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GEOLOGY OF THE STRAIT OF BELLE ISLE AREA, NORTHWESTERN INSULAR NEWFOUNDLAND, SOUTHERN LABRADOR, AND ADJACENT QUEBEC

Part 1. PRECAMBRIAN ROCKS OF THE STRAIT OF BELLE ISLE AREA
H.H. BOSTOCK

Part 2. LOWER PALEOZOIC AUTOCHTHONOUS STRATA OF THE STRAIT OF BELLE ISLE AREA
L.M. CUMMING

Part 3. GEOLOGY OF THE HARE BAY ALLOCHTHON
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Preface

The first systematic geological investigations of the Strait of Belle Isle area were initiated by James Richardson of the Geological Survey of Canada in 1860. Investigations have continued for more than a century largely by geologists from the Newfoundland government, universities in Canada, United States and United Kingdom, and mineral exploration companies, all of whom have contributed greatly to the knowledge of the area. This report represents the first large scale compilation of the geology of the Strait of Belle Isle area.

The area is important for two principal reasons. It is located near an ancient continental margin and thus provides an opportunity to study structural and sedimentological processes that were operative during plate tectonic evolution of the margin in the early Paleozoic. It is of economic importance because of extensive limestone and dolomite deposits – the raw materials for the construction industry and potential host rocks for zinc deposits. Information on the detailed geology will be essential if a tunnel is constructed beneath the Strait of Belle Isle to carry hydroelectric power cables from Labrador.

D.J. McLaren
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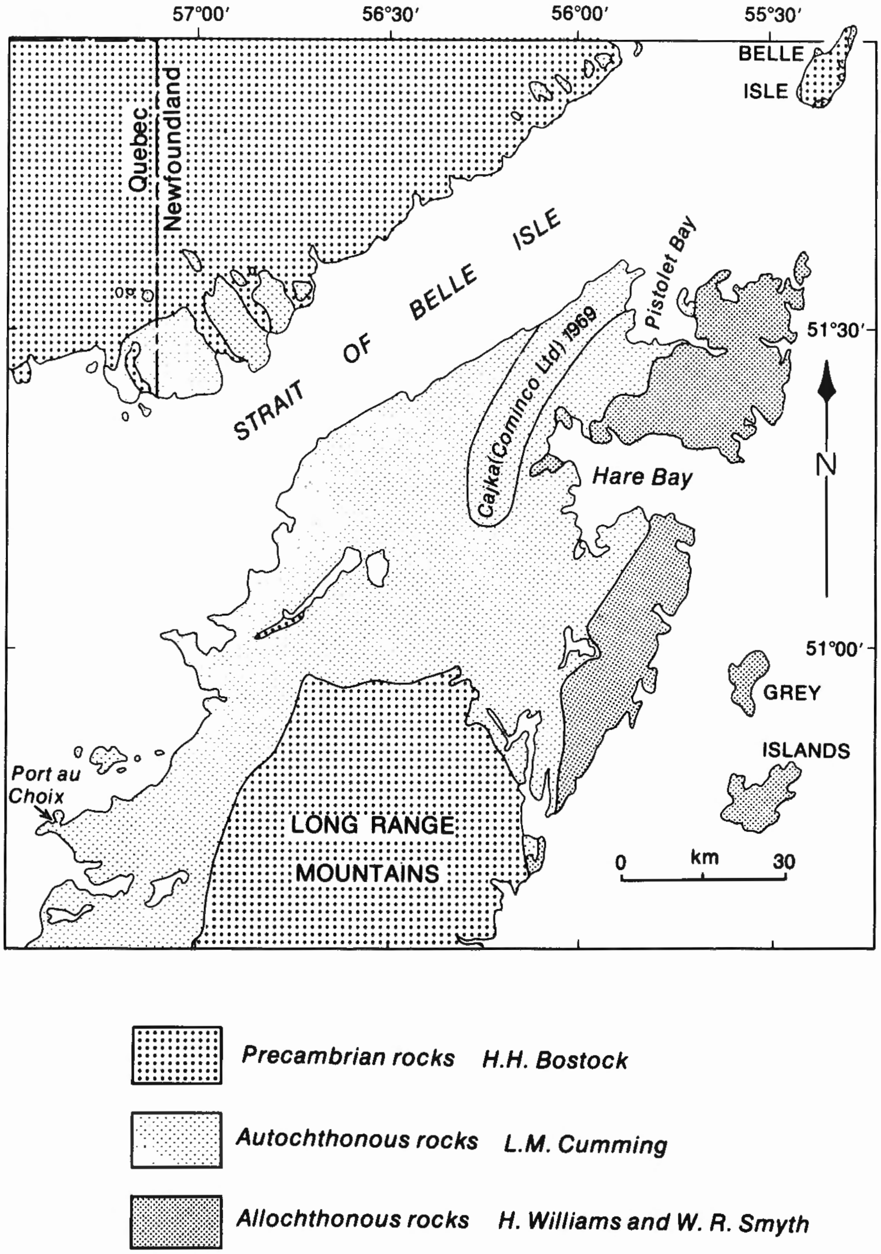

Sketch map showing the three major geological terranes of the Strait of Belle Isle map area and the areas studied by the various authors.

Foreword

The area described in this memoir includes the extreme northern part of the Great Northern Peninsula of Newfoundland and the adjacent islands lying off the coast of Newfoundland including Belle Isle, and a narrow strip along the north shore of the Strait of Belle Isle extending from the extreme southeastern part of Quebec to southeastern Labrador. The area, in general, falls within the confines of 50°30'N – 52°00'N and 55°05'W – 57°30'W and includes parts of map sheets 2 L, M and 12 I, P of the National Topographic System.

Within the Strait of Belle Isle area, the bedrock comprises three principal geological terranes (i) Precambrian crystalline basement rocks of the Grenville Province, largely of Helikian and earlier age, forming an extensive outcrop belt in the Great Northern Peninsula, southern Labrador and Belle Isle; (ii) lower Paleozoic platformal strata with wide distribution in the lowlands bordering the Great Northern Highlands of Newfoundland, as well as the rocks that form the Highlands of St. John, the remnants along the north shore of the Strait of Belle Isle in Quebec and Labrador and fringing the shores of Belle Isle; and (iii) Hadrynian and Paleozoic sedimentary and igneous rocks of the Hare Bay Allochthon.

Systematic geological mapping of the Strait of Belle Isle area, which began in 1969, was completed in 1974. H.H. Bostock was responsible for mapping the Precambrian crystalline rocks; L.M. Cumming for the lower Paleozoic platformal sequences; and Harold Williams and W.R. Smyth for the Hare Bay Allochthon.

At the conclusion of field investigations, the bedrock geology of the above described areas was integrated and compiled on a scale of 1:125 000 (see Map 1495 in pocket). Descriptions of the rocks comprising the three principal geological terranes were prepared independently and consist of the following papers:

1. Precambrian rocks of the Strait of Belle Isle area, by H.H. Bostock.
2. Lower Paleozoic autochthonous strata of the Strait of Belle Isle area, by L.M. Cumming.
3. Geology of the Hare Bay Allochthon, by Harold Williams and W.R. Smyth.

The Strait of Belle Isle area is one of the last in the Canadian Appalachian geological province to be mapped on a reconnaissance scale in modern times. The area is of economic importance because of its extensive limestone and dolomite deposits and potential for zinc deposits. Details of the geology of the region may also be of practical importance in future planning for the construction of a tunnel beneath the Strait of Belle Isle, to carry hydroelectric cables that will deliver power from Labrador to insular Newfoundland.

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GEOLOGY OF THE STRAIT OF BELLE ISLE AREA, NORTHWESTERN INSULAR NEWFOUNDLAND, SOUTHERN LABRADOR, AND ADJACENT QUEBEC

Abstract

The Strait of Belle Isle area comprises three principal geological terranes and each is described separately.

Precambrian terrane. The oldest rocks in the area are widespread leucocratic to melanocratic biotite-quartz-felspar gneiss including some quartz-rich gneiss, pelitic gneiss, amphibolite, and minor calcareous gneiss of Helikian or earlier age. The gneiss was intruded by plutons of the anorthositic suite, commencing with metagabbro throughout the area and continuing with mangerite and hornblende granite north of the Strait of Belle Isle. Small bodies of foliated granite were probably emplaced early in this sequence. Regional folding along northeast-trending axes accompanied by amphibolite to granulite facies metamorphism probably occurred after intrusion of the metagabbro and before emplacement of the hornblende granite. Later, more hydrous megacrystic to massive granitic plutons were emplaced diapirically, mostly within the gneiss of the northern Long Range. Phyllonite, up to 300 m thick, was formed by outward thrusting along the northeast margin of the megacrystic pluton at Canada Bay. Gneiss and megacrystic granite along the east margin of the area were folded along northeast-trending axes and subjected to greenschist facies metamorphism during or soon after emplacement of the megacrystic plutons.

Small bodies of fine grained, massive to foliated, mostly leucocratic rocks of probable cataclastic origin are present near the east coast of the Long Range. The protolith of these bodies is of the same age as the Grenville basement, but cataclasis is of late Grenville or later age or both. Normal faulting followed by reverse faulting is evident on Belle Isle and similar faults are present along the margin of the Long Range Grenville inlier.

A zone of greenschist facies metamorphism of Ordovician age crosses the Precambrian rocks along the east margin of the area and may extend to the southern end of the Grenville inlier.

Lower Paleozoic autochthonous strata. Hadrynian, Cambrian, and Ordovician autochthonous strata of the Strait of Belle Isle area are part of an easterly thickening wedge of strata extending to the southwest and underlying much of the northern Gulf of St. Lawrence. The strata were deposited on a shelf that gave way easterly to a shelf edge and then to a turbiditic facies which formed along the former continental slope and rise.

The oldest strata comprise quartzite, conglomerate, siltstone, and minor basalt of the Bateau Formation on Belle Isle. Tholeiitic diabase dykes of the Long Range dyke swarm of latest Hadrynian age intrude the Bateau Formation and are feeders to tholeiitic basalt flows of the overlying Lighthouse Cove Formation. These flows are interlayered with basal arkosic sandstone, conglomerate and siltstone of the succeeding Bradore Formation. In all, these strata are as much as 700 m thick. A regolith is widely evident on the basement complex beneath the Bradore beds.

The Cambrian to Middle Ordovician is represented by up to 1500 m of shallow water marine carbonates and some shale and quartz sandstone derived from the Precambrian rocks to the northwest as the proto-Atlantic ocean waters gradually transgressed to the northwest. Lower Ordovician carbonate of the St. George Group consists of about 600 m of calcitic dolomite and dolomitic limestone, which exhibits subtle lateral changes from one rock type to the other. Fossils are not abundant and include graptolites, trilobites, and gastropods. Carbonate in the top of the group contains zinc occurrences. Following the Lower Ordovician deposition, the entire continental shelf was uplifted and exposed to subaerial erosion.

After perhaps 5 million years, the shelf again subsided and 500 m of marine limestone of the Table Head Group was then deposited in a Middle Ordovician epicontinental sea upon a karst surface. Water on the shelf then deepened, and detritus for the first time came from the east when the Hare Bay Allochthon was being moved westward on to the shelf. The resulting black shale and minor greywacke of the Goose Tickle Formation, as much as 500 m thick, were deposited gradationally upon the carbonate.

Hare Bay Allochthon. The Allochthon consists of a variety of sedimentary and volcanic and plutonic rocks including greywacke, polymictic conglomerate, quartz-pebble conglomerate, siltstone, shale, mafic pillow lava, peridotite, harzburgite, dunite, gabbro, and diorite.

These rocks make up six contrasting rock assemblages which are separated by thrust faults. The lower structural slices consist of sedimentary rocks and the highest structural slice consists of an ophiolite suite. Mélange zones commonly separate the thrust slices. One of these zones contains large exotic blocks.

The commonest blocks are serpentinized peridotite, mafic volcanic rocks, amphibolite, foliated gabbro, greywacke, diorite, hornblendite and hornblende-biotite schist. These are contained in a matrix of black and green shale. Most rock types of the blocks may be matched directly to rocks in nearby structural slices.

The rocks of the Hare Bay Allochthon originated in the east, in the region near what is now Notre Dame Bay, and record a part of the development and destruction of the ancient continental margin of eastern North America.

Résumé

La région du détroit de Belle-Îsle comprend trois types principaux de terrains géologiques que nous décrirons séparément.

Terrains précambriens. Les roches les plus anciennes de la région sont formées de gneiss largement répandue, leucocrates à mélanocrates, à biotite, quartz et feldspath, comprenant de petites quantitées de gneiss quarzifères, de gneiss pélitiques, d'amphibolites et de gneiss calcaires, de l'Hélikien ou d'âge plus ancien. Les gneiss sont traversés par des plutons constitués par une suite anorthositique, formée à sa base par un métagabbro présent dans toute la région, et à son sommet par une mangérite et un granite à hornblende qu'on retrouve dans le nord du détroit de Belle-Îsle. Il existe des petits massifs de granite foliacé qui se sont probablement mis en place au début de cette série. Un plissement régional, suivant des axes orientés nord-est accompagné d'un métamorphisme du faciès amphibolite à celui granulite, s'est produit après l'intrusion du métagabbro mais avant la mise en place du granite à hornblende. Plus tard, des plutons granitiques massifs, ou macrocristallins et plus hydratés, ont formé une intrusion diapirique, surtout dans les gneiss du nord des Long Range. Des phyllonites qui atteignent parfois 300 m d'épaisseur doivent leur existence à une poussée qui s'est exercée vers l'extérieur le long de la bordure du pluton macrocristallin de Canada Bay. Les gneiss et les granites macrocristallins de la bordure orientale de la région, ont été plissés suivant des axes orientés nord-est et soumis à un métamorphisme du faciès schistes verts, pendant ou juste après la mise en place des plutons macrocristallins.

On trouve, près de la côte orientale des Long Range, des petits corps microcristallins, massifs ou feuillités, formés de roches surtout leucocrates, ayant probablement une origine cataclastique. La roche mère d'où proviennent ces corps est du même âge que le soubassement de Grenville, mais les processus cataclastiques datent au plus de la fin du Grenvillien. À Belle-Îsle, on voit que la formation de failles a été suivies de celle de failles inverses et il existe des failles similaires le long de la bordure de l'enclave grenvillienne des Long Range.

Il existe une zone métamorphique du faciès schistes verts, d'âge ordovicien, qui traverse les roches précambriennes le long de la bordure orientale de la région et qui s'étend vers l'extrémité sud de l'enclave grenvillienne.

Couches autochthones du Paléozoique inférieur. Les couches autochthones de l'Hadrynien, du Cambrien et de l'Ordovicien, de la région du détroit de Belle-Îsle font partie d'un ensemble de couches qui s'épaissit vers l'est et s'étend vers le sud-ouest où il recouvre une grande partie du nord du golfe Saint-Laurent. Les sédiments se sont déposés sur un plateau continental qui a fait place vers l'est à une flexure continentale puis à des faciès à turbidites qui se sont formés le long de l'ancien talus continental et de l'ancien glacis continental.

Les roches les plus anciennes sont des quartzites, des conglomérats, des siltstones et quelque basaltes de la formation de Bateau sur Belle-Îsle. Les dykes de diabase tholéitique du système de dykes des Long Range, qui datent de la fin de l'Hyacrynien, traversent la formation de Bateau et ont alimenté les coulées de basalte tholéitique de la formation sus-jacente de Lighthouse Cove. Les coulées sont interstratifiées avec le grès arkosiques, les conglomérats et les siltstones de la base de la formation sus-jacente de Bradore. L'épaisseur totale de ces couches ne dépasse pas 700 m. L'existence d'un régolithe étendu sur le socle et au-dessous de la formation de Bradore est évidente.

La période du Cambrien à l'Ordovicien moyen est représentée par au plus 1 500 m de carbonates d'eau peu profonde et d'un peu de schistes argileux et de grès quartzifère formés par érosion de roches précambriennes situées au nord-ouest, au fur et à mesure de la transgression de l'océan proto-Atlantique vers le nord-ouest. Les roches carbonatées de l'Ordovicien inférieur du groupe de St-George ont une épaisseur moyenne de 600 m et sont formées de dolomite calcique et de calcaire dolomitique qui passent très progressivement d'un type de roche à l'autre. Les fossiles sont peu nombreux et comprennent des graptolites, des trilobites et des gastéropodes. Au sommet du groupe, les roches carbonatées contiennent des venues de zinc. Après la sédimentation de l'Ordovicien inférieur, l'ensemble du plateau continental s'est soulevé et a été exposé à l'érosion subaérienne.

Après peut-être 5 millions d'années, le plateau continental s'est de nouveau affaissé, et 500 m de calcaire marin du groupe de Table Head se sont alors déposés sur une surface karstique, dans une mer épicontinentale de l'Ordovicien moyen. L'eau est devenue alors plus profonde au-dessus du plateau continental, et pour la première fois le matériel détritique est venu de l'est alors que l'allochtone de Har Bay glissait vers l'ouest sur le plateau continental. Les schistes noirs accompagnés d'un peu de grauwacke, de la formation de Goose Tickle, qui atteignent parfois 500 m d'épaisseur, se sont alors déposés sur les roches carbonatées, et présentent un granoclassement.

Allochtone de Hare Bay. Il consiste en un ensemble de roches sédimentaires, volcaniques et plutoniques en particulier des grauwackes, des conglomérats polygéniques, des conglomérats à galets de quartz, des siltstones, schistes argileux, laves mafiques en coussinets, péridotites, harzburgites, dunites, gabbros, et diorites.

Ces roches forment jusqu'à six assemblages bien individualisés qui sont séparés par des failles chevauchantes. Les nappes structurales inférieures consistent en roches sédimentaires alors que la tranche supérieure consiste en une série ophiolitique. Généralement les nappes de charriage sont séparées par des zones de mélange. Une de ces zones contient de gros blocs exotiques.

Les blocs qu'on rencontre le plus souvent sont formés de péridotite serpentinisée, de roches volcaniques mafiques, d'amphibolite, de gabbro foliacé, de grauwacke, de diorite, de hornblendite et de schiste à biotite et hornblende. La matrice qui contient ce matériel est formée de schiste noir et vert. La plupart des types de roches de ces blocs correspondent aux roches de nappes de charriage des environs.

Les roches de l'allochtone de Hare Bay se sont formées à l'est, dans la région proche du secteur actuel de la baie Notre-Dame, et sont les témoins d'une partie du développement et de la destruction de l'ancienne marge continentale de l'est de l'Amérique du Nord.

Frontispiece: *Barge Bay, a small fishing community on the north shore of the Strait of Belle Isle, showing racks for drying fish in the foreground and glacially rounded hills of Precambrian gneiss in the background. (GSC 160120).*

PART 1

PRECAMBRIAN ROCKS OF THE STRAIT OF BELLE ISLE AREA

H.H. Bostock

INTRODUCTION

Location and accessibility

The area surveyed in the course of Operation Strait of Belle Isle is in northern Newfoundland and in southern Labrador and Quebec between 50 00'N - 52 01'N and 55 16' - 57 25'W.

The principal settlements are Englee, Roddickton, St. Anthony, Hare Bay, Blanc Sablon, and Port Saunders. These and other smaller ports are visited regularly or on request by coastal steamer. Some of these may also be reached by all-weather gravel roads from the Trans-Canada highway at Deer Lake. Communities along the north shore of the Strait of Belle Isle west of Red Bay are accessible by gravel road and are connected to insular Newfoundland by ferry service between Blanc Sablon and Flowers Cove.

Most areas of the interior in the northern Long Range Mountains are accessible by float-equipped aircraft from Deer Lake, but the interior north of the Strait of Belle Isle is more deeply dissected, and lakes suitable for aircraft landing are locally absent. The only practical access to this area for more detailed investigations is by helicopter. River travel in the interior highlands is inhibited by many rapids and waterfalls.

Climatic conditions within the map area are variable; the best weather usually occurs in July but always with the threat of fog or rain. High winds are common and typically blow from the west or northwest during fair weather, and from the northeast to southeast during foul weather. Ice in the upland lakes of the Long Range Mountains commonly breaks up about the first of June but may persist as much as two weeks later.

Excellent outcrop is present along the coasts, particularly along the north coast of the Strait of Belle Isle and along the coast of the Great Northern Peninsula. Landings can be made from small boats at many places when the wind is calm or blowing offshore, and the ocean swell is down. Daily tide range is generally less than 2 m except in the southwestern part of the area where it is slightly greater.

