by heulandite and analcite with quartz, and a high temperature stage marked by albite, laumontite, and quartz (Coombs, *et al.*, 1959, p. 91). The prehnite–pumpellyite meta-greywacke facies is defined by Coombs (1960, p. 342) ". . . to include those assemblages produced under physical conditions in which the following are commonly formed: quartz–prehnite–chlorite or quartz–albite–pumpellyite–chlorite, without zeolites and without the characteristic minerals of the glaucophane schist facies, jadeite or lawsonite" Seki (1961, p. 409) recognized two distinct zones between those of diagenesis and the greenschist facies in the Sanbagawa metamorphic terrain of Japan. These are characterized by the minerals chlorite and pumpellyite with respect to increasing temperatures. Data collected from many other metamorphic terrains suggested to Seki the possible existence of three facies in this low temperature field: the zeolite facies; its high pressure equivalent, the chlorite facies; and the pumpellyite–chlorite or pumpellyite–prehnite facies. In the higher pressure field metamorphism would lead to the glaucophane schist facies, and in the lower pressure field the greenschist facies.

The broad zone containing intermittent development of pumpellyite in this map-area is not extensively recrystallized, and at least part of that which is recrystallized might be attributed to deuteric effects. Neither pumpellyite nor prehnite were recognized in the sediments (although the writer hesitates to accept this as evidence of their complete absence), and the primary fabric is little altered. In contrast low temperature assemblages described by Coombs (1954) and Seki (1961) appear to result from extensive recrystallization of the primary rocks. The term facies, therefore, does not seem an appropriate designation for the low temperature zone of mixed metamorphic, deuteric, and primary assemblages of this area. Equilibrium may have been achieved locally with the result that some assemblages are indicative of pressure–temperature conditions that prevailed throughout the zone, but for most of the rocks reaction rates may have been too slow to respond to these modest conditions. Accordingly, in this report the zone has been designated the subgreenschist facies zone rather than as a specific metamorphic facies.

Assemblages of secondary minerals that may be indicative of the conditions of metamorphism are as follows: pumpellyite–epidote–albite, quartz–albite–pum-



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pellyite–chlorite, and rarely pumpellyite–prehnite–chlorite. In each case primary minerals, notably plagioclase, augite, and sphene, are present in addition to those mentioned. Commonly the secondary minerals are accompanied by a fine-grained, colourless mineral or minerals that remain unidentified. These assemblages correspond to the prehnite–pumpellyite meta-greywacke facies of Coombs (1960).

At the high-temperature end of the Labrador metamorphic succession, almandine amphibolite facies rocks pass into granodiorite gneisses. It might be tempting to interpret this as a natural continuation of the series into the granulite facies. However, there is little evidence to support this. The granodiorites contain none of the high temperature minerals, such as pyroxene, garnet, and sillimanite, nor the characteristic textures, such as perthitic and antiperthitic feldspars, common to the granulite facies. The plagioclases are not more (and in places are less) calcic than in the almandine amphibolite rocks. It seems much more reasonable to regard the granodiorite gneisses as basement rocks or arkosic sediments involved in the same deformation, but not of notably higher metamorphic grade than the rocks immediately to the west.

## Thermal Metamorphism

Evidence of thermal metamorphism is confined to the borders of gabbro or ultramafic sills, where it has generally produced inconspicuous zones a few inches to a few feet wide. Shales, greywackes, and red quartzites or arkoses are the rocks most commonly found in contact with gabbros, which probably accounts for the modest metamorphic effects generally encountered. In two places impure carbonate rocks show a much more marked reaction to metamorphism than the sedimentary rocks in general.

The dark shales and greywackes and the red beds are almost invariably bleached for as much as several feet from the contact. In places, rosettes of actinolite crystals from microscopic size to several millimetres in length are present in shales within a few inches of the contact. Few other effects are noteworthy in most cases. In thin section, the rocks, particularly shales, appear to be finely recrystallized and in places new biotite or chlorite may have formed. In one area about 2 miles west of Lac de l'Axe, shales of the Seward Formation have been converted to an adinole for at least 20 feet above the upper contact of a gabbro sill. The rock is cream-white with rare pinkish layers still visible and has a porcellaneous lustre. Commonly it is marked with faint spots one or two millimetres in diameter. Thin sections of the adinole show a finely crystalline mosaic composed of quartz and albite in almost equal parts with subordinate chlorite, actinolite, epidote, muscovite, and sphene thinly distributed in layers or concentrated in clots and clusters of radiating crystals. The rock is similar to adinoles described by Harker (1932, pp. 128–130).

Carbonate rocks at the northeastern tip of Lac Ribero show evidence of a pronounced reaction to thermal metamorphism from nearby sills (p. 23). Coarsely fragmental rocks, composed originally of dolomite fragments in a matrix of dolomitic quartzites, are now extensively recrystallized to mixtures of lime silicate minerals, calcite, and quartz. Some of the fragments are almost entirely lime silicate minerals, others have a rim of lime silicate minerals and an interior of calcite. Masses of magnetite, several inches across, are interspersed in lime silicate rims of some of the large fragments. The only minerals recognized in thin sections of the rock are tremolite–actinolite, diopside, chlorite, quartz and calcite. Tremolite–actinolite is the more abundant of the lime silicate minerals.

Marls of the Seward Formation outcropping on islands just south of the large central peninsula of Lac Cramolet show considerable recrystallization of carbonate, quartz, and chlorite with a variable, but in places substantial, development of actinolite. The contact is not apparent, but gabbro that forms the adjoining peninsula must underlie the marls at a short distance.

# Chapter VI

# STRUCTURAL AND HISTORICAL GEOLOGY

## Structural Geology

Rocks of the Labrador Trough trend northwestward and dip mainly northeast. At the western margin of the Trough the rocks generally rest little disturbed on basement gneisses and dip gently eastward. Within a mile or two of the contact deformation becomes pronounced; the rocks are closely folded with folds overturned to the southwest, and they are cut into numerous slices by northeast-dipping thrust faults. Three belts of distinctive structural character appear in succession eastward from the western margin of the Trough. The western belt of about 15 miles width is characterized by closely spaced folds and thrust faults; the central belt of 15 to 20 miles width, evidently stiffened by the presence of thick gabbro sills, exhibits broad open folds with few thrust faults; and the eastern belt of 10 to 20 miles width comprises steep-dipping, partly drag-folded layers disrupted by one major thrust fault.