Previous geological investigations

Many early travellers from 1600 - 1860 (e.g. Lt. Baddely, 1829 in Packard, 1891) through the Strait of Belle Isle likely noted the impressive basalt remnants at Henley Harbour (Fig. 1), but the first systematic coastal geological survey in the area was by James Richardson in 1860 and 1862 (unpublished field notes, Public Archives, Ottawa). Limited coastal surveys continued until Foley (1937) and Fritts (1953) made geological traverses on foot across the interior of the Long Range Mountains from coast to coast. In the middle 1950s Brinco carried out a regional, aircraft-supported, geological reconnaissance of which the present map area forms a part. Specific areas within the Long Range highlands were examined by New Jersey Zinc Exploration Co. (Canada) Ltd. in 1967.

James Richardson (unpublished field notes, 1860) examined Precambrian gneiss along the north shore of the St. Lawrence River into Bradore Bay and as far east as Black Bay. He recorded structural observations, and measured sections of the Bradore and Forteau formations in the vicinity of Blanc Sablon. The results of his work were included in the volume entitled "Geology of Canada" (Logan, 1863).

O.M. Lieber (1860), while attached to an astronomical expedition of the United States Coast Guard in 1860, observed a dioritic rock at the entrance to Chateau Bay.

Packard (1891) joined an expedition organized by William Bradford of New York in 1864 for the purpose of sketching and painting icebergs and arctic scenery. He observed "Laurentian" gneiss along the north shore of the Strait of Belle Isle, and in particular described the basalt and underlying gneiss at Henley Harbour. His publication of 1891 provides an extensive bibliography of natural history and early exploration of the Labrador coast.

Alexander Murray (Chapter 2 in Murray and Howley, 1881) made the first geological investigation of the eastern coast of the Great Northern Peninsula of Newfoundland. He recognized rocks of the "Laurentian System" extending from Little Coney Arm in White Bay to Canada Bay. He recorded the presence of northeast-trending greenstone dykes intrusive into the gneiss and remarked on the absence of crystalline limestone within the Laurentian System.

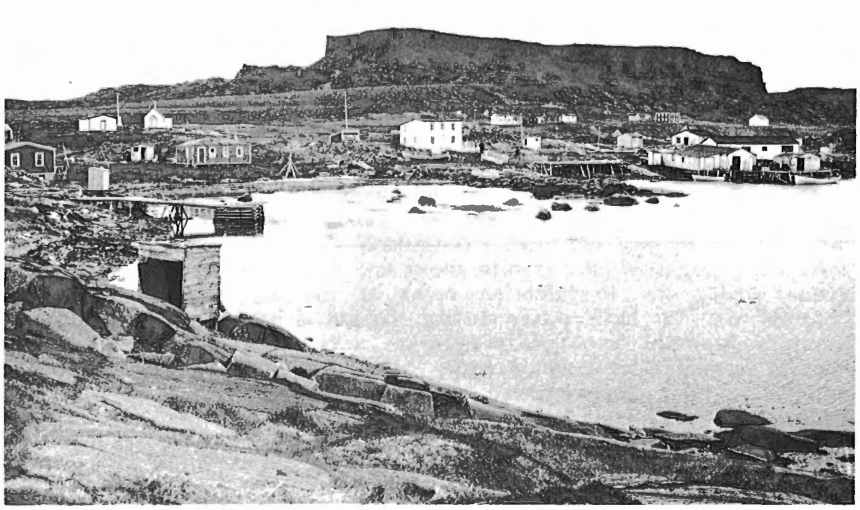


Figure 1. The Devil's Dining Table, basalt flow of the Lighthouse Cove Formation at Henley Harbour. (GSC 160126).

Robert Bell (1885), as a member of an expedition to northern Labrador and Hudson Bay in 1884, stopped briefly at Blanc Sablon and Henley Harbour. His observations were substantially the same as those of Richardson and Packard.

A.R.C. Selwyn (1890) took passage on the lighthouse steamer "Napoleon", "which accorded opportunities for examination only at widely separated points." Landings were made at Greenly Island, Chateau Bay, and Belle Isle. He observed that Belle Isle, in large part consisted of various crystalline and subcrystalline rocks like those of the Huronian mineral-bearing belts of the country north and west of lakes Huron and Superior, and was not entirely Laurentian gneiss as previously supposed from observations by Packard (1891), and others on the Labrador coast.

J.P. Howley (Chapter 22 in Murray and Howley, 1918) made a reconnaissance study in the White Bay district in 1902 during which he visited several points along the eastern coast of the Great Northern Peninsula. His observations of the Precambrian essentially confirmed those of Alexander Murray.

Schuchert and Dunbar (1934), in the course of their examinations in 1910, 1918, 1920, and 1933 of Paleozoic rocks along the coast of western Newfoundland, along the Strait of Belle Isle, and as far southeast as Canada Bay, observed the contact between the Ordovician Green Point Formation and Precambrian gneiss of the Long Range Highlands at Western Brook Pond. The former strata were found to be intensely deformed for half a kilometre or more west of the contact, and these authors suggested that the gneisses were thrust westward over the Paleozoic rocks.

Foley (1937) completed a stadia controlled geological reconnaissance of Precambrian rocks from Pikes Feeder Pond near the southwestern margin of the map area to Great Harbour Deep in 1936. He divided these rocks into four units: pre-Laurentian schist, Laurentian gneiss, post-Laurentian acidic intrusions, and basic dykes. Foley believed that the pre-Laurentian schist, which includes pelitic schist and minor quartz-rich gneiss bodies, described in this memoir, were probably remnants of an early sedimentary terrane engulfed by Laurentian gneiss and post-Laurentian intrusives. He recongnized the occurrence of unusual syenitic rocks (diorite unit Hdi of this memoir) some 3 km southeast of Pikes Feeder Pond; however he included, as part of the same body, megacrystic and related granitic rocks which occur more widely throughout the northern Long Range Mountains, but were of unusually restricted occurrence along his traverse. Foley recorded the presence of diabase dykes* up to 105 m wide in the eastern part of the Long Range and observed that "metamorphism is noticeably more pronounced in dykes close to the coast". He also noted the presence of glauconite in the Bradore Formation overlying the Precambrian gneiss. Foley suggested that the structure of the Precambrian basement complex was geanticlinal with the axis trending a little west of north and located along the height of land. Two northeast-trending thrust faults with eastern side upthrown were interpreted at Great Harbour Deep.

Betz (1939) spent the summers of 1936 and 1937 mapping the geology and marble deposits in the Canada Bay area. He recognized pink granite gneiss and dark hornblende gneiss among the Precambrian rocks of the area, and observed that the basic dykes cutting the gneiss have been intensely altered. Some of these dykes were reported to contain fragments of country rock. Basic dykes "of more definite form" were observed to intrude allochthonous Ordovician sediments at various places and to have undergone metamorphism similar to that of their host rocks. Betz suggested that east-dipping thrust faults, present in the rocks north of Canada Bay, converge in the region south of Wild Cove where some 1200 m of beds were thought to have been removed as a result of faulting and erosion. In a later paper, Betz (1943) suggested that a zone of late Paleozoic faulting extends northeasterly through western Newfoundland from St. George's Bay area to White Bay, and thence more northerly off the coast of the Great Northern Peninsula at least as far as Conche and Cape Rouge peninsulas.

Twenhofel and MacClintock (1940a) recognized three southeast-sloping erosion surfaces in Newfoundland. The highest, the Long Range Peneplane (elevation of 610 - 790 m in the Long Range Mountains), is represented by the summits of monadnocks, and was believed to be Tertiary or possibly Cretaceous in age. Other surfaces recognized in the west are at elevations of 400 - 520 m and 300 - 330 m.

Krank (1939) made a reconnaissance of the Labrador coast starting at Blanc Sablon in 1937. He described the "strong linear schistosity, which over great areas has an almost horizontal position" that is most conspicuous in the country between Wreck Bay and Chateau Bay. He observed that these gneisses are completely crystalline without evidence of cataclasis despite severe deformation. With regard to the Bradore Formation, he noted that it does not occur as an almost continuous series of outliers along the coast between Bradore Bay and Chateau Bay as had been supposed (Logan, 1863) but that the western exposures north of the Strait are limited to an area between Bradore Bay and West St. Modeste.

Christie (1951) conducted a reconnaissance of the southern Labrador Coast in the summer of 1950. His observations within the present map area confirmed those by Krank.

Douglas (1953) spent the field seasons of 1946 and 1947 examining the coast of Labrador with emphasis on the examination of mineral occurrences. His party landed at Blanc Sablon, Forteau, and Red Bay to investigate reports of coal but none was found. Iron reported at Carrol Cove (between Pinware and Red Bay) proved to consist of minor concentrations of magnetite and ilmenite associated with aplite and pegmatite dykes in a schist-gneiss complex. Landings at Chateau Bay and Henley Harbour confirmed earlier descriptions of the local geology.

Fritts (1953) made three reconnaissance traverses across the Long Range: from St. Pauls Inlet to Jackson's Arm, from Canada Bay to Bartletts Harbour, and from Eastern Blue Pond to Little Harbour Deep. In this work three principal lithologies were recognized. These comprised the Highlands Formation of acid to basic gneiss and schist including quartz-rich gneiss and some conglomerate; porphyritic granite; and massive Cloud River granite. The porphyritic granite was recognized to be intrusive into the gneiss formation which, on the basis of quartz-rich gneiss and metaconglomerate components, Fritts considered to be largely of sedimentary origin. Because of its fresh unaltered appearance the Cloud River granite was tentatively considered to be younger than the porphyritic granite. Basic dykes intrusive into the gneiss and porphyritic granite were found to be foliated near the east coast but more massive elsewhere. Fritts noted the occurrence of tourmaline in one pegmatite and purple fluorite and an opaque radioactive mineral in another. Comparison of rocks in the northern part of the Long Range Mountains with those encountered farther south suggested that the latter have been more highly altered.

*The term diabase is used in this report to refer to the basic dykes of the Long Range swarm that are of basaltic composition, and include rocks of ophitic to gabbroic textures and range from fine to medium grained.

Oxley (1953), Nelson (1955), and Woodard (1957) mapped the western margin of the Long Range Highlands and recorded contact relationships between the gneiss and the overlying lower Paleozoic rocks and faults along which rocks of both ages had been involved. Early (post-Middle Ordovician) thrusting of the gneiss over the Paleozoic rocks in the southern part of the area was less pronounced toward the north in concert with a northward reduction of folding in the Paleozoic rocks. Later, high-angle faults persisted northward and in some cases brought Precambrian rocks to the surface west of the edge of younger strata (as at Eastern Blue Pond). Fluorite was observed as coatings on the walls of some of these faults (Woodard, 1957). Along the central and northern parts of the western highland margin, the Bradore Formation overlies highly weathered Precambrian gneiss, the composition of which resembled that of the overlying Bradore (Nelson, 1955). Woodard (1957) suggested that the lower glauconite-bearing beds of the Bradore Formation, which he named the Two Mile Pond member, constitute a restricted basal unit of the formation not found elsewhere.

W.D. Harrison and F. Johnson (unpublished report, British Newfoundland Corporation, 1955) carried out a reconnaissance survey of the Precambrian highlands for Brinco in 1954. In addition to confirming observations of previous workers they suggested that biotite schist lenses mapped by Foley (1937) west of Pikes Feeder Pond may be related to similar lenticular schist bodies found as far south as Parsons Pond. The eastern porphyritic granite is in sharp contact with the Precambrian gneiss and was interpreted as a sill, presumably because of the shallow dip of the contact gneiss in many places. Basic dykes of three ages were observed to be intrusive into the gneiss west of Chimney Bay. Rocks along the western scarp margin of the highlands were found to be widely chloritized with local shearing and minor brecciation. Red alteration and minor fluorite in tiny fracture fillings were observed.

Hale (1961) conducted an aircraft-supported reconnaissance around St. Augustin and as far east as Forteau. His principal contribution was the mapping of previously little known anorthosite bodies west of the present map area.

Clifford and Baird (1962) produced the first comprehensive map of the Precambrian highlands of the Great Northern Peninsula based on data assembled from previous workers and their own reconnaissance. Their map delineates areas of granitic rocks within the older gneiss and also traces the contacts with overlying supracrustal rocks. The extent of the basic dyke swarm intrusive into the eastern part of the highlands and the related flows about Canada Bay are also shown. Several distinct lithologies in the Precambrian gneisses comprised psammitic schist and gneiss, pelitic and semipelitic schist and gneiss, hornblendic schist and gneiss, quartzite, and conglomerate. The hornblendic gneiss was recognized to be partly of intrusive and partly of sedimentary origin. They concluded that the regional metamorphic grade of the gneiss corresponded to the staurolite-quartz subfacies of the almandine-amphibolite facies, although sillimanite had been found in one sample of pelitic schist. Three types of granitic intrusions were found (i) porphyritic granite (K-Ar biotite age of Leg Pond pluton 960 Ma); (ii) granite and granite gneiss with indefinite boundaries; and (iii) the Cloud River Granite of Fritts (1953) (K-Ar age 945 Ma).

Eade (1962) compiled data of previous workers in the Battle Harbour-Cartwright area adjoining the present map area to the north, and carried out 8-mile reconnaissance of most of the remaining unmapped parts of the quadrangle during 1961. His map shows foliated granitoid gneiss grading into foliated and massive granite to granodiorite adjacent to the northern boundary of the present map area. Epidote - amphibolite facies metamorphism was thought by Eade to be normal for most rocks.

Clifford (1965) discussed the age of the Cloud Mountain basalt and described thin discontinous lenses of clastic sediments up to 6 m thick beneath the basalt. He suggested that the clastics underlying the basalts are most likely of Early Cambrian age "equivalent to the Cloud Mountains Formation of Betz" (1939) (Bradore Formation, unit l€B, of this memoir), but did not exclude the possibility that they are significantly older.

Stevens (1967) visited Belle Isle in 1967 and found the island to be chiefly underlain by Grenville-like acidic gneiss cut by abundant Lower Cambrian or Eocambrian basic dykes and sills. Small exposures of basic volcanic flows intimately associated with the basic intrusions and conformably overlain by arenite and shale of the Lower Cambrian Bradore and Forteau formations were reported. A northeast-trending cleavage cutting the basic intrusions and sediments was observed, as well as high-angle faults with similar trends.

Pontin (1967) reported on the examination of six areas in the Precambrian highlands during 1967 for New Jersey Zinc Co. (Canada) Ltd. His observations confirmed those of previous workers, those of particular interest being the presence of quartzite and some conglomerate beds within the gneiss, and the greater degree of alteration of the gneiss in the southern part of the highlands.

Harold Williams surveyed the coast of Belle Isle in 1968 (Williams and Stevens, 1969). His work introduced a new unit, the Bateau Formation, consisting of plutonic boulder conglomerate, quartzite, arkosic sandstone, siltstone, slate, and minor volcanic rocks exposed along the northeast coast of the island. The Bateau Formation is intruded by diabase dykes and was therefore thought to be older than the Bradore sandstone. The axis of a major anticlinal structure produced by Paleozoic deformation was interpreted to strike north-northeasterly through the centre of the island.

I.R. Pringle (Pringle et al., 1971) collected specimens of megacrystic granite and of diabase for radiometric dating during a British Schools Exploring Society excursion in the Hawke Bay area. The age of intrusion of the Long Range megacrystic granite plutons was determined by these authors to be 1130 ± 90 Ma (Rb-Sr, λ^{87}Rb = 1.39×10^{-11} yr^{-1}) based on an isochron calculated from samples from the Hawke Bay and Portland Creek areas. An average of two K-Ar biotite ages for grantite from the Portland Creek area, 840 ± 20 Ma, was interpreted to represent a time of uplift subsequent to Grenvillian metamorphism. Six K-Ar whole-rock age determinations were completed on five samples of a diabase dyke from the western margin of the dyke swarm. The average age, 805 ± 35 Ma, was interpreted as the minimum age for the time of their emplacement. Based on correlation of the Long Range dyke swarm with the Lighthouse Cove Formation by Williams and Stevens (1969), Pringle et al. (1971) suggested that a disconformity exists between the Lighthouse Cove and Bradore formations.

Strong and Williams (1972) studied the basic dykes and flows of northwestern Newfoundland in which samples from within the present area were included. They concluded that both flows and dykes are tholeiitic, and that the samples from Newfoundland resemble basic rocks of similar geologic setting eleswhere along the western part of the Appalachian Province. Similarities with more recent basalts from known tectonic environments were considered to suggest that the Appalachian basalts mark the initial stages of continental foundering and distension in the early development of the Appalachian geosyncline (s.l.).

Scope of the present work

Geological reconnaissance of the Strait of Belle Isle region was undertaken to provide regional geological information on an area in which previous investigations had been largely restricted to the coast, to provide background for possible engineering works in the Strait of Belle Isle, and to assess the economic mineral potential of the region. Mapping was begun in 1969 with two two-man traverse teams in the southern part of the map area where preliminary study of air photographs indicated that an extensive and reasonably well exposed section of gneiss perpendicular to foliation could be examined. Work was laid out in a strip 8 to 18 km wide across the highlands with traverses on a three by one kilometre grid. Early termination of the field season halted this more detailed work 13 km from the west end of the strip. The remaining region of contiguous Precambrian rocks to the north was mapped mostly in 1970 with similar traverse teams, but using a somewhat more dispersed traverse pattern radiating from points of access. The east coast of the Long Range Mountains and the north coast of the Strait of Belle Isle were mapped mostly in June 1971 by a single traverse team using an outboard-powered inflatable rubber boat. A Bell 47G2A helicopter was used for completion of mapping in the Torrent River area of the Long Range in July, and for inland reconnaissance in Labrador and Quebec in August and September. The latter reconnaissance was carried out with traverses spaced at 3 km intervals and landings mostly every 1.5 to 6 km. This phase of the work was severely curtailed by fog and high winds, the latter typically increasing in velocity during the day. These hazards resulted in almost all traverses being completed in patchwork fashion. Few landings could be made in north-trending valleys when flying against the wind because of the necessity of returning to the leeward ridge in order to find an updraft to clear the ridge to windward. Belle Isle was visited for two weeks in July 1972 and was mapped by a single traverse team supported by boat transport provided by the lighthouse keepers.

Topography and drainage

The Strait of Belle Isle map area lies at the eastern extremity of the Canadian Shield athwart the boundary between the Laurentian and Appalachian physiographic regions (Bostock, 1970). It includes parts of the Mecatina Plateau, East St. Lawrence Lowland, Newfoundland Coastal Lowland, Newfoundland Central Lowland, and Newfoundland Highlands.

The southeastern corner of the Mecatina Plateau, which forms the northern part of the map area, may be divided into two physiographic subregions that join approximately at the Pinware River. In the eastern subregion the land surface rises rapidly but evenly northward from the strait to upland coastal summits at elevations between 330 and 450 m. This sloping surface is dissected by several short steep-walled valleys, the most pronounced being those near Temple and Wreck bays. Farther inland to the north the surface is more even. West of Pinware River the land surface is dominated by a circular topographic feature, approximately bisected by the coastline, that extends from Pinware on the east to Baie des Belles Amours on the west. The rock surface within this circular feature has low relief and is characterized by many lakes. Only the Bradore Hills near the centre stand high and form prominent land marks. The circular feature is surrounded by more rugged topography, with highest upland summits on the north near 550 m in elevation. Toward the north boundary of the map area the summits are of lower elevation and the land surface is more subdued.

Pinware River is the principal stream draining the central part of the map area north of the Strait of Belle Isle. It, and St. Paul River, tributaries of which reach the western margin of the map area, rise on the Mecatina Plateau and have cut down through the region of higher summits and more rugged topography that borders the coast.

The Newfoundland Highlands on the Great Northern Peninsula are partly underlain by Precambrian rocks that form the Grenville Inlier of northwestern Newfoundland. The drainage divide and crest of the highlands run northeasterly near the western margin of the Precambrian rocks, but the highest point (620 m) within the Strait of Belle Isle map area is underlain by Paleozoic rocks that form the Highlands of St. John. West-facing slopes, underlain by crystalline rocks, descend fairly rapidly from the height of land to the Precambrian-Paleozoic contact. To the east, however, the upland surface slopes more gently toward the east coast, fiord-like inlets project inland for 7 or 8 km, and local relief in places reaches over 300 m. To the north, the upland surface slopes gently north and gradually merges with sea level.