East of the Trough structural trends parallel those within the Kaniapiskau Group. The dips of foliation are predominantly eastward but range from moderate adjoining the Trough through a broad flat or undulating zone to steep in the northeast corner of the area.

### Faults

The faults can be classified as strike faults and oblique faults. High-angle cross faults are fairly abundant in the western structural belt (Dufresne, 1952, p. 68; Harrison, 1952, p. 16), but their displacement is too small to show on Map 1209A (*in pocket*).

#### *Strike Faults*

All strike faults in the area are interpreted as northeast-dipping thrust faults for the following reasons. (1) Detailed mapping in the western belt (see for example Dufresne, 1952; Harrison, 1952) has adequately demonstrated that many of the strike faults there are thrust faults. In a few places the faults outcrop and their attitudes can be measured, in others they are known from drill-holes. (2) Many of the strike faults are related to folds in the classical manner of thrust faults; they break through the overturned limb of the fold and appear at surface on the west side of the crest. (3) Both folds and faults are assumed to be manifestations of a continuing orogenic action, and hence can be expected to show an overall consistency.

Thrust faults in the western belt dip moderately to steeply northeast. Dufresne (1952, p. 65) reported that the dips range from 45 to 75 degrees, Harrison (1952, p. 16) observed a fault that dips 40 to 45 degrees, and the writer measured a dip of 45 degrees on a thrust fault that outcrops a few hundred feet west of the north arm of Lac Le Fer. Elsewhere the dips are unknown, but the remarkably straight traces of major faults in the eastern part of the Trough are some indication that these are also fairly high-angle faults. Moreover, steeply east-dipping schistosity in the Murdoch Formation is consistent with a moderate to high-angle dip on the subjacent Walsh Lake–Connolly Lake fault.

Thrust displacements in the western belt appear to be mostly small. In many places folded structures cut by strike faults do not seem to be greatly distorted; the two parts of the dissected structure as they appear on the map still match reasonably well. Stratigraphic separations are generally small. Dufresne (1952, p. 65) reported average stratigraphic separations of 300 to 500 feet, with few more than 1,200 feet. In contrast, strike faults in the central and eastern structural belts produce marked disruptions in structure and stratigraphy, and displacements, although unknown, are assumed to be large. The strike separation of recognizable beds along a few of these faults ranges from 2 to 6 miles.

Three strike faults of major importance are: (1) the Ferrum River fault, which occupies Ferrum River valley and passes along the east side of Wakuach Lake; (2) the Walsh Lake fault (Frarey, 1952, p. 7; Kavanagh, 1954, pp. 141–146); and (3) the Keato Lake fault, which marks the eastern boundary of the Trough. An *en échelon* pair of strike faults, one through Lac Chassin and the other through Northern Lake, are also noteworthy. None of these faults has been seen in outcrop but their presence is readily manifested in other ways.

The Ferrum River fault separates terrains of entirely different lithology and structure. Gabbro sills and arkosic sediments are dominant rocks east of the fault but are sparse or absent west of it. Volcanic rocks and greywackes become increasingly important east of the fault. Probably the Ferrum River fault marks a line along which notable deepening of the geosyncline began. The presence of both coarse clastic deposits and abundant igneous rocks is evidence of an increased rate of geosynclinal depression. The hinge line between parts of the basin with different rates of sinking is an obvious point from which a fault can be expected to spring and the markedly greater resistance to deformation possessed by rocks east of the fault would further favour its development.

The Walsh Lake fault marks the base of the predominantly volcanic Doublet Group. This is probably not fortuitous. The thick volcanic assemblage with few sedimentary interlayers, apart from the Thompson Lake Formation, must represent a mass with little pliancy. Regional movements, therefore, might be expected to be concentrated along its boundary. For much of its course the fault nearly coincides with the lower contact of the Murdoch Formation. Locally, however, it seems to have severed thin slices of Murdoch Formation from the main mass and these now appear on the west side of the fault, just west of Murdoch and Irene Lakes. Conglomerates immediately above the fault at Walsh Lake are presumably at the base of the Murdoch Formation, and are further evidence that the fault nearly coincides with the original lower boundary of the Doublet Group. The fault is marked by a conspicuous lineament and, in its northern part, by an abrupt change in the metamorphism exhibited by rocks on either side. These rocks pass from intermittently altered on the west side to the greenschist facies on the east side.

The Keato Lake fault separates markedly contrasting lithologies and for this reason is taken to be the eastern boundary of the Labrador Trough. Quartz–feldspar–biotite schists abut massive meta-basalts and meta-gabbros. In Wakuach Lake map-area, the fault is represented mainly by a lineament of varying definition, but in the Griffis Lake area, a few miles to the southeast, it is marked by a wide shear zone (Fahrig, 1951). Fahrig recorded dips of 45°NE on foliation of the shear zone. Presumably this reflects the dip of the fault surface.

### *Oblique Faults*

Most of the oblique faults shown on the map strike northwesterly, but a few strike south to southwest. Many of the former are simply parts of strike faults that transect strata in passing from one horizon to another. The most notable example is the north part of the Walsh Lake fault that swings sharply westward through Lac Low. Other northwest-striking oblique faults, such as those just west of the Lac Chassin–Lac Low valley, have no direct relationship with strike faults and since they are unfolded, are probably of later generation. At least one of them shows minor right-hand displacement.

## Folds

The Labrador Trough contains a succession of folds that trend northwestward and are overturned to the southwest. In the western structural belt folding is complex. Folds range in size from small, individual drag-folds a few inches across to highly involved anticlinoriums or synclinoriums several miles across. According to Dufresne (1952, p. 60), individual folds are generally of short strike length and may plunge either northwest or southeast at angles of from 5 to 30 degrees. The central and eastern parts of the Trough are involved in what appears to be a single, or possibly double-crested, asymmetrical anticlinorium, the high point of which probably lies just northwest of the Ferrum River fault. Folds in the area in general culminate in a northeasterly trending zone that passes through Lac Cramolet and from there they plunge at shallow angles to the northwest and southeast.

Folds in the Tait Lake area are of special interest in that they illustrate the effects of superimposed, successive foldings of somewhat different orientation. North-northwest plunging folds are crossed by later east-southeast plunging folds. The most notable effect is shown by the fold east of Tait Lake. The outer layers of the original anticline have evidently been slipped off the core and drawn into a new fold that crosses the crest of the old. The possible successive stages of development are illustrated on Figure 28. The two sets of folds are thought to be penecontemporaneous rather than the result of two discrete episodes of folding. In a belt of rocks undergoing movement, deformation might be expected to originate and spread from a number of isolated centres. Each centre of deformation would probably respond in a slightly different manner depending upon the competency of the rocks, the depth to the basement, and other factors. Folds in the Tait Lake area probably represent the juncture of two such styles of deformation spreading from independent centres. The southeast-plunging set may have originated at a greater distance from the point of intersection than the other, hence arrived later, or it may have been initiated slightly later.

## Relation of Folds and Faults

The writer considers folds and thrust faults of the Labrador Trough as manifestations of the same orogenic action. In most cases the thrust faults are probably extensions of folding. Harrison (1952, p. 15) noted that the characteristic structure in the Burnt Creek strip is ". . . the overturned anticline, with a large part of the overturned limb truncated along a thrust fault . . .". This applies equally well to the western structural belt in the present map-area. Similarly the Ferrum River fault cuts through the forelimb of the major, central anticlinorium. The Walsh Lake fault and other major strike faults on the east side of the anticlinorium are interpreted as back-limb thrust faults (Douglas, 1950, pp. 88–95). The former provides an excellent example. It roughly parallels the fairly steep-dipping beds of the back limb for much of its course, but in the northern part of the area cuts through and repeats some of them.

A continuing interplay between movements on folds and faults during deformation might be expected to produce folded fault planes. Many of the faults appear to be sinuous, but evidence of actual folding can only be postulated in the case of the Walsh Lake fault where it swings westward near Lac Low. Part of the bend no doubt results from an original change in attitude, but it appears to have been exaggerated by continued movement in a manner reminiscent of folded faults of the cordilleran region (Douglas, 1950, p. 85). Pressure on the ends of truncated sills and interlayered strata south of Lac Low is assumed to have caused the flexures and fractures in these rocks farther south. The postulated

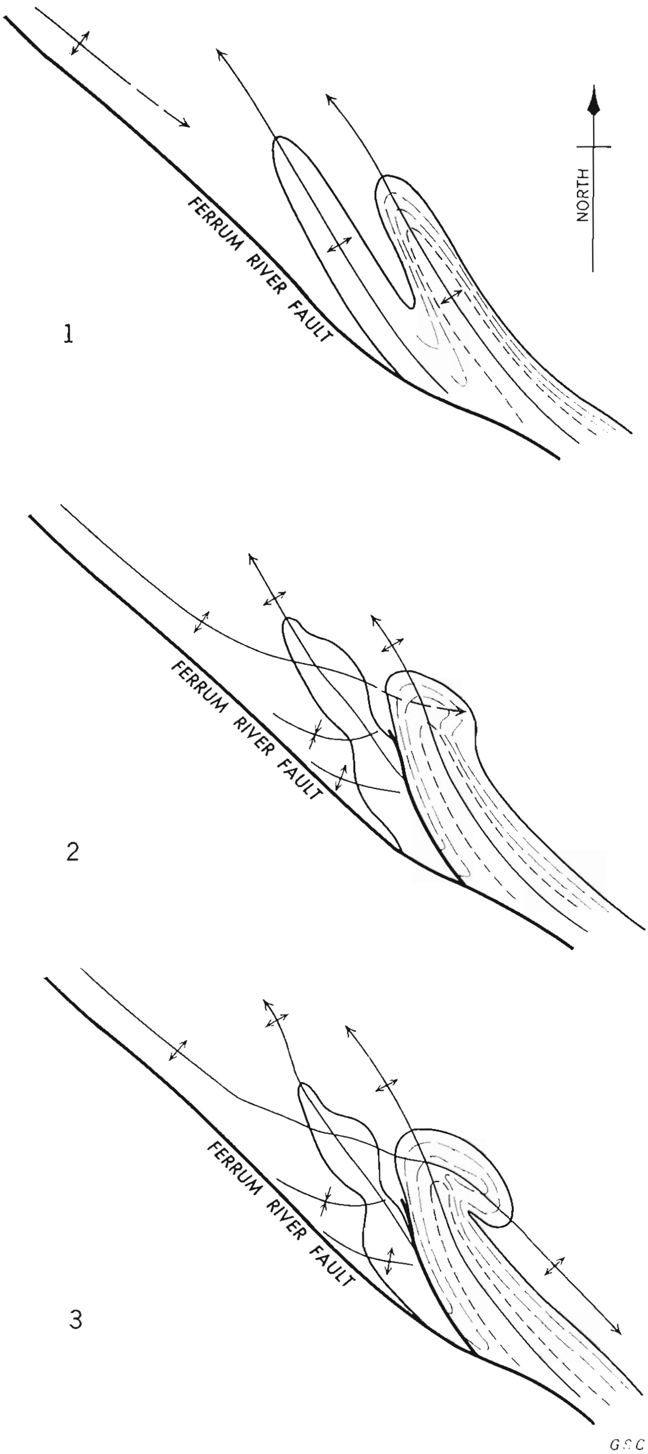


FIGURE 28.
Schematic plan-view diagrams showing successive stages in the development of the folds near Tait Lake.

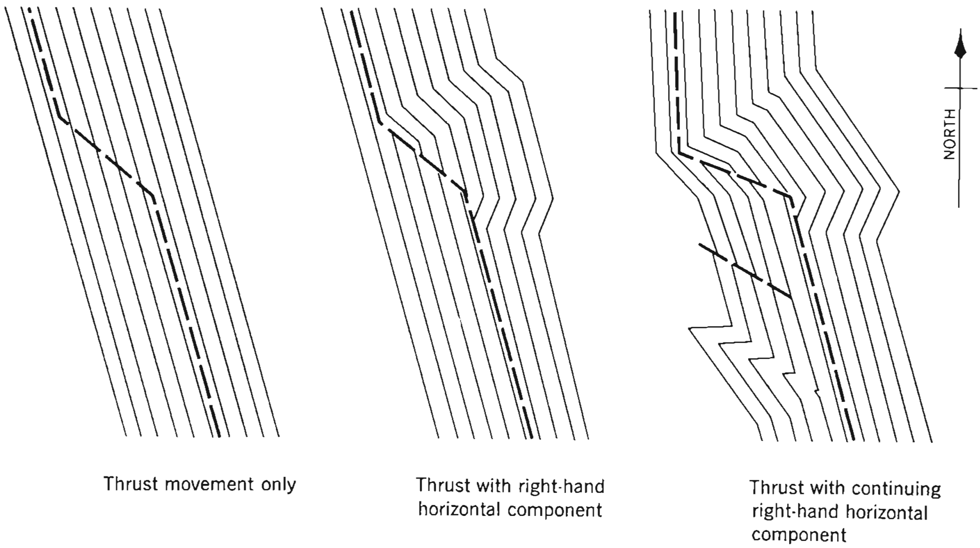


FIGURE 29. Schematic plan-view diagrams showing successive stages in the folding of the Walsh Lake fault.