Pleistocene

The Pleistocene geology of areas north and east of the northern Long Range has been investigated by Grant (1970, 1972). He suggested that four glacial events are recorded in the area:

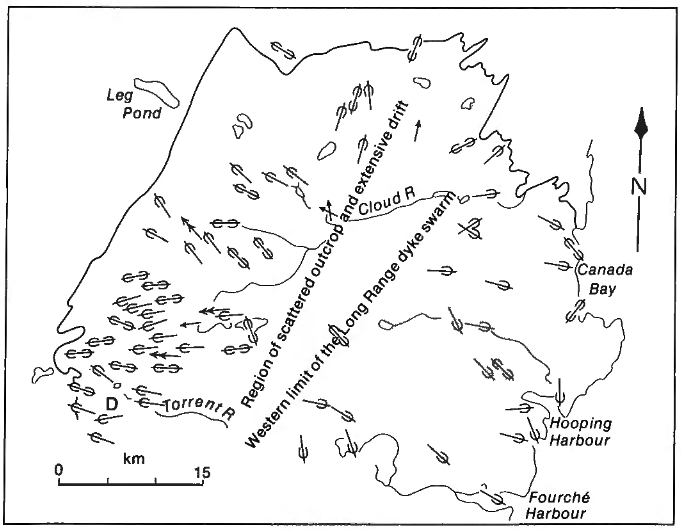
(1) Laurentide ice from Labrador advanced southeast over at least the northern lowland portion of the Great Northern Peninsula.
(2) Laurentide ice retreated under the influence of a calving bay enlarging northeastward along the Strait of Belle Isle.
(3) Ice from the Long Range highlands cap readvanced as piedmont lobes.
(4) Flow continued radially during final retreat to an ice shed along the median line of the plateau.

Striae, lunate scars, crescentic fractures (Fig. 2) and boulder trains, mapped during the present work, support the conclusions of Grant (1970) in demonstrating radial flow of late ice from an ice cap in the central part of the highlands (Fig. 3). Ice flow from this cap originated east of the present drainage divide and was located, at least for a short period, east of the western margin of the diabase dyke swarm, as suggested by the presence of diabase erratics along the western margin of the highlands. A single set of striae observed on sandstone of the Bradore Formation at the northern limit of outcrop of the Precambrian rocks may represent earlier flow or late flow related to draw-down along Castors River valley. Pleistocene features observed in Labrador indicate only the latest southward to southeastward movement of ice (Fig. 4). On Belle Isle glacial striae on the upland surface reflect the latest southeastward flow of Laurentide ice. On the northeastern part of the island striae trend close to 110° but on the central and southwestern parts more southerly trending striae were also observed. The direction of ice movement is clearly indicated by a train of white chert and limestone erratics that decrease in number eastward from the west shore of Belle Isle where the White Point Formation, from which these erratics were derived, outcrops.

Drift cover is thin and outcrop fairly abundant over most parts of the northern Long Range Mountains. However a region of subdued relief near the centre line of the uplands, where outcrop is particularly scarce, may represent the area of ice accumulation from which movement was radially outward. North of the Strait of Belle Isle, outcrop is abundant south of the higher upland summits but farther north, particularly near the headwaters of Pinware River, it is scarce. On Belle Isle, drift cover is moderate along the northwestern margin of the uplands particularly south of

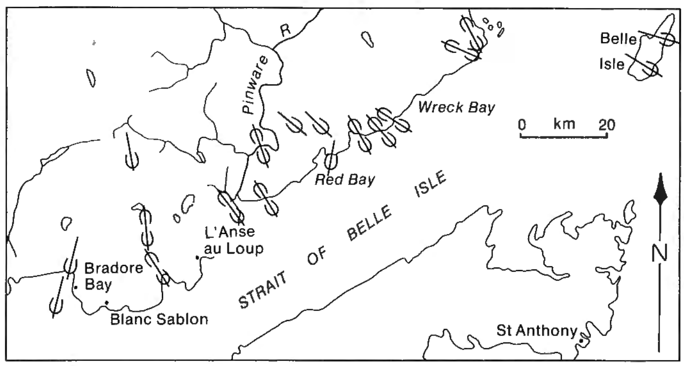
Figure 2. Striated surface with lunate scars and crescentic fractures in gneiss east of Red Bay. Ice flowed in the direction of the knife blade (190 degrees). Distribution of striae shown in Figure 3. (GSC 160098).

most of the north coast of the Strait of Belle Isle the Precambrian rocks show little sign of internal disintegration due to postglacial weathering, except where they occur close to the Bradore-gneiss unconformity. Locally, as in the valley north of Forteau, weathering of the rock surface has caused quartz veins to attain a positive relief of from 6 to 10 mm, but the bordering rock seems coherent. In contrast, rocks over an extensive region in the northwest corner of the map area show severe surface weathering. On the upland surfaces in this region, rocks struck by a hammer commonly disintegrate into coarse sand-sized mineral fragments. At several localities where the outcrop surface is horizontal it is pitted by masses of small circular basins up to 15 cm in diameter presumably caused by frost action. In spite of the ease with which they are disaggregated, these rocks are not severely argillized. The alteration may perhaps be a periglacial phenomenon related to freezing and thawing of water along grain boundaries in exposures that have not been protected by heavy snow cover either during recent winters or during some period in the past.



Glacial striae (direction of ice movement known, unknown)........
Crescentic or lunate fractures (direction of ice movement known).....
Distinctive boulder trains (direction of transport indicated)...........
Position of diabase erratic presumably derived from the Long Range dyke swarm east of the drainage divide...........................D

Figure 3. Some glacial features of the northern Grenville Inlier, Newfoundland.



Glacial striae (direction of ice movement known, unknown)..........

Figure 4. Some glacial features of the Strait of Belle Isle region.

Wreck Cove. The drift thins eastward across the island, and ponds become more numerous. Small raised beaches as high as 120 m above sea level are prominent in the cirque-like valley northeast of Barbers Cove but were not observed elsewhere. These beaches are composed of platy fragments of local greenstone, commonly 15 to 20 cm in diameter showing very little abrasion, and a few scattered cobbles of gneiss. Similar rocks with a greater proportion of fines make up sloping terraces observed along creeks inland from Green Cove and Lighthouse Cove.

Postglacial rock weathering is minimal over the greater part of the northern Long Range. Ancient rock is preserved, however, as a regolith some 1.5 m thick at the unconformity below the Bradore Formation near Pikes Feeder Pond. Along

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GENERAL GEOLOGY

Introduction

The oldest rocks within the map area comprise the basement gneiss complex of leucocratic to melanocratic gneiss with some interbanded quartz-rich gneiss. The widespread occurrence of thin quartz-rich gneiss bands suggests that the gneiss is mostly of sedimentary origin. The minimum age of the gneiss is Helikian.

TABLE OF FORMATIONS

Era	Period	Formation	Lithology	Maximum Thickness (metres)
Paleozoic	Cambrian	Forteau (includes Devils Cove at Canada Bay and White Point on Belle Isle)	Shale, limestone	55**
		Bradore	Arkosic sandstone, siltstone, conglomerate; some glauconitic sandstone; minor phosphatic sandstone and iron-rich beds	85***
Proterozoic	Hadrynian	Lighthouse Cove; Long Range dyke swarm	Tholeiitic basalt, agglomerate; diabase and gabbro dykes	300*
			Possible unconformity	
		Bateau	Quartzite, conglomerate, arkosic sandstone, siltstone, slate; minor volcanic rocks	240*
	? — ? — ? — ?		Unconformity	
	Helikian		Basic dykes of southeast Labrador	
			Intrusive Contact	
		Fourché Harbour and Cloud River plutons	Massive, largely equigranular, granodiorite, quartz monzonite, and granite	
			Intrusive Contact	
		Lake Michel, Leg Pond, Horse Chops, Hooping Harbour, and satellite plutons	Megacrystic granodiorite, quartz monzonite, and granite	
			Intrusive Contact	
			Hornblende granodiorite to granite Mangerite	
			Intrusive Contact	
			Chiefly foliated granodiorite, some quartz monzonite and granite	
			Intrusive Contact?	
			Diorite Amphibolite, hypersthene amphibolite, meta-troctolite, meta-ultramafite, norite, mangeronorite	
			Intrusive Contact	
	Aphebian and/or Helikian	Basement gneiss complex	Leucocratic to melanocratic gneiss, some quartz-rich gneiss, pelitic gneiss and amphibolite, minor calc-silicate rocks	
		Fourché Point schist	Quartz-eye schist	
		Torrent Cove assemblage	Phyllite (phyllonite), schistose mesocratic gneiss; some muscovite-chlorite schist and quartz-rich gneiss	300

*Williams and Stevens, 1967
**Schuchert and Dunbar, 1934
***Cumming, 1970

A large number of small basic plutons including what are now amphibolite, hypersthene amphibolite, metatroctolite, meta-ultramafite, norite, and mangeronorite were intruded into the gneiss throughout the map area. These were followed in the northwest part of the map area by larger mangerite and hornblende granite plutons. This sequence of intrusions forms part of the anorthosite suite that is widespread elsewhere in the Grenville Province. Regional metamorphism of amphibolite to granulite grade (throughout at least the western part of the map area) occurred about the time of emplacement of the mangerite plutons. Small bodies of mainly foliated granitic rocks, present in the gneiss along the north shore of the Strait of Belle Isle, are probably older than the mangerite, but may be younger than the metagabbro. A diorite plug in the northern Long Range is also probably older than high-grade metamorphism.

Megacrystic granitic plutons of the Long Range were intruded after the high-grade metamorphism because textures and mineral assemblages characteristic of granulite facies are sporadically preserved in the surrounding gneiss, but are not related to the plutons. Pelitic rocks along their contacts locally contain andalusite younger than sillimanite. Because the gneiss along the margins of these plutons commonly dips at moderate angles toward the plutons, these plutons are considered to have been diapirically emplaced and to have spread laterally at higher levels. A distinctive group of rocks composed of phyllonite, schistose granitic gneiss, and minor quartz-rich gneiss and muscovite-chlorite schist, here called the Torrent Cove assemblage, is thought to represent a partly mylonitized zone developed during emplacement of the southeastern megacrystic pluton. The age of emplacement of the megacrystic plutons is 1130 ± 90 Ma based on a Rb-Sr isochron (Pringle et al., 1971). Massive, mostly equigranular granite bodies also postdate high-grade metamorphism, but their age relations to the megacrystic plutons are unknown.

The Fourché Point (quartz-eye) schist is present only in the Fourché Point area where it is in fault contact with gneiss and massive granite. The schist is thought to be of cataclastic origin and, because it is veined and locally intruded by pegmatite, cataclasis is probably of late Grenvillian age.

A few small basic dykes striking approximately north and east intrude the gneiss north of the Strait of Belle Isle. A K-Ar whole-rock age suggests that these are of late Grenvillian age.

The Bateau Formation, composed of quartzite, conglomerate, siltstone, slate, and minor volcanic rocks, is present only at the northeast end of Belle Isle. In part the formation lies unconformably on Precambrian gneiss, but in part it has been faulted into wedges of variable orientation possibly prior to emplacement of the Long Range dyke swarm.

The Lighthouse Cove Formation, consisting predominantly of massive to rarely pillowed, tholeiitic basalts, locally with agglomerate, breccia and minor arkose between flows, is restricted to the eastern part of the map area. It occurs at the base of the Bradore Formation with which it is interstratified. Tholeiitic dykes, which intrude the basement gneiss complex, megacrystic and massive granitic plutons, and the Bateau Formation, locally are feeders to the Lighthouse Cove Formation. Dykes consisting of successive intrusions of diabase, and breccia dykes at Canada Bay show complex age relations that may reflect more than one period of dyke emplacement, or may result from faulting during dyke emplacement. K-Ar whole-rock dates on the flows are consistent at about 420 Ma suggesting that the flows have been degassed during the Ordovician orogeny and uplifted. Their apparently conformable position at the base of the Lighthouse Cove-Bradore-Forteau formational sequence suggests that the flows are not very much older than the first fossil-dated rocks above them (the Forteau Formation, 540 Ma).

The Bradore Formation (including the Cloud Mountains Formation of Betz, 1939) lies unconformably upon the basement gneiss complex, its base being widely marked by a regolith up to 2.5 m or more thick. The formation is 45 to 85 m thick and consists predominantly of arkosic siltstone, sandstone, and conglomerate with some local glauconitic, and minor quartz-rich, iron-rich, and phosphate-bearing variants.

The Forteau and Devils Cove formations of late Early Cambrian age, lie conformably upon the Bradore Formation and are composed chiefly of shale with some limestone beds. The White Point Formation, known only on Belle Isle, is composed of chert, fragmental limestone, and siltstone. The formation may be younger than the Forteau Formation (Williams and Stevens, 1969) but the contact is faulted and relative ages are not known. On map 1495A (in pocket) it is shown as part of the Forteau Formation.

Rocks of cataclastic origin including protomylonite and felsite-like rocks are present in fault wedges and as other small bodies in the gneiss along and near the northeast margin of the Precambrian inlier. The age of cataclasis of these rocks is not clearly defined but because they locally overlie Bradore arkose, cataclasis may in part or entirely postdate the Bradore Formation.

Deformation of the basement gneiss has been complex with northeast-trending axes of deformation predominating but of differing age in different parts of the map area. In part this deformation is thought to have occurred during Elsonian orogeny when it was accompanied by plutonism related to the anorthosite suite. Younger deformation with northeast-trending axes affected rocks mainly in the eastern part of the map area probably during or soon after emplacement of the megacrystic granite plutons and was accompanied by a greenschist facies metamorphic overprint. A final phase of deformation with northeast-trending axes occurred over a more restricted region along the east margin of the map area where, during the Ordovician orogeny, it affected chiefly supracrustal rocks. This phase of deformation was probably accompanied by a second greenschist facies metamorphic overprint that is evident in the Long Range dyke swarm within the basement gneiss complex. Of more restricted extent are areas dominated by northwesterly trending structures that are also probably of more than one age in different parts of the map area, but which may in part be related to the emplacement of the megacrystic plutons.

Reverse, strike-slip, and normal faults are evident within and along the margins of the basement gneiss complex, but the first two types are concentrated along the east and west margins of the Grenville inliers of Belle Isle and the Long Range Mountains. The movements associated with the eastern faulting appear to have occurred from late Grenville to post-Mississippian time, but those along the west margins of the inliers are of post-late Early Cambrian age and may be entirely post-Early Ordovician.

Northeast-trending normal faults are evident between the marginal fault zones where the contact between gneiss complex and supracrustal rocks has been offset. Both in the Long Range and on Belle Isle this normal faulting is post-late Early Cambrian and on Belle Isle it can be seen to be partly earlier than reverse faulting.

In Labrador and Quebec north- to northeast-trending faults are probably normal faults of late Precambrian age. More easterly trending normal faults have offset the supracrustal rocks and are of post-late Early Cambrian age.

Basement gneiss complex

Introduction

The basement gneiss complex, which underlies roughly 65 per cent ($3000\,km^2$) of the Precambrian terrane within the map area, comprises mainly medium-grained schistose gneiss and lenticularly banded gneiss, with a lesser proportion of layered gneiss and some schist. By far the greater part is biotite (± chlorite)-quartz-microcline-plagioclase gneiss, but hornblende is a prominent constituent of some of the gneiss, and some quartz-rich gneiss, pelitic schist, amphibolite, and calc-silicate gneiss are present.

The gneiss complex is probably mostly of sedimentary origin because thin bands and remnants of quartz-rich gneiss and pelitic schist are widespread. Metaconglomerate (?) though rare is further suggestive of this origin. The possible absence of pelitic gneiss from the eastern part of the complex may reflect a significant easterly change in the original sediments from which the gneiss complex was derived.

The interrelationships of the various units of the gneiss complex have been obscured by deformation so that their relative ages are unknown. There is some suggestion that the part of the complex in the Long Range may consist mostly of two major lithostratigraphic layers such that a layer of predominantly darker coloured gneiss lies on top of lighter coloured gneiss. Discussion of this speculation is deferred to the section on structural geology. The age of the sedimentary sequence from which the gneiss complex was derived is also unknown, but emplacement of plutons of the anorthosite suite (mangeronorite, mangerite, and hornblende granodiorite) within the complex suggests that the sedimentary sequence is Helikian or Aphebian or both.

Subdivision of the gneiss

The greater part of the gneiss complex, which consists of biotite (± chlorite)-quartz-microcline-plagioclase gneiss, has been divided arbitrarily on the basis of colour index (volume per cent of dark minerals) into leuco-, meso-, and melanocratic units. Unit ranges are leucocratic gneiss 0 to 5 per cent, mesocratic gneiss 5 to 15 per cent, and melanocratic gneiss 15 per cent or greater, and are based on estimates of dark minerals made in the field. The less abundant but more distinctive lithologies comprising pelitic schist, quartz-rich gneiss, amphibolite, and calc-silicate gneiss only locally form bodies of mappable size. Rocks of significantly different colour index in terms of the present classification are in places interbanded or interlensed at varying scales up to the point where they become mappable as separate units. Where lithologies of two distinct colour indices are interbanded, both forming significant proportions of the rock, the gneiss has been generalized as mesocratic.

Variation in relative detail of mapping involving systematic traversing on foot in the southern part of the map area and helicopter reconnaissance in the interior north of the Strait of Belle Isle has resulted in nondifferentiation of most gneiss units in Labrador and Quebec. It is possible that large areas of undifferentiated, predominantly leucocratic gneiss are present in the northwest corner of the map area between the hornblende granite plutons and northwest of Wreck Bay, but even these areas contain some darker gneiss and meaningful contacts could not be drawn.

Variation in metamorphic history of the gneiss units in different regions of the map area has produced variations in texture and mineralogy of the gneiss within their respective gneiss units. It is therefore expedient to outline regions of differing metamorphic history while leaving discussion of the data on which they are based to the section on metamorphism. These regions are referred to as 'terranes' (rather than zones) because each is, or may be, the product of a different, broadly defined, dominant phase of regional metamorphism, the criteria for which include both mineral assemblages and mineral textures. The distribution of metamorphic facies terranes is as follows:

(1) The Granulite terrane underlies a region north of the Strait of Belle Isle between Bradore Bay and Red Bay, and extends north-northwestward to the northern limit of the map area. The south margin of this terrane lies close to the coast.

(2) The Greenschist terrane underlies a region along the east coast of the Long Range extending inland about 27 km at Fourché Harbour. Farther north the boundary is less well known but it appears to veer northeasterly near Cloud River. Belle Isle is included within this terrane.

(3) The Amphibolite terrane comprises the remaining regions where basement gneiss is exposed.

Leucocratic gneiss

Distribution. Bodies of leucocratic gneiss are evenly distributed throughout the map area, except that they are not as common on Belle Isle. In the northern Long Range they commonly occur in belts or at the cores of domes where they reach over 300 m apparent thickness (perpendicular to foliation), but they are also present as bands or lenses down to a few millimetres thick in the darker gneiss. Perhaps the greatest apparent thickness of leucocratic gneiss is exposed at Canada Bay where over 1000 m are present without the base being exposed; however, in spite of local regularity of foliation and jointing displayed by these rocks, the preservation here and there of linear amphibolite inclusions indicates that the rocks have been severely deformed. The original thickness of these rocks is thus unknown.

Megascopic Features. The leucocratic gneiss is pale pink, or rarely pale grey to white or pale green, and where weathered, commonly develops a pale pink to white crust with an internal deep orange stain. Massive ledges or layers from several centimetres to 6 m or more thick are evident locally (Fig. 5). Bands of mesocratic or melanocratic gneiss are present locally (Fig. 6). In places, as along the north shore of Fourché Harbour east of Williamsport, bands of schistose and lineated melanocratic gneiss (possibly originally beds or sills) up to 2 m thick are intercalated in the leucocratic gneiss from 3 to 30 m apart. Generally the rock is fine to medium grained with granoblastic or schistose texture, but fine-grained varieties locally appear granulose. Thin sections of such granulose gneiss, however, reveal no trace of original sedimentary detritus. The coarser grained leucocratic gneiss commonly resembles granite in hand specimen. Small bodies of pegmatite are widespread but are not abundant.

Mineralogy. The principal minerals in the leucocratic gneiss are quartz, microcline and plagioclase (chiefly albite to sodic oligoclase, but rarely calcic oligoclase). Variants range to quartz-rich gneiss. In feldspar-rich varieties either microcline or plagioclase may predominate. Plagioclase crystals commonly have sodic rims that are widest adjacent to microcline crystals. Small amounts of quartz-plagioclase myrmekitic intergrowth are common.

Minor minerals in the leucocratic gneiss are mostly brown to green biotite and magnetite or dark opaque minerals. In places these opaques are the only mafic constituent. Hornblende occurs in leucocratic gneiss from the core of the dome some 14 km east of Pikes Feeder Pond (the north Torrent River dome), and is common in the granulite terrane of Labrador and Quebec. Clinopyroxene and hypersthene are present locally in leucocratic gneiss in the granulite terrane. Epidote, chlorite, and muscovite are common alteration products.

The most common accessory minerals are zircon, apatite, sphene, and allanite in order of decreasing abundance. Rare accessories are carbonate, pyrite, fluorite, and garnet, the latter occurring at widely scattered localities as isolated crystals or groups of crystals. Modes of examples of leucocratic gneiss from the Long Range are given in Table 1, and from Labrador and Quebec in Table 2.