mechanism is presented diagrammatically on Figure 29. It should be noted that the Walsh Lake fault would need to have a significant component of right-hand, strike-slip movement. That this is so is confirmed by the right-hand offset given the Denault Formation by this fault near the east end of Lac Low.

Forces responsible for the Hudsonian orogeny along the Labrador Trough obviously acted southwestward. However, in Wakuach Lake map-area they appear to have had a significant southeastward component as evidenced by right-hand, strike-slip movement on the Walsh Lake fault, the prevailing southeasterly plunge of folds, and the attitude of major drag-folds involving the Thompson Lake Formation in the southern part of the map-area.

### Cretaceous Movements

Cappings of redeposited iron ore containing Cretaceous fossils are found at a few of the iron-ore deposits in the main ore zone. Some of these "rubble ores" are in basins a few hundred feet across and as much as 200 feet deep. Stubbins, *et al.* (1961, p. 57) postulated that the rubble ores formed in basins marginal to active faults. As the basin floor was lowered by intermittent faulting, it was rapidly filled by the disintegration products of adjoining bedded ores. Thus at least some deformation took place during Cretaceous times.

## Historical Geology

The geological history of the area is essentially the history of an orogenic belt. Broadly it differs little from that of other orogenic belts up to the stage when discordant granites are normally emplaced. In Wakuach Lake map-area granitic stocks and batholiths related to the orogeny are absent. Possibly they occur to the east, but even so it is remarkable that so much of a geosyncline should have escaped the destructive effects of granitic intrusions. The geological history of the region, as presently interpreted, can be divided into geosynclinal and orogenic phases with a further subdivision of the geosynclinal phase into an earlier predominantly sedimentary and a later predominantly igneous stage.

## Geosynclinal Phase

The first sediments known to have been deposited in the Labrador geosyncline were mainly arkoses washed in from granitic terraces to the west, and possibly to the east; these were distributed along the basin by lateral currents. Local volcanism added basaltic material to some of the sediments. With further broadening and deepening of the geosyncline, the arkoses were buried beneath Attikamagen argillites that overlapped them westward and possibly eastward. Volcanism made substantial contributions to the offshore deposits, both in the form of lava flows and as comminuted material, which mixed with the sediments. Fluctuating quantities of quartz and feldspar sands were still being carried into the basin, but most of the externally derived material was in the form of clay or silt. A long period of stability followed, broken only by intermittent outbursts of offshore volcanic activity. During the first phase of the stable period the Denault carbonates were deposited in widespread basins separated by areas of non-deposition. Clastic sedimentation had virtually ceased. Locally the Fleming chert deposits completed this phase. A mild uplift that disrupted the newly deposited chert layers was succeeded by a slow broadening of the geosyncline and a westward creep of the seas. As these spread over the basement rocks, granitic material derived from the margin of the basin was redistributed by currents and wave action over wide areas of the geosyncline. The slow advance and shallow depth of the seas resulted in intensive reworking of the deposits, and remaining sediments were predominantly quartz sands, later to become quartzites of the Wishart Formation. Deposition of an extensive layer of iron-formation, accompanied by volcanism south of the area, completed the stable interval and it was followed by intermittent emergence and erosion. The final event in the early stage of geosynclinal development was the deposition of the Menihek Formation, a thick assemblage of argillaceous sediments accompanied in the offshore regions by much volcanic material.

The igneous stage of geosynclinal formation was confined to the central part of the belt where the thickest deposits had accumulated. It began with a period of explosive volcanism that built a layer of pyroclastic deposits as much as 6,000 feet thick. Near the end of the explosive cycle argillaceous and arenaceous sediments collected in local basins. When volcanic activity ceased these spread over the remaining pyroclastic deposits and accumulated to a depth of several hundred feet. Some of this material was undoubtedly of volcanic origin, but the arenaceous fraction may have been largely derived from external sources. This interval when the Thompson Lake Formation was deposited was followed by a long period of volcanism during which tholeiitic basaltic lavas accumulated in thicknesses in excess of 15,000 feet. Most of the volcanism was effusive and submarine. Accompanying this, and possibly preceding episodes of volcanism as well, scores of basaltic sills were intruded into weaker members of the underlying geosynclinal deposits. These were formed of the same magma as that which reached the surface through volcanoes. Ultramafic sills were intruded late in the last major volcanic episode.

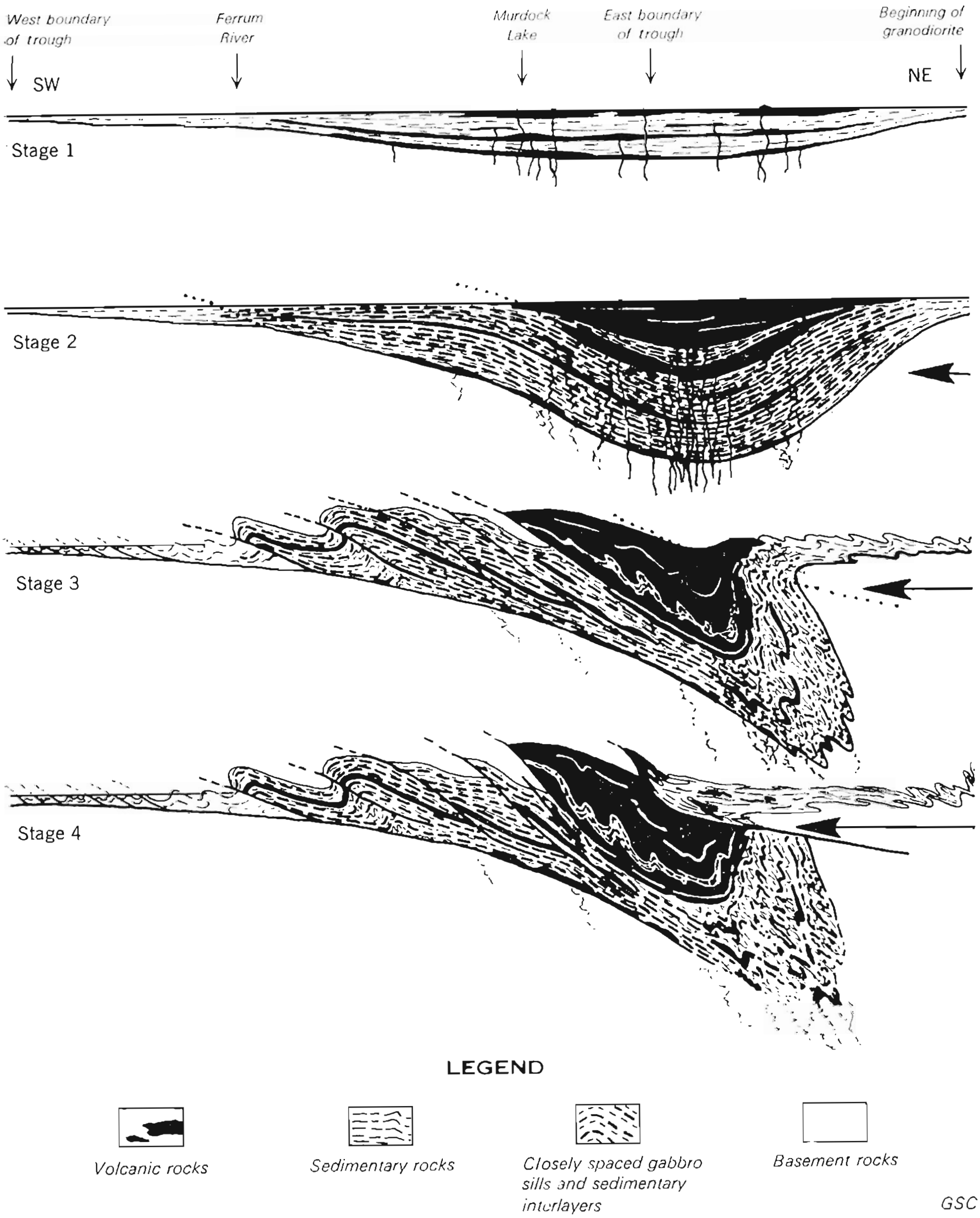


FIGURE 30. Hypothetical cross-sections of the Labrador orogenic belt at successive stages in its development: 1, at the close of Menihek deposition; 2, after eruption of Willbob Lake lavas and emplacement of the Wakuach and Retty sills; 3, after the major episode of folding and faulting; 4, after development of the Keato Lake fault.

## Orogenic Phase

The orogenic phase of the geosynclinal history is largely speculative. The writer's interpretation is given on Figure 30. Stage 1 pictures the geosyncline after deposition of the Menihek Formation. It is considered to have been a fairly flat-sloping basin with wide marginal shelves. Volcanic rocks were liberally scattered through deposits in the interior of the basin, whose maximum thickness was probably between 10,000 and 15,000 feet. Stage 2 shows the geosyncline at the commencement of folding after the addition of some 20,000 feet of volcanic and sedimentary rocks and possibly a further 20,000 feet of sills. The sills were probably distributed far more widely and evenly than their volcanic counterparts, for they are more slowly cooled, retain a higher quantity of volatiles, and are driven by hydrostatic pressures. Thus they are shown extending over a far greater width than the volcanic rocks and terminating only at the marginal shelves. Possibly incipient orogenic forces contributed to their distribution. Stages 3 and 4 represent the development of structures in the present map-area. The most important features are: (1) the development of thrust faults at the base of the central volcanic pile and at the edge of the marginal shelf where the swarm of sills abruptly terminated; (2) the clockwise rotation of the volcanic pile, which squeezed geosynclinal deposits between it and the basement and piled them up in front; and (3) the development of a thrust fault through the east limb of the geosyncline, which carried highly altered sediments from that limb to the centre of the belt. Thus the interpretation attempts to explain respectively: (1) the Ferrum River and Walsh Lake faults; (2) the steep-dipping Doublet Group, which appears to rest on the back of the overturned central anticlinorium; and (3) the Keato Lake fault, which brings into juxtaposition gentle-dipping, highly altered schists and steep-dipping lavas of the Willbob Lake Formation.

# Chapter VII

# ECONOMIC GEOLOGY

## Introduction

Iron is the metal of major economic interest in the Wakuach Lake map-area and the only one yet found in commercially workable deposits. Nevertheless, since the beginning of exploration in this region non-ferrous minerals have attracted considerable attention and at the time of writing (1963) are still vigorously sought after. Iron orebodies are confined to the western part of the Trough, chiefly along the belt known as the main ore zone, whereas non-ferrous mineral deposits are generally restricted to the areas of basic and ultramafic igneous rocks in the eastern Trough region. The locations of the principal mineral deposits and occurrences are shown on Map 1209A (*in pocket*).

Hollinger North Shore Exploration Company, until 1962, held under concession most of the ground contained within Wakuach Lake map-area (Fig. 31). The parts of the map-area in Labrador were held on a similar basis by the associated company, Labrador Mining and Exploration Company Limited. Early exploration work by the concession companies was directed mainly towards the search for base metals, but in 1945, after discovery of several iron orebodies and realization of likely post-war demands for this commodity, the emphasis was shifted to exploration for iron (Gustafson and Moss, 1953, p. 398). This phase continued until 1952 when sufficient iron ore (more than 400,000,000 tons) had been proved to assure the feasibility of production. Exploration for non-ferrous minerals was continued with increased vigour in 1953. In 1955 the concession areas in both Quebec and Labrador were surveyed with airborne magnetic and electromagnetic instruments. In subsequent years the geophysical results were followed up by geological examination, ground geophysical work, and, in places, by drilling. Between 1956 and 1959 a total of 83,500 feet of diamond drilling was completed on the Quebec concession, mainly in Wakuach Lake map-area. Canadian Johns Manville Company, through an arrangement with Hollinger North Shore Exploration Company, explored the Quebec concession area for asbestos during the years 1954 to 1957. At the end of 1962 the concession companies reduced their holdings to the blocks of ground which they are entitled to retain under lease (Fig. 30). In 1963 part of the released ground was acquired by St. Lawrence Columbium and Metals Corporation Limited and was explored in a joint