Textures. Leucocratic gneiss is typically medium grained (1 to 3 mm) and commonly granoblastic but foliation due to alignment of biotite or chlorite may be evident. More rarely, felsic minerals are conspicuously flattened. In some areas a linear fabric is evident either in linear concentrations of biotite, or in rodding of quartz. Locally the rock appears granulose but such rocks when seen in thin section consist of an interlocking mosaic of unrounded grains.

Mesocratic gneiss

Distribution. Mesocratic gneiss occurs throughout the map area. No estimate of thickness of this gneiss is practical with available data.

Megascopic Features. Mesocratic gneiss is typically pale grey, grey, pale green, or pink. Pronounced ledges (cuestas) occur locally where leucocratic gneiss is interbanded but, for the most part, outcrops are less rugged than those of the leucocratic gneiss. Banding of varying thickness is evident locally, some outcrops being finely banded (up to 30 cm), but commonly banding shows varying degrees of disruption to lenticular layers, and in many places parallel orientation of mineral grains and lenses is the only obvious element of foliation. Locally, such as south of Booney Lake, melanocratic and mesocratic interbands are abundant in leucocratic gneiss. Elsewhere melanocratic gneiss remnants are present in leucocratic to mesocratic gneiss, and such combinations have been generalized as mesocratic gneiss.

Mineralogy. The principal minerals of the mesocratic gneiss are quartz, microcline, and plagioclase (albite to sodic andesine, mostly calcic oligoclase) but in contrast to the leucocratic gneiss, biotite is present in almost all exposures. Variants range to quartz-rich gneiss. Either microcline or plagioclase may predominate but plagioclase is most commonly predominant in mesocratic gneiss from the Long Range.

Figure 5. Leucocratic gneiss in massive layers west of Hooping Harbour. (GSC 153336).

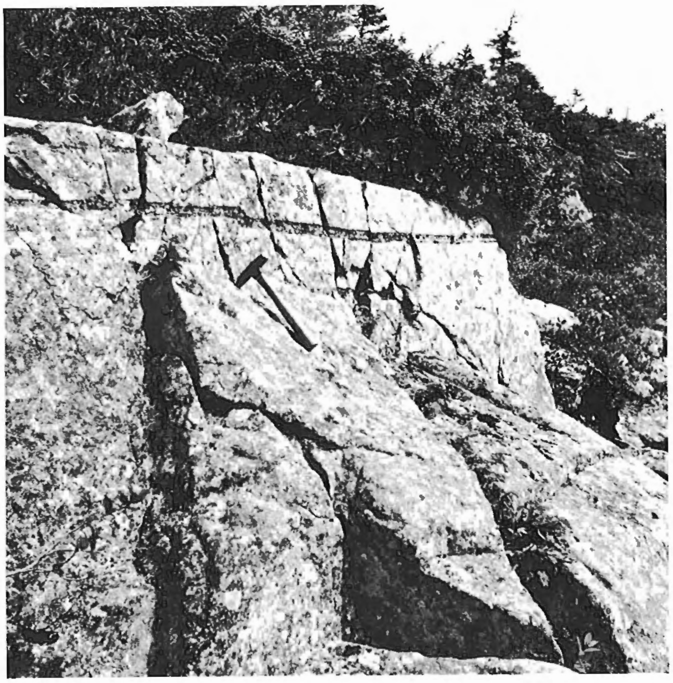


Figure 6. A 5-cm band of melanocratic gneiss forms the only evident structure in a massive leucocratic gneiss layer south of Cloud River. (GSC 153354).

Minor minerals are mostly biotite and hornblende, the latter being present in about one third of the mesocratic gneiss. Pale green hornblende is common along the east coast of the Long Range but elsewhere the pleochroism is, *x* ochre, *y* green or brown-green, *z* blue-green to green or olive. Hornblende is most abundant in mesocratic gneiss north of the Strait of Belle Isle. Cummingtonite and anthophyllite were observed in mesocratic gneiss in the vicinity of the twin lakes on Cloud River where they are associated with red-brown biotite similar to that found in the pelitic rocks, garnet, and hornblende. Hypersthene rimmed and veined by fine-grained alteration products, and showing bright red and green pleochroism, occurs with garnet and anthophyllite in mesocractic gneiss in the same area. Pale green hypersthene and clinopyroxene are present at scattered localities in mesocratic gneiss in the granulite terrane. Epidote, chlorite, and muscovite are alteration products in mesocratic gneiss.

The most common accessory minerals are apatite, zircon, dark opaque minerals, and sphene in order of decreasing frequency of occurrence. Carbonate and pyrite are less abundant and fluorite is rare. Garnet, although more frequently observed than in the leucocratic gneiss, is not common. Modes of examples of mesocratic gneiss from the Long Range are given in Table 1 and from Labrador and Quebec in Table 2.

Textures. The mesocratic gneiss is mostly medium grained, but some is fine grained, and in contrast to the leucocratic gneiss, it is locally megacrystic with feldspar megacrysts larger than 3 mm. The rock is mostly foliated with parallel or linear arrangement of mafic minerals in lenses, or within banding where it is present. Other textures and mineral intergrowths are similar to those of the leucocratic gneiss.

Table 1. Modal analyses of gneiss from the basement gneiss complex, Long Range Mountains

Colour Index	Metamorphic Terrane	Quartz	Mirco-cline	Plagio-clase	An	Biotite	Horn-blende	Chlorite	Epidote	Muscovite	Dark Opaques	Sphene	Apatite	Garnet	Allanite	Zircon	Pyrite
1	Greenschist	36.8	36.1	25.5	11	0.3	-	0.2	0.3	0.3	0.3	-	tr	-	-	tr	-
2	Amphibolite	36.9	19.8	40.1	10	2.0	-	tr	tr	0.9	0.3	-	tr	-	-	tr	-
4	Greenschist	32.1	27.0	36.4	0	3.3	-	0.6	0.3	tr	-	-	tr	0.3	-	tr	-
5	Amphibolite	29.8	29.7	35.8	4	4.5	-	0.2	-	-	-	-	-	-	-	tr	-
6	Amphibolite	23.3	31.8	39.1	17	3.7	0.4	0.3	0.7	-	0.5	0.2	tr	-	0.1	tr	-
7	Greenschist	28.2	17.1	47.4	21	6.1	-	0.2	0.1	0.3	-	0.5	0.1	-	tr	-	-
9	Greenschist	31.6	22.7	36.2	11	7.8	-	0.1	0.6	0.8	0.1	-	tr	-	0.1	tr	-
14	Greenschist	29.9	3.8	52.8	27	13.0	0.4	-	0.1	-	-	-	tr	-	-	-	-
14	Amphibolite	33.3	-	52.5	32	13.8	-	0.3	-	tr	-	-	0.1	-	-	tr	-
17	Amphibolite	31.1	10.2	42.0	28	11.0	5.2	-	-	-	-	0.4	0.1	-	-	tr	-
18	Greenschist	23.6	3.3	54.8	26	11.4	5.8	0.1	0.3	-	-	0.5	tr	-	0.1	-	-
20	Amphibolite	28.1	0.4	51.1	27	19.9	-	tr	0.1	tr	-	0.3	0.1	-	tr	tr	-
4*	Greenschist	57.1	14.3	23.1	S	2.9	-	-	0.8	1.5	0.3	-	tr	-	-	tr	-
5*	Greenschist	53.7	30.2	11.1	S	3.0	-	-	0.1	0.2	1.3	0.4	-	-	-	tr	-

*Quartz-rich gneiss; tr trace.

Note: Twelve hundred points were counted for each specimen spread over 1, 2, or 3 thin sections depending on the grain size requirement as determined by Chayes (1956) to yield an average analytical error <2.0 for major minerals. (Hypersthene-bearing gneiss is rare and therefore none was included among the specimens counted.) Analyses are arranged in order of increasing colour index (volume per cent of dark minerals). An values refer to per cent anorthite determined by refractive index except that plagioclase too altered for determination is indicated by S.

Table 2. Modal analyses for gneiss from the basement gneiss complex, Labrador and Quebec

Colour Index	Metamorphic Terrane	Quartz	Micro-cline	Plagio-clase	An	Biotite	Horn-blende	Pyroxene	Chlorite	Epidote	Muscovite	Dark Opaques	Sphene	Apatite	Calcite	Zircon	Pyrite
2	Amphibolite	33.8	36.9	27.7	2	-	-	-	0.8	-	-	0.8	tr	tr	-	tr	-
3	Amphibolite	16.6	15.3	64.3	15	0.9	0.9	-	0.4	-	1.0	0.7	-	tr	-	tr	-
4	Granulite	21.3	46.4	28.8	12	1.6	0.7	-	-	-	-	1.0	0.2	tr	-	tr	-
8*	Granulite	53.5	15.4	23.5	34	4.8	-	1.3[1]	0.7	-	-	0.8	-	tr	-	tr	-
9	Granulite	23.3	25.7	42.3	27	5.3	2.3	-	-	-	-	1.0	-	0.1	-	tr	-
10	Granulite	-	36.1	53.8	15	5.3	1.8	-	0.2	-	tr	1.1	1.3	0.4	0.1	tr	-
11	Granulite	10.8	33.4	44.5	18	3.8	4.0	2.0[2]	tr	tr	-	1.4	tr	0.1	-	tr	tr
16	Amphibolite	7.7	37.0	38.5	0	8.8	5.4	-	tr	-	-	0.3	1.6	0.8	-	tr	-
38	Amphibolite	9.8	6.8	45.0	25	15.8	18.0	-	-	3.6	-	-	0.8	0.2	-	tr	-
38	Amphibolite	3.5	1.8	55.3	24	17.5	16.8	-	1.2	1.5	-	tr	1.2	1.3	-	-	-

*Quartz-rich gneiss; [1]Hypersthene; [2]Clinopyroxene; tr trace.

Note: Twelve hundred points were counted for each specimen spread over 1, 2, or 3 thin sections depending on the grain size requirement as determined by the method of Chayes (1956) to yield an average analytical error <2.0 for major minerals. Analyses are arranged in order of increasing colour index (volume per cent dark minerals). An values refer to per cent anorthite determined by refractive index.

Melanocratic gneiss

Distribution. Melanocratic gneiss is more restricted in occurrence than either of the lighter coloured gneiss units, but nevertheless is represented in all regions of the map area. Well banded melanocratic gneiss with an apparent thickness of 180 m or more is unusually well preserved in a fault-bounded remnant just west of the greenschist terrane 10 km south of Cloud River.

Megascopic Features. The melanocratic gneiss is chiefly grey or grey-green. Except where stained red, the weathered surface is grey, or locally grey-brown and crumbly. Where the rock is banded melanocratic layers of varying thickness are commonly separated by mesocratic or leucocratic bands. Over extensive areas, however, banding persists only as remnants or the rock is homogeneously schistose. At a single locality in the hills immediately west of Hooping Harbour a band of melanocratic gneiss (20 cm or more thick) in leucocratic gneiss contained elongate ellipsoids, possibly stretched pebbles (Fig. 49). Other examples of minor metaconglomerate beds within the basement gneiss complex have been reported elsewhere by Fritts (1953); and by Pontin (1967), but none of these were found during the present study. Highly schistose zones of melanocratic to mesocratic gneiss, in which the banding has presumably been destroyed by penetrative deformation, are present along the northeastern margin of the northwestern megacrystic pluton (Leg Pond pluton) and along the valley between Northwest Brook and twin lakes on Cloud River. Large silver-like remnants of leucocratic gneiss are included locally within these schistose zones. Small bodies of pegmatite, though not abundant, are widespread and more obvious than those in the lighter coloured gneiss.

Mineralogy. The principal minerals in the melanocratic gneiss are plagioclase (albite to sodic andesine, chiefly sodic oligoclase) and biotite with quartz present in 90 per cent and hornblende and microcline each present in half of the gneiss examined. Most of the melanocratic gneiss has more than 10 per cent of quartz and this characteristic has been used to distinguish some hornblende-rich gneiss bands from metagabbro and amphibolite. The pleochroic formula for hornblende is, x ochre, y green to brown-green, z sea-green, green, grey-green, or brown-green. Crystals from the granulite and amphibolite terranes commonly have bluish green rims parallel to the z direction. Anthophyllite, associated with garnet and red-brown biotite, occurs in some melanocratic gneiss near the twin lakes on Cloud River. Cummingtonite, showing polysnthetic twins, is present at one locality in the same area. Pale green hypersthene and clinopyroxene are found almost exclusively in the melanocratic gneiss of the granulite terrane. Epidote, chlorite, and more rarely muscovite are present locally in melanocratic gneiss, and garnet, though widely distributed, is not common. Accessory minerals in the melanocratic gneiss are apatite, zircon, sphene, allanite, dark opaque minerals, pyrite, carbonate, and fluorite in order of decreasing frequency of occurrence. Modes of examples of melanocratic gneiss from the Long Range are given in Table 1 and from Labrador and Quebec in Table 2.

Textures. The melanocratic gneiss is most commonly medium grained (1 to 3 mm) but finer and coarser grained examples are known. The finer grained melanocratic gneiss tends to be banded whereas the coarser gneiss is commonly schistose and contains plagioclase megacrysts as in the area north of Torrent River. Linear fabric is common in the melanocratic gneiss and is typically expressed as elongate patches of biotite.

Quartz-rich gneiss

Distribution. Quartz-rich gneiss of unknown thickness is the predominant lithology over an area of about 10 km^2 at the far northern extremity of the Long Range, where it appears in the cores of two antiformal structures. Farther south an extensive area containing a large proportion of quartz-rich gneiss within the Hooping Harbour satellite pluton is poorly exposed. Smaller bodies of quartz-rich gneiss are present at Bradore Bay and elsewhere in southeastern Labrador and Quebec. Bands of similar gneiss from 15 cm to 2 m thick are widespread within other gneissic rocks though rare throughout the map area, but they appear to be most common along the margins of, or as inclusions within, the granitic plutons.

Megascopic Features. In outcrop the quartz-rich gneiss is mostly pale grey and weathers pale grey, but locally is purplish, white, or pale grey-green. The rock varies from massive to irregularly fractured, jointed, or thinly banded.

Mineralogy. Quartz forms from 40 to 90 per cent of most of the quartz-rich gneiss bands, and rarely, the gneiss is a true quartzite containing over 90 per cent quartz. Minor minerals are mostly microcline, sodic plagioclase, and biotite. Hornblende, garnet, sillimanite, and clinopyroxene are important constituents locally. Epidote, chlorite, muscovite and dark opaque minerals are present in some specimens. Zircon is almost always present but apatite, sphene, and carbonate are local accessory minerals.

Textures. Quartz-rich gneiss is mostly medium grained (1 to 3 mm) but some coarse-grained and fine-grained gneiss occurs. Quartz grains mostly form granoblastic mosaics but are in places flattened (Fig. 7). Commonly a medium to coarse-grained quartz host in which original quartz grains have coalesced to form large amoeboid crystals contains dispersed, fine-grained inclusions of potash feldspar or plagioclase.

Pelitic schist and gneiss

Distribution. Pelitic schist and gneiss are present in most parts of the map area but were not found in Labrador east of Pinware, on Belle Isle, or in the greenschist terrane east of Hooping Harbour pluton. At many exposures pelitic rocks occur in thin bands up to 6 m thick, but locally, such as east of Ocean Pond, they are extensively exposed and apparently thicker though complexly deformed. At the Torrent River exposures pelitic and semipelitic gneiss is up to 100 m wide, but again the rock, although forming a more or less linear band along the hillside, has been severely deformed.

Megascopic Features. The pelitic rocks weather a typically yellowish to reddish brown, and are crumbly and migmatitic in aspect. Fresh surfaces, where observed, are typically pale bluish or greenish grey. Foliation is in most cases contorted, but sillimanite is commonly oriented parallel to the regional mineral lineation and locally forms spectacularly lineated schist (Fig. 47).

Mineralogy. The principal minerals in the pelitic rocks are quartz, biotite, plagioclase (typically calcic oligoclase, locally sodic oligoclase, and rarely andesine or labradorite), potash feldspar, and sillimanite (Fig.8). Cordierite and garnet are prominent constituents locally. Chlorite and muscovite are fairly common alteration products and epidote is less common. Andalusite is present with sillimanite in pelitic gneiss along the southeast contact of the Leg Pond pluton and in an inclusion within the Cloud River pluton.

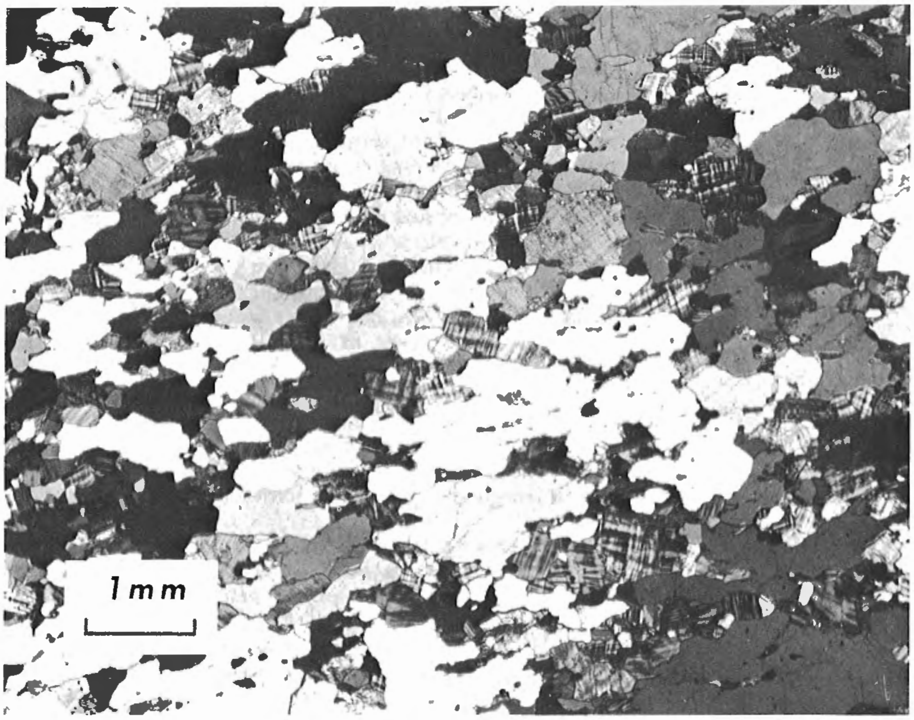


Figure 7. Flattened quartz grains sectioned perpendicular to foliation in quartz-rich gneiss from the granulite terrane. Crossed nicols (GSC 201902-B).

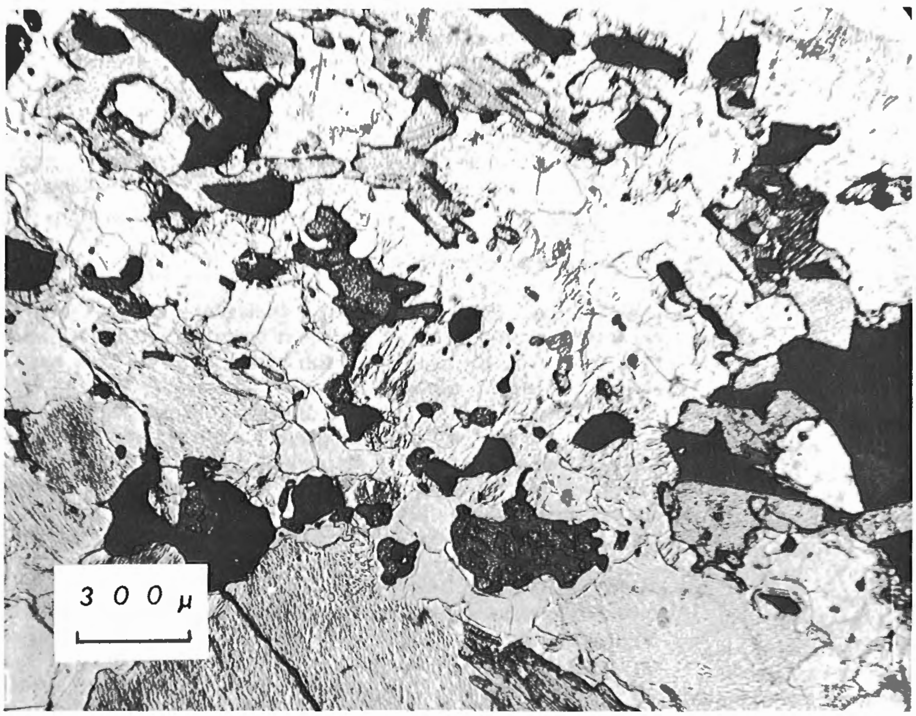


Figure 8. Pelitic gneiss from the northwest part of the granulite terrane. Mesoperthite is visible in the lower left corner. Other minerals are cordierite, chlorite, spinel, garnet, sillimanite, magnetite-ilmenite, and quartz. Polarized light (GSC 201902-P).