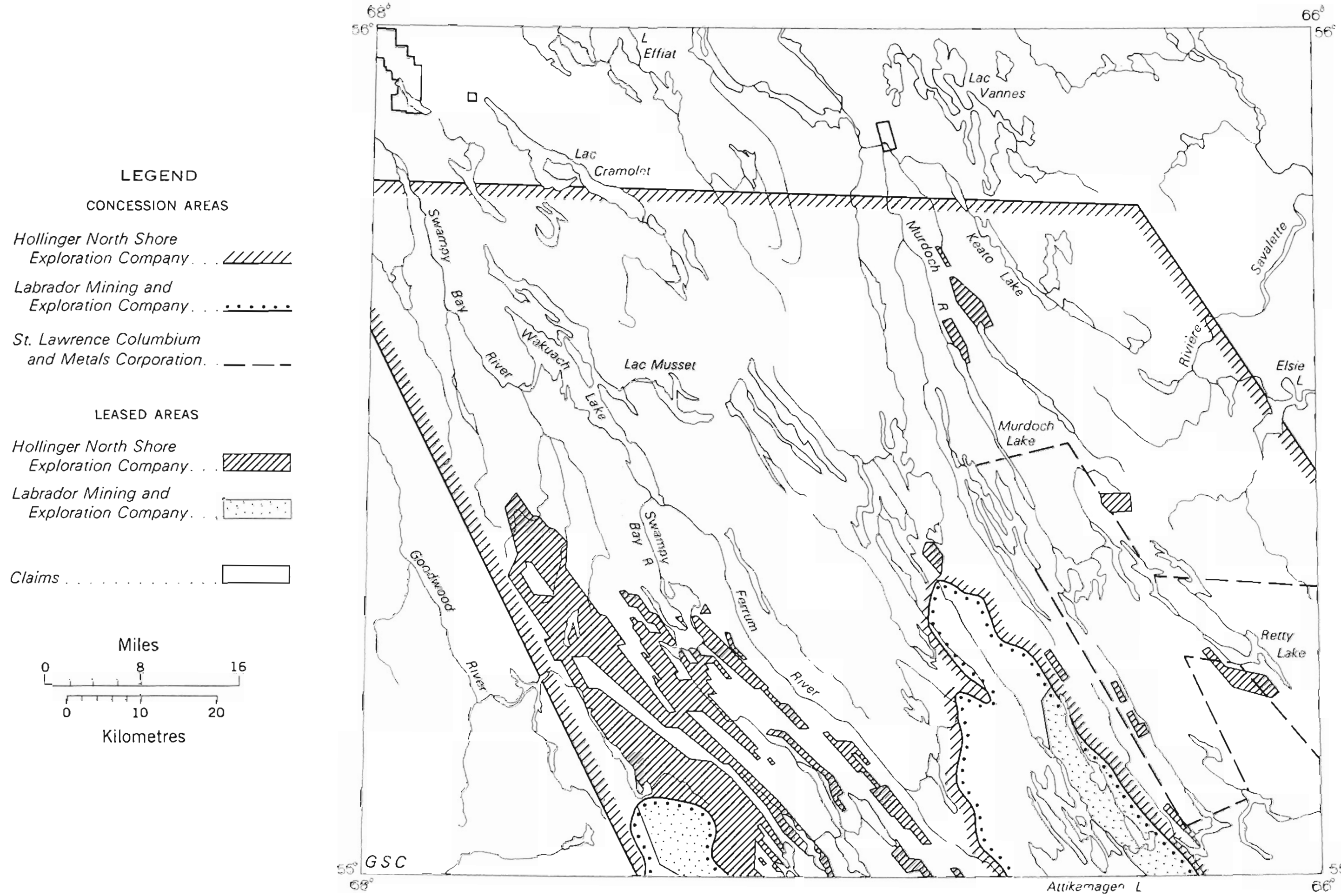


FIGURE 31. Wakuach map-area showing part originally contained in the Hollinger North Shore Exploration Company and Labrador Mining and Exploration Company concession areas and the reduced holdings now retained by these companies under lease. Also shown are the concession areas recently acquired by St. Lawrence Columbium and Metals Corporation and claims in good standing in October, 1963.

venture with Anaconda Company Limited. Immediately north of the map-area, a copper discovery in the late summer of 1961 prompted extensive staking near Lac Romanet and Dunphy Lake, 10 to 15 miles north of Lac Cramolet. To date (1964) nothing of commercial value has been uncovered in the staked area, but the activity focused attention on the northern part of Wakuach Lake map-area and resulted in the discovery of two new lead–zinc showings.

The writer has had little opportunity to examine the mineral deposits of this area. Accordingly the following descriptions, based mainly on the literature and personal communication with company personnel, is necessarily brief. No attempt will be made here to give a balanced summary of the voluminous literature on iron of this region; this has already been well done in some of the references to be cited.

## Iron Deposits

### Deposits of Soft Ores

Nearly all known deposits of soft ores in the Labrador Trough are contained in a belt 4 to 6 miles wide by 60 miles long that parallels the western boundary of the Trough and is situated a few miles east of it. The northern extremity of the belt is represented by the Eclipse deposit (shown on Map 1209A (*in pocket*)). Most of the deposits are concentrated about two centres, one near the town of Schefferville and the other in the southwestern corner of the present map-area. To date mining has been confined to orebodies of the southern group. Within Wakuach Lake map-area thirteen separate orebodies contain an estimated 139,300,000 tons of iron ore. One of these, the Goodwood orebody, is the largest single deposit of soft ores in the Labrador Trough. Little information is available about specific deposits in this area but excellent accounts of soft ores of the Labrador Trough generally are to be found in the following references: Schwellnus (1957); Stubbins, *et al.* (1961); and Gross (*in press*). A summation of the main features of the Labrador soft ores based on the work of these authors is given briefly below.

#### *Mineralogy*

The orebodies are composed essentially of hematite and goethite in varying proportions with, in places, prominent amounts of manganese and hydrous oxides. The hematite occurs in various forms; as earthy red hematite, metallic hematite, specular hematite, and martite. Essentially no magnetite, iron silicates, nor iron carbonates from the original iron-formation survive in the ores. Hollandite and pyrolusite are manganese minerals that have been identified in these orebodies.

#### *Type of Ore*

Ore is defined as mineable material containing more than 50 per cent iron and less than 20 per cent silica on a dry basis (Stubbins, *et al.*, 1961, p. 44). Schwellnus (1957, p. 115) divided the Labrador soft ores into two principal types:

leached ores and enriched ores. Leached ores are formed from the primary rock by simple removal of constituents other than iron, notably silica. They are typically porous, vuggy, and friable; the ore mineral is mainly hematite; and the ore typically reflects the characteristics of the parent rock. Enriched ores are those in which part of the iron content has been introduced, usually in the form of goethite fracture-fillings and vug or pore deposits. Enriched ores are characteristically hard and dense; they are spatially associated with regions of intense deformation, faulting, or brecciation; and they are generally enriched in manganese oxides and hydrous oxides. The grade of enriched ores is high[1] and more uniform than that of leached ores. Correlation between enriched ores and their parent rock is commonly difficult.

Several varieties of leached ores result from initial differences in the parent rocks. Red ores, composed mainly of earthy red hematite, are derived from the Ruth slates; yellow ores predominantly of goethite are from the silicate–carbonate iron-formation; and blue ores mainly of metallic hematite and martite are from cherty metallic iron-formation. A further distinction can be made between Bessemer (low phosphorus, low manganese) and non-Bessemer (moderate phosphorus and manganese) ores. Both are present in Labrador orebodies, but non-Bessemer ores predominate. In Wakuach Lake map-area about 95 per cent of the ore reserves are of non-Bessemer type. Red and yellow ores as well as enriched ores are generally high in manganese and phosphorus and are therefore of non-Bessemer quality. Blue ores are commonly of Bessemer type.

Rubble ores are composed of transported material derived from the erosion of bedded ores during Cretaceous times. They cap a number of the orebodies in the Knob Lake area and range in thicknesses up to several hundred feet. Fossil plant remains are found interbedded with the rubble deposits in many places. Stubbins, *et al.* (1961, pp. 55–56) recognized three types of rubble deposits; largely unconsolidated sand and gravel deposits, talus deposits, and breccia ores. The sand and gravel ores are generally well-bedded and are composed of subangular to well-rounded ore fragments. Talus ores are poorly sorted deposits consisting of angular to subangular blocks of ore and other material. Breccia ores are composed of angular fragments of widely varying sizes generally cemented by iron and manganese oxides. Thus they would fall into the class of enriched ores. The breccia ores, which are composed largely of fragments of the same ore type, are considered by Stubbins, *et al.* (1961, p. 56) to be mainly tectonic breccias produced by Cretaceous deformation.

### *Ore Controls*

Localization of the soft ore seems to depend upon the following four stratigraphic and structural factors.

(1) Character of the parent rock. Apart from the obvious control exercised by the Sokoman and Ruth Formations, generally, relatively minor differences between sub-units of the iron-formation exert a further control on localization of

[1] About 63 per cent iron by dried analysis (Schwellnus, 1957, p. 132).

ore. Thus the 'Lower Red Cherty' and 'Upper Red Cherty' members (Table I, p. 49), which are about 10 per cent richer in iron than other sub-units of the iron-formation, contain the bulk of the leached ores (Schwellnus, 1957, p. 119). The ease of weathering is also a factor in ore control. Rocks in which the matrix contains considerable carbonate and iron silicates are more readily weathered than those in which the matrix consists of chert and iron oxides (Schwellnus, 1957, p. 120), and should be more favourable sites for ore formation. This applies to the silicate–carbonate, the 'Lower Red Cherty', the 'Brown Cherty', and the 'Upper Red Cherty' iron-formation members.

(2) Secondary structures. Most orebodies occur in faulted or unfaulted synclinal troughs (Schwellnus, 1957, p. 147; Stubbins, *et al.,* 1961, p. 45), which Gross (*in press*) believes provided permeable conduits through which surface waters could circulate. Permeability, he suggests, is afforded by fracture zones related to the folding. Another particularly favourable site for ore localization, according to Schwellnus (1957, pp. 70–71, 147–148), is the region of overlap between *en échelon* strike faults with reverse dip-slip movement. Each strike fault must show diminishing displacement towards its termination. As a result, in the region of *en échelon* overlap a strike-slip component will develop, which in adjoining faults has an opposite sense of movement. The enclosed block between the overlapping faults will, therefore, be subjected to a couple and will become severely fractured with resulting increase in permeability. Other factors likely to increase the permeability of the iron-formation and render it more susceptible to leaching are faults of all types, anticlines, and drag-folds.

(3) Absence of an impermeable cover. If penetrating surface waters are responsible for converting iron-formation to ore, then the absence of an impermeable cover constitutes an ore control. Schwellnus (1957, p. 150) noted that the Menihek slates are the only possible barrier in this region and that their absence from the height of land coincides with the main ore zone.

(4) Topography and water-table. The orebodies of the main ore zone are concentrated about the present height of land. This may be a reflection of the topography existing at the time of ore formation (Gross, *in press*) and the height of land would be a region with maximum depth to the water-table. Such a condition would favour the vigorous circulation of surface waters to considerable depths (Schwellnus, 1957, p. 150), and in this way would constitute an ore control.

### *Origin*

Schwellnus (1957), Stubbins, *et al.* (1961), and Gross (*in press*) agree that the soft ores of the Labrador Trough are formed by the leaching action of meteoric waters on the Sokoman and Ruth Formations with or without subsequent enrichment. Evidence for this view is as follows: (1) orebodies and partly leached and enriched zones are closely associated with the present surface; (2) the grade and tonnage of iron ore diminishes with depth; (3) orebodies are invariably related to permeable zones that connect with the surface; (4) oxidization of iron in the orebodies indicates the close relationship that must have existed between the leaching medium and the atmosphere. Schwellnus (1957, p. 150) postulated that ore formation took place initially during the Permian or Permo-Triassic periods when the land was high and arid. During or after a second episode of deformation in the Cretaceous Period, new channels of access were opened to groundwaters and the process of ore formation was renewed. Many of the previously formed leached ores were enriched by hydrous iron and manganese oxides, and rubble ores were formed by normal processes of erosion and sedimentation.