Dark opaque minerals are a minor constituent of most pelitic schist, and dark green spinel is present locally. Iron sulphide is evident in some thin sections and is concentrated in amounts probably reaching several per cent of the rock in a few places. The rusty weathered surface that almost everywhere accompanies the pelitic schist probably reflects a higher content of iron sulphides than is present in the enclosing gneiss. Accessory minerals are zircon and apatite.

Textures. The pelitic rocks are mostly medium grained, but may be coarse or fine grained. Where the rock is coarse grained, megacrysts of potash feldspar are commonly present. Sillimanite ranges in habit from fibrolite to coarse prismatic crystals but is mostly prismatic. It typically lies within the foliation planes of the schist and is oriented parallel to the regional lineation. Locally sillimanite appears to have formed at the expense of muscovite but more commonly it is altered to muscovite and persists where enclosed in plagioclase. Andalusite is clearly later than sillimanite in schist south of the Leg Pond pluton where it is associated with patches of green biotite. Spinel does not occur in contact with quartz but is separated from it chiefly by sillimanite, cordierite, or opaque minerals.

Calc-silicate gneiss

Distribution. Calc-silicate gneiss is widespread but rare throughout the map area. It is most abundant along the synformal axis at the crest of the dome-like structure northeast of Pikes Feeder Pond (Pikes Feeder antiform). In no case are more than about 6 m known to be exposed at one place.

Megascopic Features. The calc-silicate gneiss is mostly fine grained, thinly bedded (in places down to a few millimetres), pink and green, dark green, yellow-green or pale grey-white. Outcrops are flaggy, massive, or hackly fractured.

Mineralogy. The principal minerals in the calc-silicate gneiss are quartz and diopside, although any one of a number of minerals may be the predominant component of an individual occurrence. These include plagioclase (albite to bytownite), microcline, epidote, hornblende, garnet, chlorite, calcite, scapolite, and vesuvianite. Olivine is a major constituent in one gneiss. The most common accessory minerals are apatite and sphene. Zircon, dark opaque minerals, pyrite, fluorite, and chabazite (?) were found in one specimen only. Two gneiss bands containing hypersthene, labradorite-andesine, hornblende, cummingtonite, biotite and quartz, and a third containing quartz, andesine, and diopside are thought to be calcareous metasediments.

Textures. Calc-silicate gneiss is mostly fine grained (less than 1 mm) but less commonly it is medium grained (1 to 3 mm). Nonbedded gneiss may be massive or schistose.

Gneiss of Belle Isle

The basement complex of Belle Isle is distinct from that elsewhere in the map area in being abundantly intruded by diabase dykes that form 30 to 50 per cent of continuous exposure along the coast, and more than 90 per cent inland. Furthermore, the gneiss, although it includes rocks of a wide variety of colour indices, is mostly mesocratic to melanocratic, and where not overprinted by deformation with northeasterly trending axes displays a distinctive northwest-trending foliation.

Megascopic Features. The gneiss of Belle Isle ranges from leucocratic to melanocratic, and from well banded and lenticularly banded to schistose and locally massive. Exposures along the coast indicate that leucocratic bands, which may be 30 m or more thick, are subordinate to the darker gneiss. Banding is commonly complex with megabands up to 20 m thick including thinner bands up to 1 m thick. The banding is in many places obviously lenticular having been modified by deformation. Thin bands of quartz-rich gneiss and of calc-silicate gneiss are present within the gneiss sequence in a few places but pelitic gneiss was not observed. Along the shore south of Bateau Cove banding is less common than foliation characterized by parallel alignment of mineral grains. Locally the gneiss is intruded by amphibolite (metagabbro) bodies which have been deformed and subsequently veined by leucocratic material. One small body of slightly schistose granitic rock perhaps 200 m wide was observed within the gneiss on the south shore of Green Cove. This body has been deformed and it is not clear whether it represents an intrusion within the gneiss or is a leucocratic band that has been more severely recrystallized than similar rocks elsewhere. Small bodies of pegmatite, both parallel to and cross-cutting the foliation, are common. The gneiss is abundantly, and typically irregularly, fractured so that gneissic structures are rarely expressed in landfroms.

Mineralogy. The principal minerals are plagioclase (chiefly albite), quartz, microcline, biotite, and chlorite with some hornblende particularly in the northern part of the island. Muscovite is rare, carbonate is common, and epidote is an ubiquitous alteration product. Accessory minerals are dark opaque minerals, zircon, sphene, apatite, and allanite.

Textures. The gneiss of Belle Isle is typically medium grained (1 to 3 mm) and equigranular with schistose to granoblastic textures. Locally a linear fabric is expressed by elongate patches of biotite or more rarely by hornblende. On the shore near Bateau Cove and locally in the central part of the island the gneiss is porphyroblastic with subhedral feldspars up to 15 mm long. Quartz is locally flattened parallel to foliation and commonly shows strain shadows under crossed nicols. Microcline is mostly finely perthitic. Mafic silicates in the northeastern part of the island are extensively replaced by fine-grained, green, randomly oriented biotite.

Summary of regional variations in the gneiss

Gneiss of the granulite terrane is characterized by the presence of submesoperthitic potash feldspar. Hypersthene, though far from ubiquitous, is much more extensively present than in gneiss of the other terranes. Clinopyroxene and hornblende are more common in this terrane and hornblende shows the browner colours in thin section. Sphene was not observed in the leucocratic gneiss and is present only sparingly in the darker gneiss mostly as rims on magnetite-ilmenite. The olive-green colour typical of feldspar in gneiss of granulite facies elsewhere, though locally present, is not characterisitic of the granulite terrane.

Gneiss of the greenschist terrane is greener than in other terranes and is characterized by a greater proportion of epidote or chlorite or both. In thin section amphiboles tend to be shades of pale green, or to have *z* optic direction bluish green. Two generations of biotite are commonly present, the earlier being characterized by alignment of c crystallographic axes perpendicular to foliation in the host gneiss. This early biotite also contains many fine lenses of epidote along cleavage planes and in zig-zag patterns across it. Locally acicular (rutile?) inclusions form triangular patterns visible in sections parallel to cleavage (Fig. 43B and C). Later biotite is usually clear, randomly oriented, and in fine-grained clusters of crystals (Fig. 43D). Most plagioclase in the greenschist terrane is sodic and has complex alteration patterns. Crystals with sericitized patches of albite surrounded by clear oligoclase and rimmed by clear albite are common. Allanite crystals are commonly diffusely rimmed by epidote.

Gneiss in the amphibolite terrane typically lacks the features described above but scattered occurrences of mesoperthitic to submesoperthitic potash feldspar, and of hypersthene have been encountered in the Long Range, suggesting that this part of the amphibolite terrane barely reached granulite facies conditions or underwent amphibolite facies retrogression after granulite facies metamorphism or both. Furthermore, local retrogression has in places produced concentrations of epidote and chlorite within the amphibolite terrane. Most of the gneiss, however, has a single generation of biotite that is the predominant mafic component and is of unaltered appearance relative to that in the greenschist terrane (Fig. 43A).

Comparison of gneiss from the Long Range with gneiss from Labrador and Quebec is of interest because the two areas are separated by about 90 km of supracrustal cover. Differences between gneiss from these two areas in part reflect the better preservation and perhaps more extensive development of granulite facies metamorphism north of the strait, and the more extensive development of greenschist facies overprint in the south. Other differences, which may be related to differences in the original sediments from which the gneiss were derived rather than to metamorphism, are evident. Leucocratic and mesocratic gneiss with low quartz content (quartz less than roughly 20 per cent) are more common in Labrador and Quebec than in the Long Range and the only quartz-free gneiss bands from these units were found there (mostly in the vicinity of Red Bay). Gneiss, with plagioclase more abundant than potash feldspar, is more frequently encountered in the Long Range. Although hornblende is distinctly more common in gneiss from the granulite terrane, it is also somewhat more common in gneiss from the amphibolite terrane in Labrador and Quebec than from the amphibolite terrane in the Long Range. Pelitic gneiss on the other hand occurs in the western part of the map area and is apparently absent in the east, both north and south of the Strait of Belle Isle.

A comparison of gneiss from each of the major regions of the map area and from each of the map units is presented in Table 3, which shows the percentage of thin sections examined from each specified subdivision that contains a given mineral or mineral proportion. The number of thin sections involved in each percentage figure is given immediately below that figure.

Age relations of the basement gneiss complex

The interrelationships of the various gneiss units of this complex have been obscured by deformation, therefore, relative ages cannot be assigned to any of them. The absolute age of a sedimentary sequence from which these units were derived has been obscured by metamorphism; however minimum ages can be determined from the radiometric ages of the intrusive plutonic rocks and by comparison with similar suites of rocks in other parts of the Grenville Province where radiometric ages are more plentiful.

The basement gneiss complex throughout the northern Long Range was intruded by megacrystic granite plutons. The emplacement of the largest of these plutons, the Lake Michel pluton, has been dated at 1130 ± 90 Ma ($\lambda^{87}Rb = 1.39 \times 10^{-11}\ yr^{-1}$) by Pringle et al. (1971) on the basis of a Rb-Sr isochron. High-grade regional metamorphism of gneiss surrounding the megacrystic plutons preceded their emplacement, and hence the deposition of sediments, which eventually gave rise to the basement gneiss complex, occurred before the emplacement of the megacrystic granite plutons.

The basement gneiss complex north of the Strait of Belle Isle is intruded by small mangerite bodies. Insofar as mangerite bodies in the Grenville Province farther west are related to the anorthosite intrusions, dating of the anorthosites may provide a minimum age for the basement gneiss complex. By this criterion the minimum age of the gneiss complex would be about 1460 Ma (i.e. Helikian or earlier) based on radiometric age determinations on biotite, hornblende, and zircon from the Michikamau Intrusion (Emslie 1970; Emslie et al., 1976; Krogh and Davis, 1973).

Torrent Cove assemblage

The name Torrent Cove assemblage is proposed on an informal basis to refer to rocks of mostly distinctive petrography along the shore in the vicinity of Torrent Cove on the west side of Canada Bay. These rocks extend southward along the shore to within 2.5 km of the head of Wild Cove, where they apparently pass below sea level beneath metagabbro. Gneiss immediately overlying the metagabbro in this vicinity locally shows somewhat similar phyllitic textures and it may be related in origin to gneiss in the Torrent Cove assemblage. Rocks of the Torrent Cove assemblage extend for some 6 km northwest of Torrent Cove, beyond which they become indistinguishable from the surrounding gneiss. They comprise up to about 300 m of

Table 3. Mineral frequencies in the basement gneiss complex[1]

	No. of Thin Sections per Region			Leucocratic Gneiss (107 thin sections)											Mesocratic Gneiss (121 thin sections)											Melanocratic Gneiss (103 thin sections)										
	Leucocratic Gneiss	Mesocratic Gneiss	Melanocratic Gneiss	Quartz Absent	Quartz 1 to 20%	Quartz >20%	Kspar ≥ Plagioclase	Hornblende	Pyroxene	Epidote	Chlorite	Muscovite	Opaques	Sphene	Quartz Absent	Quartz 1 to 20%	Quartz >20%	Kspar ≥ Plagioclase	Hornblende	Pyroxene	Epidote	Chlorite	Muscovite	Opaques	Sphene	Quartz Absent	Quartz 1 to 20%	Quartz >20%	Kspar ≥ Plagioclase	Hornblende	Pyroxene	Epidote	Chlorite	Muscovite	Opaques	Sphene
Labrador	34	42	16	3	26	71	85	18	12	3	59	44	97	24	5	40	52	64	45	14	21	55	43	100	45	19	75	6	6	88	19	31	38	25	75	56
Long Range	73	79	87	0	15	85	77	7	0	41	63	64	70	34	0	33	67	28	24	4	44	63	52	60	52	13	55	32	3	41	3	46	28	25	32	61
Granulite	12	17	3	8	25	67	92	33	25	0	33	33	100	0	6	41	53	65	59	35	12	41	24	100	29	0	100	0	0	100	33	66	33	33	100	33
Amphibolite (Labrador)	22	25	13	0	27	73	82	9*	5	5	73	50	95	36	4	44	52	64	40	0	28	68	56	100	56	23	69	8	8	85	15	23	38	23	69	62
Amphibolite (Long Range)	34	40	40	0	15	85	88	15	0	18	68	50	79	35	0	40	60	35	30	8	15	65	48	60	48	20	43	38	5	33	8	22	48	28	38	55
Greenschist	39	38	47	0	15	85	67	0**	0	62	59	77	62	63	0	26	76	21	18	0	76	63	58	61	58	6	66	28	2	49	0	66	32	23	28	66

[1] The table shows the percentage of the total number of thin sections from each of two geographic areas and four metamorphic terranes that contain the designated mineral or mineral proportion.

* Hornblende remnants within chlorite. ** One thin section contained chlorite pseudomorphous after hornblende(?).

gently southwesterly dipping quartz-feldspar-mica phyllite that is overlain by a huge lens some 6 km long of schistose mesocratic gneiss northwest of Torrent Cove. Overlying the latter, and separating the rest of the Torrent Cove assemblage from megacrystic granitic rocks above, are up to 200 m of muscovite-chlorite schist and quartz-rich gneiss that probably include retrograded remnants of pelitic gneiss. Near Torrent Cove, where the schistose mesocratic gneiss pinches out, the overlying schist thickens and becomes indistinguishable from the phyllite.

The phyllite of the Torrent Cove assemblage is fine grained, and grey to white, pink, or pale green. Locally it appears banded, but commonly banding is lenticular at scales varying down to a few millimetres. South of Torrent Cove the rock is characterized by a silvery micaceous parting (Fig. 9) that exhibits a remarkably fine, regular, southwest-plunging crenulation lineation. Pegmatite boudins (Fig. 10), around which this lineation is warped, are elongate at high angles to the lineation trend. Northwest of Torrent Cove, where the phyllite is overlain by the schistose gneiss lens, the lineation is scarcely evident and is replaced by a whispy schistosity.

The principal minerals present in phyllite are quartz, plagioclase (chiefly albite), and microcline with muscovite, chlorite, and epidote. Biotite occurs locally and tremolite-actinolite and garnet are rare constituents. Accessory minerals are dark opaque minerals, carbonate, apatite, sphene, and zircon in order of decreasing frequency. The phyllite is typically fine grained (0.1 to 1 mm) but quartz or feldspar crystals up to 3 mm in diameter occur locally either scattered or in lenses. Typically the rock is schistose (and lineated) but rarely there is a massive mosaic texture. Some specimens are slightly cataclastic.

The schistose mesocratic gneiss of the Torrent Cove assemblage is medium grained, grey to pale grey or buff pink. To the northwest the rocks are more leucocratic and granitic in character whereas to the southeast they approach melanocratic gneiss. No banding or linear features were observed in this rock, although in other respects it resembles mesocratic gneiss elsewhere.

The schistose mesocratic gneiss consists principally of quartz and plagioclase (sodic oligoclase) with lesser amounts of microcline and 5 to about 15 per cent of biotite. Small amounts of muscovite, chlorite, and epidote are common. Accessory minerals are dark opaque minerals, apatite, and zircon.

The contact between phyllite and schistose mesocratic gneiss is exposed along the valley of the northern tributary to the creek flowing into Torrent Cove, where a transitional zone some 9 to 12 m thick exists between fine-grained quartzofeldspathic phyllite and medium-grained schist. This zone is characterized by whispy clots of biotite-chlorite and an increase in grain size of the matrix. In the valley of the southern tributary no schistose mesocratic gneiss was encounted, but phyllitic rocks containing biotite-chlorite whisps were observed at one outcrop.

Contacts between the Torrent Cove assemblage and the basement gneiss were not observed, but the two are exposed on either side of a small bay at the northeast corner of Torrent Cove where there is an abrupt change in grain size from about 0.5 mm or less in phyllite to 2 mm in leucocratic gneiss. Micaceous (muscovite) partings are locally present in the leucocratic gneiss near the contact. Foliation on either side of the contact is conformable, although farther north, dips in the phyllite are slightly steeper than those in the underlying gneiss. Crenulation lineation, so evident in the phyllite on the southern shore of Torrent Cove, is apparently absent in the leucocratic gneiss although a parallel mineral lineation was observed within a melanocratic gneiss band in leucocratic gneiss near the north shore of Torrent Cove.

Figure 9. Phyllite (phyllonite) of the Torrent Cove assemblage showing well developed parting parallel to foliation. (GSC 158037).

Figure 10. Pegmatite boudins in the Torrent Cove assemblage. (GSC 158039).

A contact between metagabbro and phyllite of the Torrent Cove assemblage (possibly along a large inclusion within the metagabbro) is well exposed in Canada Bay just south of Green Cove where the metagabbro, represented by amphibolite, dips gently beneath the phyllite. At this point offshoots of amphibolite appear to penetrate the phyllite and fragments of phyllite are apparently engulfed in amphibolite.

Age relations and origin

The Torrent Cove assemblage consists of a variety of lithologies some of which may be part of the Grenville basement (muscovite-chlorite, and quartz-rich schist), some of which were derived from the Grenville basement by cataclasis (phyllite), and some of which could have been intrusive into the basement complex (schistose mesocratic gneiss), although an intrusive contact has not been recognized. Because at least a part and possibly all of the assemblage has been derived from rocks of the Grenville basement, the Torrent Cove assemblage is here considered to be ultimately of the same age as the Grenville basement.

The age of deformation involved in the formation of the phyllite and possible intrusion of the schistose mesocratic gneiss is also conjectural. It is suggested, however, that the phyllite may be considered to be phyllonite developed from Grenville gneiss as a result of cataclasis and metamorphism accompanying diapiric emplacement and spreading of the northern part of the Hooping Harbour megacrystic granitic pluton. This pluton has gently to moderately south-west-dipping contacts where it structurally overlies the Torrent Cove assemblage, and bulges northeastward over it where phyllonite is best developed. Intrusive relations between metagabbro and phyllonite may reflect preservation of the original contact between Grenville gneiss and gabbro related to the anorthosite suite. The low metamorphic grade and finer grain size of the phyllonite in contrast to the gneiss results from the postmetamorphic maximum emplacement of the megacrystic plutons. The minimum age of cataclasis involved in formation of the phyllonite is indicated by a K-Ar muscovite age, 843± 24 Ma (Wanless et al., 1973), from a pegmatite which intrudes the phyllonite.

Fourché Point Schist

The Fourché Point schist is a distinctive grey quartz-eye schist that, within the map area, occurs only in a fault-bounded mass a little more than 1.5 km long and about 0.75 km wide immediately northwest of Fourché Point. To the northwest, the schist lies against southeast-dipping leucocratic gneiss of the basement gneiss complex, and to the southeast it is overlain by massive granite of the Fourché Harbour pluton. Similar rocks may occur within unmapped areas near the coast farther south.

The rock is medium to dark grey and commonly contains transparent quartz crystals up to 1 to 2 mm in diameter in a fine-grained or aphanitic matrix. Locally pale brownish white feldspar crystals with indistinct boundaries give the rocks a prophyritic appearance. At one locality fine-grained quartzofeldspathic lenses are separated by thin anastamosing, whispy stringers of fine-grained chlorite-biotite. Although the rock is schistose, the orientation of schistosity on a larger scale is difficult to detect because of the hackly fracture and locally dense texture of the rock. The schist is cut by numerous white quartzofeldspathic veins and by some pink pegmatite, the former dipping moderately to steeply southeast.

In thin section the rock appears strongly schistose and fine to medium grained. Quartz crystals, and locally amphibole crystals up to 2 mm long, are present in a fine-grained matrix. The principal mineral components in order of decreasing abundance are albite and thoroughly altered plagioclase, quartz, pale green amphibole or biotite or both and epidote, and chlorite. Muscovite is an important constituent in places and minor potash feldspar is present locally. Accessory minerals are sphene, carbonate, apatite, magnetite, zircon, allanite, and fluorite.

Textural evidence of cataclasis is prominent. Quartz grains are commonly composed of several smaller crystals with serrated margins and moderate to extreme irregular extinction due to strain. Pockets of mortar texture are present locally. Disaggregated quartz crystals may be strung out into lenses parallel to schistosity indicated by amphibole and muscovite alignment. Fine-grained biotite on the other hand typically is not oriented parallel to the major schistosity.

The rock is thoroughly broken up slightly less than 1.5 km north of Fourché Point where massive leucocratic gneiss and Fourché Point schist are in contact along the slope. Pink and white pegmatite, grey fine-grained quartz-bearing gneiss, epidote-rich gneiss, and chloritic leucocratic gneiss are present in patches with irregular schistosity. Talus is predominantly grey schist similar to that farther south indicating that the Fourché Point schist is present in the cliffs above the contact zone at this locality. In the small cove immediately north of Fourché Point the schist is separated from brecciated massive granite by a steeply south-dipping fault. On the north shore of Fourché Harbour pale greenish grey schist resembling the Fourché Point schist is present at one locality, but it appears that the Fourché Point schist is largely cut out by convergence of the northwest fault margin of the schist and the fault along the granite contact.

Age relations and origin

The Fourché Point schist lies within the Grenville basement and has most probably been derived from rocks of that basement by cataclasis. The schist is therefore a variety of metamorphic rock ultimately of the same age as the Grenville basement.

Cataclasis involved in formation of the Fourché Point schist is probably of Grenville age because the schist is intruded by pegmatite. Muscovite in a pegmatite intrusive into rocks of the Torrent Cove assemblage to the north has been dated at 843± 24 Ma (Wanless et al., 1973) and no Paleozoic intrusions have been established within the Precambrian inlier (although Neale and Nash, 1963, have suggested the presence of a Paleozoic pluton cutting Precambrian rocks in the southern part of the inlier). The most probable age of the pegmatite intrusive into the Fourché Point schist is therefore Grenvillian. Post cataclastic fine-grained biotite within the Fourché Point schist resembles biotite that has been dated from the Grenville basement immediately to the north. The K-Ar age of this biotite, 434± 18 Ma (Wanless et al., 1973), provides an independent minimum limit for the age of the cataclasis in the Fourché Point schist.

Felsitic cataclastic rocks

Largely massive rocks of felsitic appearance are present in a fault wedge at the coast below Sugarloaf hill, in a small brecciated lens along the fault on the highlands above, and in several small but poorly known bodies exposed within the Precambrian inlier along its northeastern margin between Sugarloaf hill and the cirque-like valley north of Cloud River. The best exposed felsitic body occurs at the coast near Sugarloaf hill where a wedge in excess of 30 m thick is present. The body north of Cloud River was seen at two localities where it appeared to be about 9 m thick and tabular in form. Other occurrences are exposed in isolated outcrops and hence are of unknown form and size.

In hand specimen the felsitic rocks are fine-grained to aphanitic, pink to grey, and largely massive although a weakly developed flow texture is apparent locally. Quartz crystals, commonly 0.5 mm and locally 2 mm in diameter, are visible in some bodies.

Thin section examination of the rocks near Sugarloaf hill suggests that they are protomylonites (classification of Higgins, 1971). They consist of scattered small fragments of leucocratic gneiss in an abundant quartzofeldspathic matrix containing pockets, veins and tiny shears of very fine grained felsic minerals. Locally classic mortar texture is developed in which fragments of leucocratic gneiss forming 75 per cent of the rock are surrounded by similar fine-grained felsic minerals. Quartz crystals have extreme undulatory extinction and are serrated about their margins. Potash feldspar fragments show coarse microcline twinning characteristic of this mineral in the leucocratic gneiss. A little muscovite is present but mafic minerals have been destroyed except for patches of magnetite, and hematite along grain boundaries.

When examined in thin section the felsitic rocks northwest of Sugarloaf hill and within the basement complex comprise quartz crystals and fragments of medium-grained quartz-feldspar rock in a massive, equant, fine-grained (0.05 to 0.5 mm) matrix which typically forms more than 50 per cent of the rock. Patches of muscovite, commonly adjacent to rock fragments, poikilitically enclose matrix grains. Fine-grained greenish biotite is present in one specimen. Accessory minerals are magnetite, apatite, zircon, sphene, and fluorite, but all of these do not occur together. Quartz crystals are equant and in places consist of several crystals with serrated contacts. Grain boundaries with the matrix are serrated and embayments typical of volcanic quartz are not present. Extinction of quartz crystals is undulatory, but extreme strain shown by quartz in the rocks near Sugarloaf hill is not evident. Feldspar, which is in included rock fragments, shows abundant but patchy microcline twinning. No euhedral or subhedral crystals are evident.

The felsitic rocks at Sugarloaf hill (protomylonite) at their western contact overlie a lens of Bradore sandstone-conglomerate some 5 m or more thick, which in turn overlies leucocratic gneiss. The sandstone wedge strikes 05° and dips 35° east. At its basal contact, pockets of conglomerate and lenses of greenish siltstone persist on the east-dipping upper surface of the leucocratic gneiss, the contact being unconformable rather than faulted. The contact between Bradore sandstone and the overlying felsitic rocks, though not exposed, appears to dip east at about the same angle. The upper fault contact of the felsite-like rock is located in a small sharply defined cove where felsitic rock underlies mesocratic to melanocratic gneiss resembling that on the north shore of Hooping Harbour. These rocks in turn are in fault contact with grey schist of the Hare Bay Allochthon (Sugarloaf schist, Williams and Smyth, this volume) about 0.5 km notheast.

Age relations and origin

The felsitic rocks (protomylonite) at Sugarloaf hill are cataclastic. Those within gneiss to the northwest differ principally in having larger proportions of a more homogeneous matrix. Both are devoid of megacrysts of likely volcanic origin, and it seems probable that both are of similar cataclastic origin; however the northwestern bodies may represent a more advanced stage of cataclasis and recrystallization than those at Sugarloaf hill. Three of the felsitic bodies immediately northwest of Sugarloaf hill are approximately colinear and could represent a single zone of cataclasis.

The protolith from which the felsitic rocks were derived was the Grenville basement, as fragments of this basement in advanced stages of cataclastic digestion are present within the felsitic rocks and included material from younger formations has not been recognized. Furthermore the felsitic rocks lie predominantly within the Grenville basement. The felsitic rocks are therefore considered to be a variety of metamorphic rock ultimately of the same age as the Grenville basement.

The probable common origin and lesser degree of schistosity and recrystallization shown by the felsitic rocks in comparison to the Fourché Point schist suggest that cataclasis involved in the formation of these felsitic rocks is post-Grenvillian. To some extent this hypothesis is supported by the superposition of protomylonite on top of the Bradore Formation, for if cataclasis is of Grenvillian age two distinct periods of faulting would be required to achieve the present disposition of felsic rocks. Felsitic rock within the Hooping Harbour pluton has developed late micas and these may have formed as a result of late greenschist facies metamorphism. Because this metamorphism may be present in the Devils Cove Formation at Wild Cove, the younger limiting age for this metamorphism may be taken from the K-Ar biotite age of 434 to 512 Ma derived from secondary biotite in gneiss in the Fourché Harbour area. Cataclasis in the felsitic rocks is thus probably Ordovician or earlier, but post-Grenvillan.

Intrusions of the anorthosite suite

Intrusions of the anorthosite suite, which are widespread in the Grenville Province, include "rocks of intermediate to silicic composition variously referred to as mangerites, charnockites, adamellites, syenites and quartz syenites" in addition to the anorthositic rocks (Emslie, 1973). Hale (1961), mapping in the Grenville Province immediately west of the present map area, recognized that, although not everywhere found together, gabbro and anorthosite are commonly associated and that the smaller the intrusive mass the greater the proportion of gabbro likely to be present. Thus the occurrence of many small gabbroic bodies without extensive anorthosite is nevertheless consistent with anorthositic plutonism. The association of mangerite and hornblende granite showing some chemical characteristics of the anorthosite suite, with these gabbro bodies within the map area is further indicative of this type of plutonism. The metagabbro, mangerite, and hornblende granite plutons are therefore considered in one group although the metagabbro may include some undifferentiated rocks of different age and origin, and other intrusive bodies (namely dioritic rocks of the Long Range and foliated granitic rocks of Labrador and Quebec) may have been intruded during the interval over which anorthositic plutonism took place.

Emplacement of intrusions of the anorthosite suite within the map area is thought to have occurred in succession with the gabbro bodies followed by mangerites and ultimately by the hornblende granite plutons. Further, there is some indication (to be discussed in relation to metamorphic and structural geology) that regional folding along northeast-trending axes, which accompanied the metamorphic maximum, occurred after emplacement of the metagabbro, perhaps about the same time as the emplacement of the mangerite plutons.

Metagabbro and related rocks

Metagabbro intrusions have been found in all regions of the map area. They include amphibolite (hornblende and plagioclase commonly in about equal proportions), hypersthene amphibolite, norite, mangeronorite, metatroctolite, and meta-ultramafite. Some amphibolite bodies included in this unit (mostly small) may be ultimately of sedimentary or volcano-sedimentary origin and as such may be more correctly termed para-amphibolite. It is furthermore possible that bodies of more than one albeit early period of intrusion are involved. Nevertheless many if not most of the metagabbro bodies (including the Red Bay metagabbro with its anorthositic phase) probably form part of the anorthosite suite of intrusions. The metagabbros as defined above underlie about 40 km^2, of which the largest body, that at Red Bay, accounts for about one third. Many bodies are tabular or lenticular in shape and less than 200 m thick, and therefore, many are too small to be shown at the present scale of mapping.

Some metagabbro bodies produce prominent dark rounded ridges such as that south of Booney Lake but more commonly they have no distinctive topographic expression. The rocks may be schistose or massive and locally, compositional banding is preserved. The body at Red Bay varies in composition from norite and mangeronorite to anorthositic metagabbro; the possibility of gross layering was identified from the air, although none was recognized on the ground. Elsewhere remnants of layering (15 to 30 cm bands of tremolite-chlorite-carbonate rock in metagabbro) have been observed in the intrusion at Canada Bay. An ultramafic zone, which pinches and swells up to about 12 m thick, was found in the metagabbro body south of upper Cloud River. A small, poorly exposed, possibly dyke-like body perhaps 5 m across with bands of differing mineral composition along its margins was found some 19 km west of Canada Bay and 6 km south of Cloud River. Between Red Bay and Pinware several thin but complex bands (gabbro dykes?) parallel to surrounding gneissosity were found in which a layer of feldspathic lenses lay along one side of a band (Fig. 11).

The metagabbro is chiefly medium grained (1 to 3 mm), but about 10 per cent of specimens examined are finer grained and a further 10 per cent are coarser. The principal minerals present are hornblende and plagioclase, however, biotite is also common. Quartz, a minor constituent of one quarter of the metagabbro examined, is most common in metagabbro from Labrador and Quebec. Quartz typically occurs as small anhedra along grain boundaries or is poikilitically enclosed in hornblende. Plagioclase is present as equant grains. Locally relict ophitic texture is evident but zoning is not preserved. The most common plagioclase in all terranes is andesine although labradorite was found locally. In the greenschist terrane plagioclase is more or less saussuritized and albite or sodic oligoclase is widespread. Potash feldspar, commonly perthitic microcline, is found in coarse antiperthitic and mesoperthitic intergrowths in mangeronorites from the granulite terrane (Fig. 12, 13). It also occurs in some metagabbro bodies engulfed in granitic rocks in the Long Range. Hornblende with pleochroic formula *x* ochre, *y* brown-green, *z* blue-green, grey-green, or pale green is typical in the greenschist terrane. Elsewhere *z* is typically green to olive-brown and may be blue-green particularly along crystal margins. Biotite is olive to greenish in the greenschist terrane but is commonly red-brown elsewhere. In many thin sections biotite and magnetite are associated in patches. Cummingtonite, commonly showing polysynthetic twins, is present in places within hornblende in the amphibolite terrane. Colourless amphibole showing parallel extinction (probably anthophyllite) in clots rimmed by hornblende is present in two lenticular gabbro plugs northwest of Henley Harbour (Fig. 14), and anthophyllite with pleochroic formula *x* colourless, *y* = *z* pale red-brown was observed in a single specimen from the Long Range.

Hypersthene in metagabbro of the amphibolite terrane occurs as anhedral remmants (Fig. 15) that are locally poikilitically enclosed in hornblende or biotite. More rarely it occurs as reaction rims about olivine which are in turn surrounded by hornblende (Fig. 16). In the granulite terrane hypersthene exists as discrete crystals typically accompanied by clinopyroxene. It is found rarely as remnants in hornblende along the western margin of the greenschist terrane. An apparently large, but poorly exposed body of metagabbro near the south-central limit of mapping in the northern Long Range contains relatively massive hornblende-rich, hypersthene-bearing metagabbro in its central part, but more schistose and biotite-rich debris scattered about the northern periphery of the body contains no hypersthene. The latter lithology may therefore represent a late alteration ring within which higher grade mineral assemblages are preserved. Abundant olivine was found in a small body of pyroxene-olivine-plagioclase rock (metatroctolite) and as remnants in a more typical hypersthene-bearing metagabbro. Garnet is present in a lens of dark green garnetiferous amphibolite within the metagabbro at Canada Bay and in four of ten specimens examined from the amphibolite terrane in Labrador and Quebec (Fig. 17). It was not found in gabbroic rocks of the granulite terrane or in those of the amphibolite terrane in the Long Range. Chlorite and epidote are common alteration products in the greenschist terrane. Prehnite (?) and muscovite, although more widespread, are rare.

Accessory minerals include apatite, magnetite, sphene, allanite, carbonate, and chabazite. Apatite is found in almost all metagabbro and magnetite is typically present but is less common in matagabbro from the greenschist terrane. Carbonate is uncommon but widespread and may be most common in metagabbro from the greenschist terrane.

In the meta-ultramafic rocks olivine is present in variable amounts along with one or more of anthophyllite, hypersthene, tremolite, chlorite, serpentine, phlogopite, talc, spinel, chromite, carbonate, apatite, and magnetite. Olivine is commonly interlaced with serpentine. A small dyke-like meta-ultramafic body south of Cloud River contains a core zone up to 2 to 3 m wide of coarse-grained phlogopite-rich rock in which phlogopite crystals are present in a matrix of finer grained olivine, chlorite, spinel, and serpentine with minor chromite and apatite. Coarse spinel in this rock is illustrated in Figure 18. This is bordered by olivine-hypersthene-tremolite and olivine-anthophyllite-chlorite bands of unknown width.

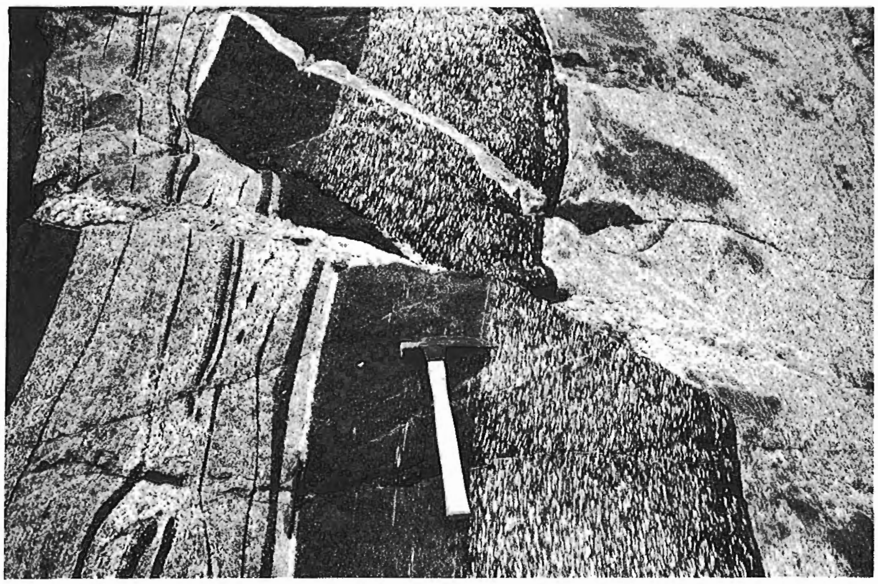
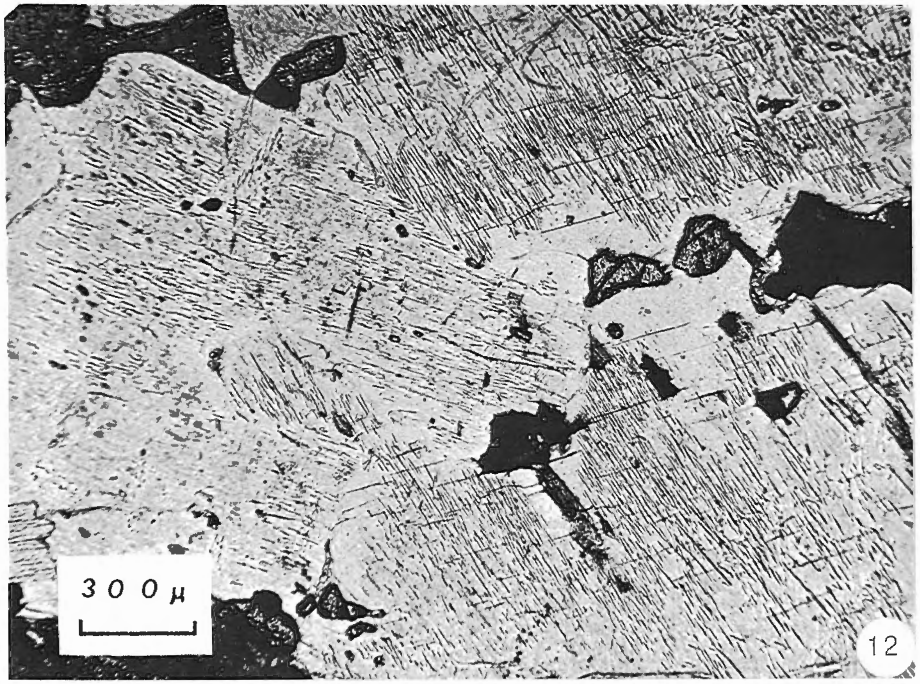
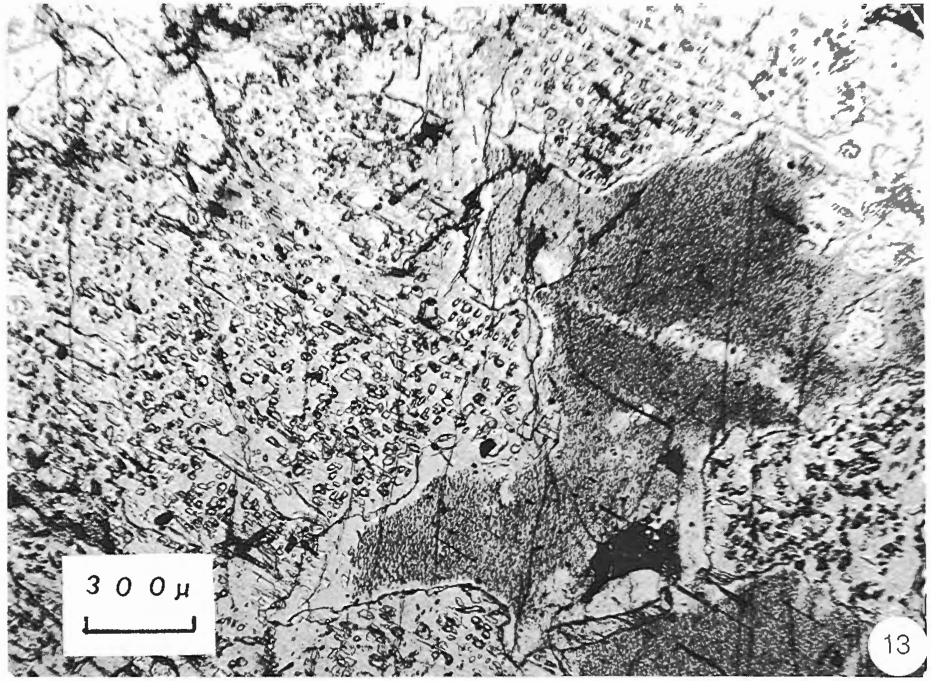


Figure 11. A metagabbro (?) band between Red Bay and Pinware. White feldspathic lenses in the part of the band on the right may be deformed or corroded phenocrysts or glomerocrysts. (GSC 160095).