### Deposits of Beneficiating Ores

Iron-formation of the Doublet Group near the north end of Murdoch Lake (pp. 65–69) is possible beneficiating ore. It is in the greenschist facies zone and as a result is somewhat coarser grained than the Sokoman Formation and therefore more amenable to concentration. The grain size of the metallic minerals ranges generally from 0.05 to 0.1 mm with aggregates of crystals up to 1 mm across. The Sokoman Formation on the other hand has a metallic grain size ranging mainly between 0.005 and 0.05 mm. The Doublet Group iron-formation is continuous for at least 2 miles and possibly for 3½ miles. It has a maximum thickness of about 600 feet. The iron minerals of possible commercial interest are hematite and magnetite.

## Non-Ferrous Mineral Deposits

Non-ferrous mineral deposits in the map-area are most simply classified by their mode of occurrence and mineralogy. Generally each mode of occurrence has a characteristic mineralogy.

### Copper–Zinc Deposits in Sedimentary Rocks

Dark argillaceous rocks of the Menihek Formation just west of the Walsh Lake fault, of the Thompson Lake Formation, and of interflow layers in the Willbob Formation are commonly rich in finely crystalline pyrite and pyrrhotite. Chalcopyrite is generally a minor constituent (Kavanagh, 1954, p. 153) and specks of sphalerite are not uncommonly present. Locally chalcopyrite and sphalerite are sufficiently concentrated to form deposits of interest. Frederickson Lake and Vera Lake[1] are examples of such deposits. In addition to the minerals already named, the Frederickson Lake deposit, described by Frarey, carries some gold[2] and minor silver (Frarey, 1967). The origin of these deposits is uncertain. Kavanagh (1954, pp. 157–158) suggested that the pyrite is syngenetic, but that pyrrhotite, chalcopyrite, and presumably sphalerite are the result of hydrothermal emanations from the associated gabbros. Frarey (1967) believes that the close

[1] The names of deposits are those used by Labrador Mining and Exploration Company, Hollinger North Shore Exploration Company, and Iron Ore Company of Canada.

[2] G. M. Hogg, Labrador Mining and Exploration Company; personal communication.

physical association of sulphide-rich shales with mafic and ultramafic igneous rocks, when elsewhere similar shales are barren, demonstrates a genetic link between them. However, it should be noted that such a relationship does not preclude the possibility that the sulphide-rich shales are syngenetic. Elsewhere in this report the writer has attempted to show that the volcanic rocks and gabbros are derived from the same magma. Hydrothermal solutions emitted by submarine volcanoes might be expected to deposit a suite of minerals on the ocean floor similar to that deposited by emanations from the same magma at modest depths.

### Copper Deposits in Gabbro

These deposits consist generally of sulphide-rich lenses in gabbro or glomeroporphyritic gabbro. Pyrrhotite, pyrite, and chalcopyrite in this order of abundance are the major metallic minerals present; pentlandite and magnetite may be minor constituents. Connolly Lake, Dobbin Lake, and in part Walsh Lake are deposits of this type; minor occurrences include the two shown on Map 1209A (*in pocket*), about 1½ miles west of the mid-point of Murdock Lake and about 3 miles northwest of Irene Lake, respectively.

### Copper–Nickel Deposits in Ultramafic Sills

Deposits of this type appear to be localized at or near the base of ultramafic sills. The sulphide minerals are mainly pyrrhotite–pentlandite and chalcopyrite, and the gangue is the serpentinized peridotite host rock. Magnetite is said to be an important associate of the sulphide minerals, but whether or not it is present in greater amounts than in the unmineralized serpentinite is not known to the writer. Deposits of this type are present as follows: Glance Lake, Chance Lake, Ahr Lake, Thompson Lake, Blue Lake, Berry Lake, and Nancy Lake. Most of these occur in relatively thin serpentinite bodies appearing at about the same stratigraphic horizon throughout the belt, viz., at or just below the contact of the Thompson Lake and Willbob Formations. Although these do not seem to form a continuous sepentinite sill at surface, it is possible that they are either joined at depth or were once continuous, but have been separated by tectonic movements. Only the Chance Lake and Glance Lake occurrences have been seen by the writer. At the south end of Chance Lake the sulphide-bearing zone is at least 30 feet thick; it overlies the basal tremolite zone of the sill, which appears to be about 20 feet thick, and grades up into barren serpentinite. The entire thickness of the sill at this locality is less than 200 feet. At Glance Lake sparsely mineralized serpentinite was intermittently observed along a strike length of 8,000 feet. Near the mid-point of Glance Lake the mineralized zone appears to be only a few feet thick; elsewhere its thicknesses are unknown. The sill in this region is generally less than 100 feet thick.

The presence of the sulphide minerals in the basal part of the ultramafic sills with no exogenous gangue minerals is strong evidence that the deposits are magmatic. That many are related to what may have been the same sill suggests that this primary magma was extraordinarily rich in these constituents.

### Copper Deposits in Quartz and Carbonate Veins

These deposits, found mainly in the Murdoch Formation, are generally composed of disseminated pyrite and chalcopyrite in calcite or quartz veins. One deposit of this type, at the northern tip of a meta-gabbro body just east of Walsh Lake, consists of a calcite vein with an exposed length of 30 feet and a maximum width in this distance of 9 feet (Kavanagh, 1954, p. 160). A grab sample yielded an assay of 1.68 per cent copper. Other occurrences have been found on islands in Kalko and Murdock Lakes. So far as known none are of sufficient size to be of economic interest.

### Lead–Zinc Deposits in Sedimentary Rocks

The only two occurrences of this type known in the area are in sedimentary rocks of the Attikamagen Formation between sills of Wakuach Gabbros. One is about 3 miles east of Lac Louis, the other about a mile west of the north tip of Lac Cramolet. Both are recent discoveries and little information on them is yet available.

### Asbestos Deposits

Chrysotile is found in the ultramafic sills at a number of localities, of which Katherine and Retty Lakes are the most notable. Of these the Katherine Lake deposit is said to be the more promising. Crocidolite is found in the iron-formation principally at two localities, Trough Lake and Lac Hameau. Both deposits are reported to be small and of little economic interest.

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# INDEX

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