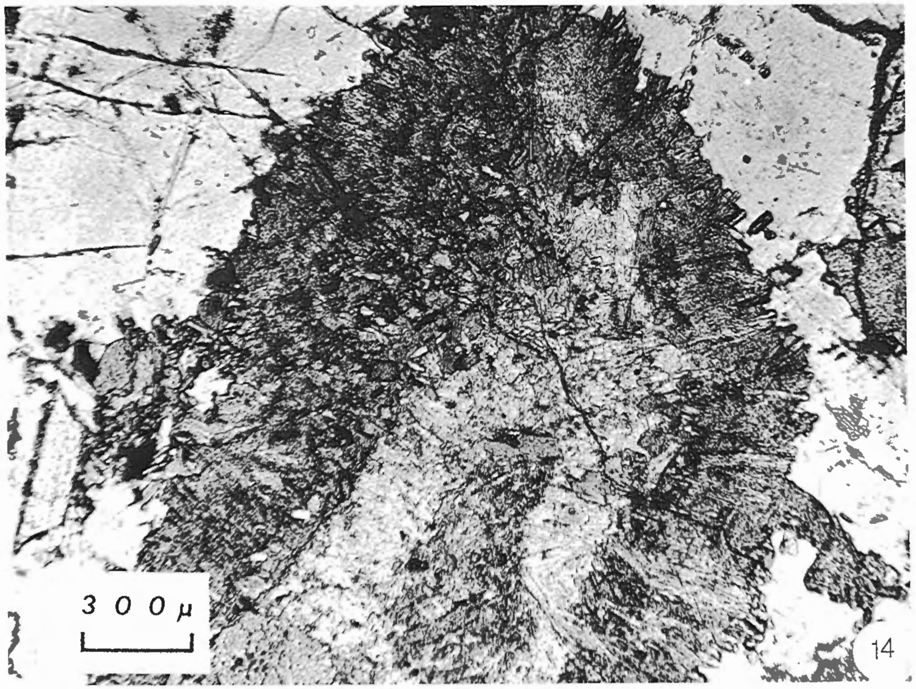
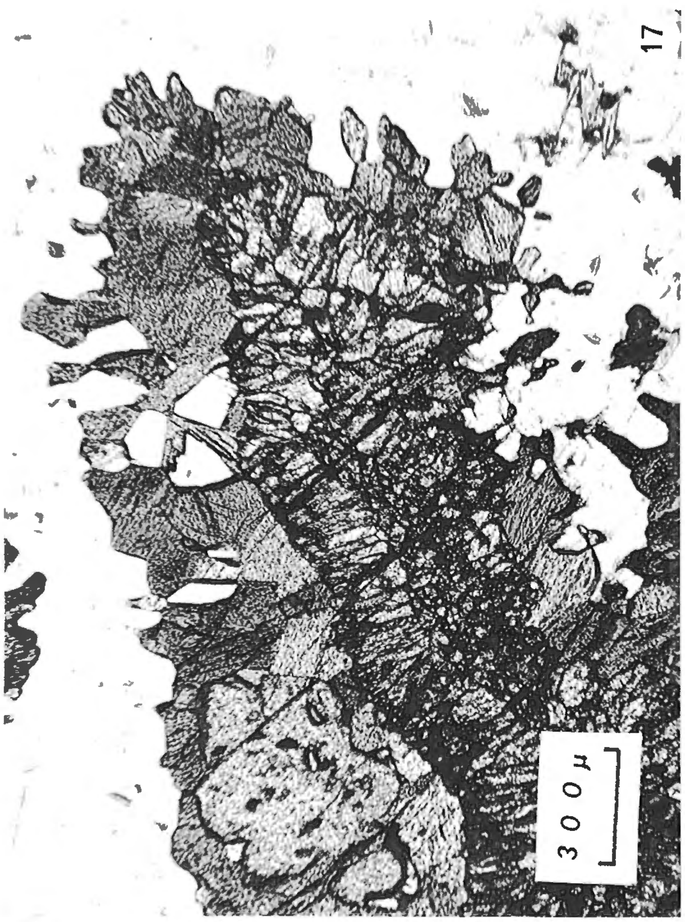


Figure 12. Antiperthite in mangeronorite from Red Bay. Polarized light (GSC 201902-H).

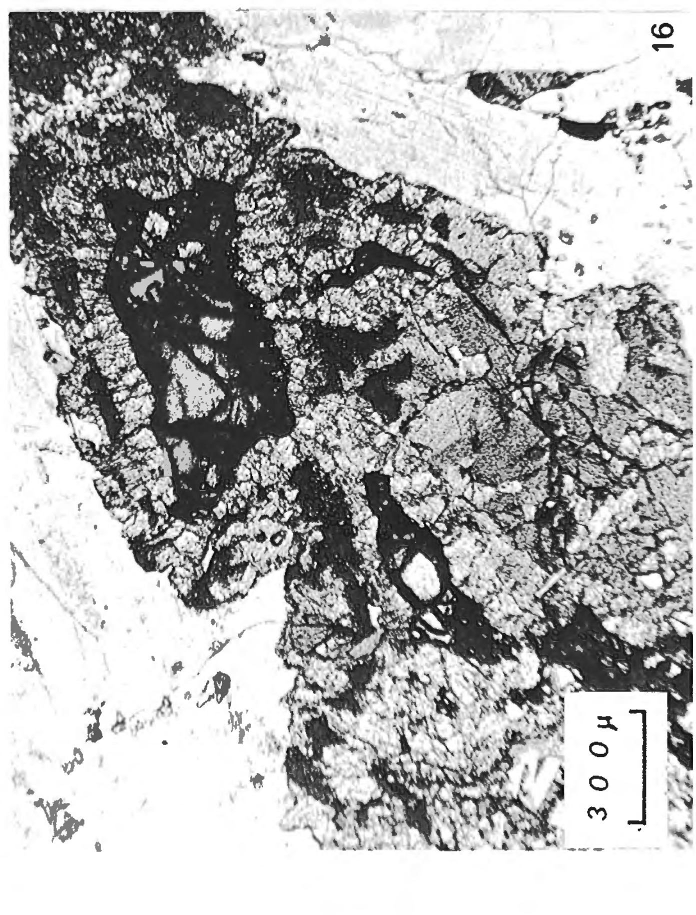
Figure 13. Mesoperthite and antiperthite in mangeronorite near West St. Modeste. (GSC 201902-E).

Figure 14. Hornblende rims anthophyllilte in metagabbro plug west of Henley Harbour. Plane polarized light (GSC 201902-O).


17
300 μ


18
300 μ


15
1 mm


16
300 μ

Age and origin. Although most of the metagabbro bodies form elongate, lens-like masses within the basement gneiss and are locally schistose parallel to the regional foliation, a few, such as the dyke-like body near Northwest Brook, cross-cut the foliation in the gneiss at a small angle. These local cross-cutting relationships and the association of ultramafic bands in some metagabbro bodies indicate that many if not most are probably intrusive bodies. On the other hand, among the smaller basic bodies there appears to be a gradation between amphibolite of intrusive origin (metagabbro) and amphibolite which forms an intimate part of local gneiss sequences. The latter are strongly schistose and thin (mostly 9 m or less thick) but in gross mineralogical character most are not distinctive. A few however have unusally high biotite and quartz content and are transitional to melanocratic gneiss. Clifford and Baird (1962) have suggested that amphibolite of both intrusive and sedimentary origin exist in the Long Range. If those of the latter origin were partly derived from tuffs and flows associated with early basic intrusions, then the apparent gradation from gneiss to amphibolite might be explained.

The age of the metagabbro is therefore complex. A few of the smaller amphibolite bodies are likely of the same age as the sediments from which the enclosing gneiss was formed, whereas others intruded the gneiss. In the granulite terrane most of the metagabbro contain mineral assemblages characteristic of granulite facies and therefore were probably emplaced before culumination of high-grade metamorphism. In the Long Range the local preservation of hypersthene suggests that their age relations are probably the same.

Age relations between the metagabbro and the various granitic plutons of Labrador and Quebec are not directly known because few contacts have been found between them. An apparently lenticular body of noritic(?) metagabbro is present within the hornblende granite intrusion north of L'Anse-au-Loup, but the granitic is schistose in the vicinity of this metagabbro and it appears that the granitic magma incorporated a body of older metagabbro. No metagabbro is known within the mangerite plutons although metagabbro is present close to the margins of the large body on Pinware River. Because the mangerite bodies appear less altered than the metagabbro it is perhaps more likely that they are the younger.

No metagabbro bodies clearly intrude the massive or megacrystic granitic plutons, but numerous irregular bodies of metagabbro exist within the satellite pluton southwest of the Hooping Harbour pluton (Fig. 21), and in places this pluton is unusually mafic. These relationships are thought to result from contamination of the granitic rocks through incorporation of metagabbro prior to final emplacement of the former. Similar relations are present on a more local scale within the Cloud River pluton.

Figure 15. Remnant anhedral hypersthene in metagabbro from the amphibolite terrane, northern Long Range. Polarized light (GSC 201902-ZZ).

Figure 16. Olivine with successive rims of hypersthene and amphibole in metagabbro from the amphibolite terrane, northern Long Range. Polarized light (GSC 201902-V).

Figure 17. Garnet within an amphibole rim about hypersthene in metagabbro from the amphibolite terrane at Lourdes du Blanc Sablon. (GSC 201902-R).

Figure 18. Large crystals of olivine and spinel from a minor metaultramafic body south of Cloud River. Polarized light (GSC 201902-C).

The metagabbro bodies thus appear to be the earliest intrusions within the basement gneiss complex. In large part they are thought to have been intruded contemporaneously with anorthosite plutons present within the Grenville Province to the north, west, and south. They were probably emplaced before or during high-grade metamorphism and early northeast-trending folding (for more detailed discussion see section on Structural Geology), and before the other plutons of the anorthosite suite, mangerite and hornblende granite. Correlation of the metagabbro with anorthosite plutonism suggests that they were emplaced about 1460 Ma (Emslie et al., 1976).

Mangerite

Three plutons and one small lenticular body of mangerite aggregating about 65 km^2 are known within the area mapped in Labrador and Quebec. Other small bodies may be present in the granulite terrane but were not detected.

Mangerite outcrops are characterized by a brown to ochre stain that penetrates the surface deeply. Fresh surfaces, which are difficult to obtain, are deep olive-green making mafic percentage difficult to estimate. Quartz is rarely recognizable. The predominant constituent is coarse-grained untwinned feldspar.

The mangerite is medium to coarse grained with anhedral mesoperthitic alkali feldspar, commonly 3 to 5 mm in diameter, accounting for about 75 per cent of the rock. Mesoperthite is composed of fine, homogeneously distributed lenses of potash feldspar and sodic plagioclase in approximately equal proportion (Fig. 19), except at crystal margins where the potash phase predominates. The larger mesoperthitic alkali feldspar crystals are commonly surrounded by smaller oligoclase crystals that may be abundantly myrmekitic and are commonly zoned to albite when in contact with the mesoperthitic feldspar. Quartz, if present, may constitute up to about 5 per cent of the rock and commonly accompanies oligoclase in the interstices between alkali feldspar crystals. Mafic minerals, which comprise up to about 15 per cent of the rock, include in order of decreasing abundance, very pale green pyroxene (clinopyroxene with or without orthopyroxene), hornblende, dark opaque minerals, and locally biotite. Accessory minerals include zircon, apatite, carbonate, and rarely sphene and allanite.

Age relations and origin. The mangerite bodies are probably intrusive into the basement gneiss which wraps around them, but contacts or cross-cutting relationships were not observed. The possibility that the mangerite bodies might be unconformably older than the basement gneiss appears remote because (i) in contrast to the basement gneiss, the mangerite bodies are only locally foliated in spite of the relatively small size of some of these plutons, and (ii) mineral textures in the mangerites are relatively simple in comparison to those found in the gneiss, suggesting that it is the mangerites that have had the simpler metamorphic history. Age relationships between the mangerite and metagabbro suite are not directly known, but because the metagabbro appears to be the more altered of the two it is likely the older. Contacts between mangerite and hornblende granite were not found.

Hornblende granite to granodiorite

Hornblende granite intrusive rocks are confined to the granulite terrane of Labrador and Quebec. One large pluton and two smaller bodies lie wholly within the map area, and a second major body is present in Quebec at the northwest margin of the map area. They cover about 1000 km^2 within the map area.

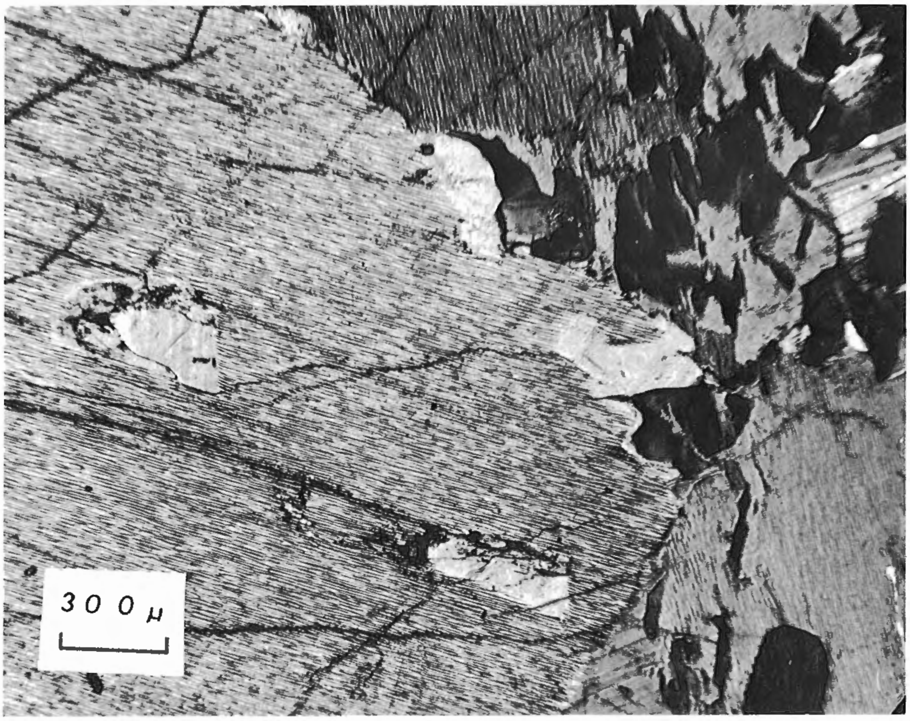


Figure 19. Mangerite from the large body northeast of Pinware showing fine, regular mesoperthitic intergrowth in alkali feldspar. Crossed nicols (GSC 201901-B).

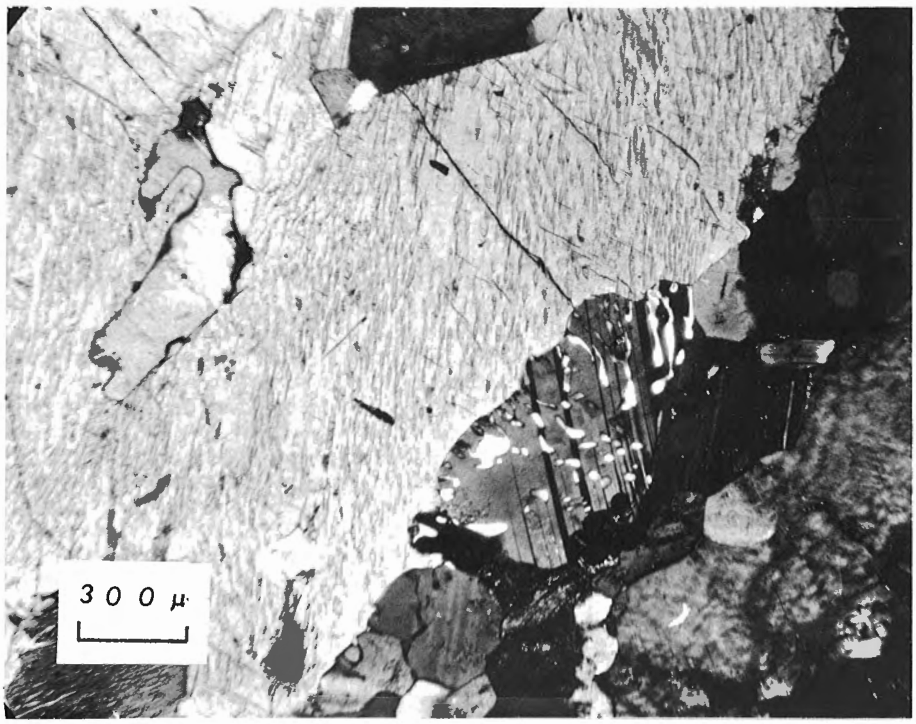


Figure 20. Submesoperthite and myrmekite in hornblende granite. Crossed nicols (GSC 201902-E).

The hornblende granite is typically massive, medium to coarse grained, pink to olive-brown, with grey quartz, twinned and untwinned feldspar, and a variable proportion of mafic minerals, mostly hornblende and biotite. Two varieties, leucocratic granite (granite to quartz monzonite) and granodiorite (quartz monzonite to granodiorite) have been mapped with a dividing line for mapping purposes at roughly 10 per cent of mafic minerals. These varieties are distinctive in the north and southeastern part of the large pluton, but in the western parts the rocks at many sites have close to 10 per cent of mafic minerals and a division is difficult. Steeply dipping schistosity is present along the north and east margins of the large intrusion and is parallel to schistosity in the country rocks. At the southeast corner of the large pluton, schistosity is particularly well developed in the vicinity of a large metagabbro inclusion and is parallel to the northward striking contact of the pluton. In the northeast however, schistosity in both gneiss and plutonic rock intersects the contact at a high angle. In the northwestern part of the pluton the rock, though massive, has a conspicuous subhorizontal sheeting that is emphasized along bluff faces by weathering. Fine-grained massive granitic rocks resembling leucocratic gneiss and a large remnant of quartz-rich gneiss are present near the north margin of the pluton, and several large mesocratic gneiss inclusions were found near the south margin. Small inclusions (of less than outcrop size) are rare.

The hornblende granite is coarse grained (3 to 20 mm), the leucocratic granite is slightly more coarsely crystalline than the granodiorite. In the leucocratic granite, quartz is commonly in excess of about 25 per cent, plagioclase is albite or sodic oligoclase, and submesoperthitic microcline exceeds plagioclase in abundance. Biotite, muscovite, chlorite, and epidote are more common in the leucocratic granite, and hornblende is more common in the granodiorite. In contrast to the leucocratic granite, the granodiorite mostly has less than 25 per cent quartz, and a plagioclase (typically calcic oligoclase) that exceeds submesoperthitic microcline in abundance. Accessory minerals are zircon, apatite, dark opaque minerals, sphene, and carbonate. Submesoperthitic microcline (Fig. 20), which is found in both leucocratic granite and granodiorite, is characterized by coarse lenses and 'blebs' of albite distributed in irregular, commonly unsymmetrical patches within the host microcline. The two feldspar phases may be in nearly equal proportions locally, but most commonly microcline appears to be slightly predominant. Hypersthene was not observed even in the most melanocratic granodiorite, but clinopyroxene, commonly overgrown by hornblende, is present in some specimens. Sphene commonly rims magnetite-ilmenite. Quartz occurs as large independent grains although myrmekite is also conspicuous (Fig. 20).

Age relations and origin. Although the hornblende granite has not been found in contact with the mangerite and the age relations are therefore not directly known, they appear to be related because both are restricted to the granulite terrane and both have features suggestive of high-grade metamorphism. On the other hand the mangerite bodies, which commonly contain both clino- and orthopyroxene as well as mesoperthite, were clearly dry intrusions, whereas the hornblende granite, which contains only local clino- pyroxene and coarse irregular submesoperthitic intergrowths, may have been emplaced under slightly less severe (wetter?) conditions. It seems unlikely that gneiss and mangerite could have been raised to conditions of granulite metamorphic facies subsequent to emplacement of the hornblende granite, in a manner such as to almost completely surround the major granitic pluton without producing either hypersthene or mesoperthite (as opposed to submesoperthite) within it. Rather, it seems probable that the gneiss was first raised to granulite facies perhaps during emplacement of the mangerite bodies and that the granitic rocks that followed were emplaced under conditions only slightly less severe, so that the mineralogy of the surrounding gneiss was not severely altered.

Chemical Analyses. Two samples (4.5 kg each) of (pyroxene)-biotite-hornblende granodiorite from the hornblende granite body were collected for rapid method chemical analysis (Table 4). These samples represent normal (sample 470 BK69) and particularly mafic, pyroxene-bearing (sample 471 BK69) parts of the granodiorite phase of the intrusion. The analyses are unusual and resemble granogabbro (Johannsen, 1932) in having low SiO_2 and high FeO (total) to MgO ratios but the plagioclase in the granodiorite is oligoclase and atypical of granogabbro. These features are typical of the chemistry of the anorthosite suite of intrusions outlined by Emslie (1973).

Foliated Granitic rocks

Several small bodies of schistose to banded or massive granitic rocks, chiefly of granodioritic composition and commonly bearing augen of potash feldspar, are present within the map area north of the Strait of Belle Isle. They have been mapped along the coast, but their distribution and extent inland are uncertain due to the reconnaissance nature of the inland mapping. The main bodies north of the Strait from west to east are at Isle de Bassin, Blanc Sablon, West St. Modeste, East St. Modeste, Wreck Bay, and York Point. These foliated granitic rocks are varying shades of pink or grey. They are banded, lenticularly foliated, simply schistose, or locally massive. In places potash feldspar augen or megacrysts are conspicuous, and these may be present throughout or may appear in some bands only.

The foliated granitic rocks are chiefly granodiorite but some quartz monzonite and granite are also present. Quartz, plagioclase (albite to sodic andesine), microcline, and biotite are characteristic. Hornblende (x ochre, y brown-green, z blue-green) is present in most specimens. Chlorite, muscovite, and epidote are present in some. Accessory minerals in order of decreasing abundance are magnetite, sphene, apatite, zircon, allanite, carbonate, and pyrite. Sphene crystals up to 3 mm are present locally.

In most of the foliated granite bodies microcline is slightly perthitic, but the granite-granodiorite near West St. Modeste contains microcline with coarse inhomogeneous irregular submesoperthitic intergrowths. Some crystals have non-perthitic rims. Myrmekitic quartz-plagioclase intergrowths are common.

Schistose granodiorite on the Isle de Bassin contains lenses and bands of amphibolite, pegmatite, and dykes and patches of pink granite. Similar rocks were found along strike on the islands farther north, on the mainland south of the prominent fault scarp that forms the shoreline farther west, and at one locality 3 km north of Bradore Bay at the east margin of the map area. Schistosity at all outcrops visited dips moderately to steeply east and is roughly parallel to that in the gneiss to the east and north. At Blanc Sablon, pink schistose quartz monzonite in which the schistosity is weak and contorted is exposed near the shoreline.

Near West St. Modeste for about 8 km along the coast, schistose granite to granodiorite contains many lenses and bands of gneiss marked by fairly sharp contacts. The granitic rocks may be divided (as interpreted on the enclosed map) into two bodies by a belt of gneiss that passes through the townsite. Schistosity and gneissosity in the foliated granitic rocks strikes easterly and dips moderately south from Pinware to a point about 2.5 km south of West St. Modeste. There the earlier east-striking schistosity is clearly interrupted by zones of superimposed schistosity that strike northerly and dip moderately west. Farther west the older schistosity is not evident and the later trend continues northward across the earlier trend and along the contact of the hornblende granite pluton where it appears both within the margin of the pluton and in the country rock. Linear features are rare in both schistosity domains, but a few measurements of crenulation suggest that southwest-plunging lineations near West St. Modest are replaced by more southerly plunging lineations in the rocks near Capstan Island. The contact between gneiss and schistose granite near Capstan Island is gradational over a restricted zone.

East of Pinware River along the coast for about 5 km, bands of schistose augen granodiorite are interlayered with gneiss striking east-northeast and dipping steeply south. East

Table 4. Chemical analyses of the granodiorite phase of the hornblende-bearing intrusive granitic rocks*

Sample Number 470 (BK69)				Sample Number 471 (BK69)			
Oxide	Wt%	Norm Mineral	Wt%	Oxide	Wt%	Norm Mineral	Wt%
SiO_2	53.7	Quartz	8.91	SiO_2	57.2	Quartz	13.05
TiO_2	1.84	Corundum	-	TiO_2	2.19	Corundum	0.36
Al_2O_3	15.1	Orthoclase	21.54	Al_2O_3	15.0	Orthoclase	26.21
Fe_2O_3	6.1	Albite	29.09	Fe_2O_3	5.9	Albite	26.41
FeO	5.2	Anorthite	15.47	FeO	3.8	Anorthite	13.13
MgO	2.4	Diopside	0.89	MgO	2.5	Enstatite	6.27
CaO	5.6	Hedenbergite	0.22	CaO	4.0	Magnetite	6.50
Na_2O	3.4	Enstatite	5.67	Na_2O	3.1	Ilmenite	4.19
K_2O	3.6	Forsterite	1.78	K_2O	4.4	Hematite	1.56
H_2O	0.9	Magnetite	8.95	H_2O	0.8	Apatite	2.43
MnO	0.22	Ilmenite	3.53	MnO	0.17		
P_2O_5	1.71	Apatite	4.01	P_2O_5	1.04		
CO_2	<0.1			CO_2	<0.1		
Total	99.8		100.06	Total	100.1		100.11
Fe total as Fe_2O_3	11.9			Fe total as Fe_2O_3	10.1		
Fe total as FeO	10.7			Fe total as FeO	9.1		
Sr	0.063			Sr	0.054		
Ba	0.22			Ba	0.21		
Zr	0.40			Zr	0.16		
Ce	0.085			Ce	0.060		
La	0.046			La	0.017		
Yb	0.002			Yb	0.00056		
Cu	0.0011			Cu	0.0025		
Pb	0.0020			Pb	0.0024		
Zn	0.026			Zn	0.018		
Sn	0.00054			Sn	0.00041		

*Chemical analyses by the rapid method staff, and by the spectrochemical laboratory staff of the Geological Survey of Canada.
Samples 470(BK69) and 471(BK69) represent the south and southeast margin of the main hornblende granite intrusion.

Approximate modal composition

Sample Number	470	471	Sample Number	470	471
Quartz	<15	>15	Chlorite	tr	-
Microcline (Submesoperthite)	>30	>30	Clinozoisite	R	-
Oligoclase	<30	<30	Magnetite	m	m
Hornblende	M	M	Apatite	tr	tr
Biotite	m	tr	Sphene	R	tr
Clinopyroxene	tr	-	Zircon	R	R

M major component; m minor component; tr trace; R rare.

of this hybrid zone more massive granodiorite containing melanocratic zones of granodiorite or quartz diorite is present. A second smaller massive granodiorite body is present within the gneiss about 1.5 km farther east. The location of contacts between normal gneiss and the eastern granodiorite is known within 2 m, but that between schistose and more massive granodiorite appears gradational. It is possible that two ages of granodiorite are present, the older being schistose and the younger nearly massive. The latter rocks are therefore tentatively included in unit Hgd. This is based on the absence of submesoperthite in the massive granitic rocks and its sporadic preservation in the surrounding gneiss.

A partly schistose body of pink leucocratic granodiorite extends along the coast for 3 km west of Wreck Bay. At its western contact this body displays a strong southwesterly plunging mineral lineation parallel to fold axes in the gneiss farther west. A similar mineral lineation was found in a gneiss inclusion at one locality within the pluton about 1.5 km to the east. The gneiss and granodiorite are interlensed at the contact which strikes northerly and dips 60° east toward the pluton. At its eastern margin the granodiorite contains numerous inclusions; the contact was not precisely located, but gneissosity in the marginal gneiss dips 13° east and exhibits isoclinal mesoscopic folds with axial plane schistosity cutting across fold crests. Fold axes plunge northeastward at 10°, approximately parallel to fold axes and crenulation in the gneiss farther east (but at a high angle to linear features in the granodiorite and gneiss farther west). Inland the contacts of the body are not known. These field relations suggest that folding of the gneiss east of the pluton predated its emplacement whereas deformation of the western margin of the pluton and the adjacent gneiss at least partly postdated its emplacement.

A schistose, augen granodiorite showing strong mineral lineation extends along the coast for about 6 km southwest of York Point (near Henley Harbour). At the southwest contact of this body phyllitic rocks resembling those of the Torrent Cove assemblage dip gently southwest away from the pluton. About 200 m northeast within the pluton a layer of gneiss appears to rest on top of the granodiorite. Both the granodiorite and the surrounding gneiss display strong planar and linear fabrics indicating that pluton and gneiss have been folded together.

Age and origin. The age and origin of the foliated granite bodies is poorly known because contacts in the inland areas have not been examined and deformation has obscured most of the primary structures exposed along the coast. Contact relations along the eastern margin of the pluton at Wreck Bay suggest that it is intrusive into the surrounding gneiss, but for the other foliated granite bodies it is not known whether they are ultimately of intrusive or metasomatic origin. No evidence of unconformable relations between gneiss and foliated granite (basal conglomerate or other distinctive lithology near contacts) was found in the surrounding gneiss.

The foliated granite bodies have not been observed in contact with any of the other abyssal intrusions and hence their relative age relations are not directly known. The York Point body has been deformed during folding thought to have occurred during the metamorphic maximum, whereas the body at Wreck Bay was emplaced after this folding, but has been deformed by folds thought to be related to emplacement of the mangerite or hornblende granite plutons or both. In the body at West St. Modeste two directions of schistosity are present locally, but only the younger persists into the neighbouring hornblende granite to the west. Thus by inference the foliated granitic bodies are thought to have been formed at least partly during and after emplacement of the metagabbro but before emplacement of the hornblende granite plutons.

Diorite

A single roughly equant body of altered diorite with an exposed area of 4.3 km^2 is located 5 km southeast of Pikes Feeder Pond. This body does not produce a topographically distinct erosion pattern but is characterized by a high aeromagnetic anomaly.

The rock consists of massive to slightly schistose, grey-weathering, grey and green, medium- to coarse-grained diorite. A little quartz is visible in many specimens. Quartz-rich gneiss inclusions were observed in the central and northern parts of the body, and pods of fine- to medium-grained leucocratic granite and pegmatite are common within it.

The rock consists mostly of plagioclase and blue-green hornblende. Augite, rimmed by hornblende, and about 5 per cent quartz are typically present. Up to 10 per cent feldspar is present in coarse irregular submesoperthitic intergrowths and in antiperthitic intergrowths in plagioclase. Most plagioclase is severely saussuritized but andesine was identified in one thin section. Minor epidote, chlorite, and biotite are common late alteration products. Magnetite-ilmenite, commonly rimmed by sphene and apatite, form up to several per cent of the rock locally. Zircon was observed in one thin section.

The diorite is intruded by minor granitic bodies probably related to the megacrystic granitic rocks, and was presumably severely altered during their emplacement. Remnants of coarse patchy submesoperthite suggest that the diorite may have originated during or prior to high-grade metamorphism. The relative age relations of the diorite and metagabbro intrusions are unknown.

Northeast of the diorite body the regional lineation pattern in the gneiss plunges toward the diorite with steepest plunges in its vicinity. As melanocratic gneiss is prominent to the north of the body, it is possible that the diorite represents an exceptionally deeply downwarped and intensely recrystallized part of the melanocratic gneiss. On the other hand the high magnetite content of the diorite, evident both in the aeromagnetic anomalies and locally in hand specimens, is not characteristic of the melanocratic gneiss and may indicate that the diorite was initially an independent intrusive.

Megacrystic granite to granodiorite

Three megacrystic granitic plutons, one large satellite body, and several smaller bodies lie within the map area. These bodies aggregate about 430 km^2 and are distributed about the northern periphery of the Long Range Precambrian inlier. Hybrid rocks, in which gneiss remnants and megacrystic and submegacrystic to massive granite rocks are intermixed, lie near the west margin of the Precambrian inlier between major plutons. For convenience of reference the megacrystic granitic plutons have been named in succession counter-clockwise as follows (Fig. 21):

(4) The Hooping Harbour pluton lies southwest of Canada Bay and extends southward approximately to the latitude of Fourché Harbour. The small body to the southwest of this pluton is called the (Hooping Harbour) satellite pluton.

(3) The Horse Chops pluton is northwest of Canada Bay at the northeast extremity of the Precambrian inlier.

(2) The Leg Pond pluton is east of Leg Pond at the northwest extremity of the Precambrian inlier.

(1) The Lake Michel pluton lies along the west margin of the Precambrian inlier with its broadest section south of the map area near Lake Michel. Only the northern extremity of the pluton is within the map area.

The megacrystic granitic rocks are commonly knobbly weathering and chiefly flesh coloured to pale grey, but green, pink and dark green varieties exist particularly within the greenschist terrane. The composition of these plutonic rocks varies from granite to granodiorite with quartz monzonite probably the most abundant. Colour index varies from about 2 to 20 (per cent of dark minerals), colour indices between 10 and 15 being most common. The Horse Chops pluton is most homogeneous in this respect with colour index estimates made in the field almost entirely between 5 and 15. Other plutons contain a substantial number of outcrops which exceeded these limits, but a regional pattern is not clearly evident. Megacrystic granodiorite of high colour index may be most abundant in the southern part of the satellite pluton where included metagabbro is prominent. Darker granitic rocks also appear to be more than normally abundant along the northeastern margin of the Hooping Harbour pluton but there they have been altered, and comparison with other parts of the pluton is difficult. Although leucocratic to

Table 5. Chemical analyses of megacrystic granitic rocks from the Long Range Mountains*

Sample	(BK70)	(BK70)	(BK70)	(BK69)	(BK69)	(BKJ69)	(BK70)	(BK69)	(BK70)	(BK69)	(BK70)		
Sample Number	336	157	136	477	466	308	158	465	119	469	159	Mean	SD
SiO_2	61.0	61.4	63.3	66.6	67.4	67.5	68.5	68.8	69.0	72.4	78.3	67.8	5.19
TiO_2	1.34	1.80	2.32	1.61	1.22	1.00	0.86	0.87	1.04	0.32	0.90	1.17	0.556
Al_2O_3	16.6	13.1	12.3	13.1	14.3	14.5	14.7	13.9	13.7	14.2	9.0	13.6	1.97
Fe_2O_3	3.1	4.2	4.2	3.2	2.8	2.5	1.9	2.6	2.5	2.0	2.0	2.8	0.838
FeO	1.7	5.0	5.4	3.3	3.5	1.6	2.6	3.5	2.0	0.5	2.2	2.8	1.55
MgO	2.0	1.5	1.6	1.1	0.8	1.2	0.7	0.8	1.4	0.5	0.8	1.1	0.483
CaO	3.2	3.9	4.1	3.0	2.2	2.1	1.7	1.6	2.2	0.5	1.6	2.3	1.12
Na_2O	3.3	2.8	2.2	2.4	2.9	2.9	3.0	2.8	2.6	4.0	1.9	2.8	0.572
K_2O	5.9	3.1	2.8	4.6	4.0	5.1	5.0	4.0	4.7	4.5	2.0	4.1	1.19
H_2O	0.6	0.7	1.1	0.4	0.6	0.7	0.5	0.4	0.5	0.5	0.5	0.6	0.197
MnO	0.07	0.15	0.14	0.11	0.09	0.06	0.11	0.10	0.05	0.03	0.06	0.09	0.039
P_2O_5	0.60	0.98	1.07	0.94	0.42	0.50	0.31	0.26	0.40	0.15	0.43	0.51	0.298
CO_2	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	-
Total	99.4	98.6	100.5	100.4	100.2	99.7	99.9	99.6	100.1	99.6	99.7		
Fe_2O_3 total	5.0	9.8	10.2	6.9	6.7	4.3	4.8	6.5	4.6	2.6	4.4		
Sr	0.11	0.039	0.027	0.039	0.022	0.058	0.027	0.020	0.054	0.076	0.11	0.047	0.034
Ba	0.25	0.10	0.087	0.13	0.080	0.12	0.14	0.086	0.13	0.052	0.063	0.11	0.056
Zr	0.12	0.15	0.17	0.099	0.10	0.15	0.080	0.13	0.13	0.038	0.093	0.12	0.039
Ce	0.073	0.085	0.083	0.060	0.070	0.091	NF	0.070	0.085	NF	NF	-	-
La	0.031	0.028	0.025	0.023	0.030	0.055	0.017	0.027	0.041	0.012	0.015	0.028	0.013
Yb	<0.0004	0.00086	0.0010	0.0016	0.00052	0.00065	0.00053	<0.0004	<0.0004	<0.0004	<0.0004	-	-
Cu	0.0031	0.0013	0.00091	0.0014	0.0019	0.0017	0.0026	0.0014	0.0028	0.00092	0.0011	-	-
Pb	0.0029	0.0018	0.0021	0.0026	0.0016	0.0035	0.0021	0.0017	0.0036	0.0016	0.0011	0.0022	0.0008
Zn	0.0094	0.0190	0.0160	0.0120	0.0110	0.0084	0.0150	0.0100	0.0070	0.0022	0.0079	0.0105	0.0049
Sn	0.00038	0.00072	0.00082	0.0011	0.00032	0.00049	0.00047	0.00022	0.00055	0.00039	0.00051	0.00049	0.00018
Q	11.19	24.55	30.50	29.18	29.57	26.49	27.35	32.17	30.33	30.78	56.18		
C	0.41	0.40	0.77	0.97	2.21	1.60	2.01	2.70	1.30	2.22	1.85		
Or	35.32	18.72	16.66	27.22	23.75	30.48	29.76	23.84	27.92	26.86	11.93		
Ab	28.26	24.19	18.72	20.31	24.63	24.79	25.54	23.87	22.09	34.15	16.21		
An	12.09	13.21	13.41	8.73	8.20	7.22	6.44	6.28	8.33	1.51	5.17		
En	5.04	3.81	4.01	2.74	2.00	3.02	1.75	2.01	3.50	1.26	2.01		
Fs	-	3.08	2.89	0.96	2.28	-	2.00	3.05	-	-	1.02		
Mt	1.85	6.22	6.12	4.64	4.08	2.48	2.77	3.80	3.61	0.79	2.92		
Il	2.58	3.49	4.43	3.06	2.33	1.92	1.64	1.67	1.98	0.61	1.72		
Hm	1.86	-	-	-	-	0.82	-	-	0.02	1.47	-		
Ap	1.41	2.32	2.50	2.18	0.98	1.17	0.72	0.61	0.93	0.35	1.00		
Quartz	<20	<20	<20	<20	20	<20	20	20	20	20	>20		
Microcline	<30	<30	30	>30	30	>30	>30	30	>30	>30	<30		
Plagioclase	>30	>30	30	<30	30	<30	<30	30	<30	30	30		
Biotite	m	m	m	m	m	m	m	m	M	-	m		
Hornblende	-	m	m	tr	m	-	m	-	-	-	m		
Chlorite	-	-	tr	-	m	-	tr	tr	tr	m	tr		
Epidote	m	-	tr	tr	-	tr	-	-	m	-	tr		
Muscovite	-	-	tr	-	-	tr	tr	-	tr	tr	tr		
Magnetite	m	m	tr	m	m	tr	tr	m	m	tr	tr		
Sphene	tr	tr	-	tr	tr	tr	tr	tr	tr	tr	tr		
Apatite	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr		
Zircon	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr		
Allanite	-	-	-	R	tr	tr	-	R	tr	-	-		
Pyrite	-	-	tr	tr	-	-	-	-	tr	-	tr		

*Chemical analyses by the rapid method staff and by the spectrochemical laboratory staff of the Geological Survey of Canada.
(BK70)336: Hooping Harbour Pluton, north part of 'tail'. (BK70)157: Leg Pond Pluton, southeast margin. (BK70)136: Leg Pond Pluton, southeast margin. (BK69)477: Lake Michel Pluton, southwest part. (BK69)466: Lake Michel Pluton, northern extremity. (BKJ69)308: Hooping Harbour Pluton, south part of 'tail'. (BK70)158: Leg Pond Pluton, central part. (BK69)465: Lake Michel Pluton, northwest central part. (BK70)119: Hooping Harbour Pluton, central part of 'head'. (BK69)469: Hooping Harbour Pluton, north part of 'head'.
NF not found; m minor; M Major; tr trace; R rare (BK70)159: Leg Pond Pluton, west margin.

melanocratic gneiss inclusions are known within the megacrystic plutons, quartz-rich gneiss appears to be represented in a higher proportion than it is within the gneiss complex as a whole. Assimilation of quartz-rich gneiss may explain local quartz-rich granitic variants within the megacrystic plutons (Table 5).

Most plutons are characterized by an abundance of subhedral to euhedral pink to white, potash feldspar megacrysts from 1 to 3 cm long, set in a medium-grained, quartz-bearing matrix. Megacrysts are unusually scarce, though present, in parts of the Horse Chops pluton. Biotite is commonly the only major mafic mineral evident. The Leg Pond and Horse Chops plutons are chiefly massive, but the southern and eastern parts of the Hooping Harbour pluton and most of its western satellite are schistose.

The megacrystic granitic rocks are composed principally of quartz, plagioclase, and microcline. Biotite, typically greenish or olive-brown, rarely makes up more than about 15 per cent of the rock. Pale green actinolite is a local constituent of megacrystic rocks in the greenschist terrane, but small amounts of hornblende with pleochroic formula x ochre, y brown-green, z sea-green to dark green, are present in 7 of 19 specimens examined from the amphibolite terrane. Epidote is most common in megacrystic rocks from the greenschist terrane whereas muscovite and chlorite are most common in those from the amphibolite terrane. Accessory minerals, dark opaques, apatite, zircon, and sphene are typically present; allanite is common and carbonate is rare.

Perthitic textures are common in microcline and locally perthite is zoned with coarse-grained albite lamellae in crystal cores, fine-grained lamellae in intermediate zones, and no visible perthite in marginal zones. Myrmekitic intergrowths are common though not abundant. Plagioclase commonly has albitic rims where in contact with microcline. Mafic minerals occur in clots, but sphene is typically separate from magnetite and locally is zoned.

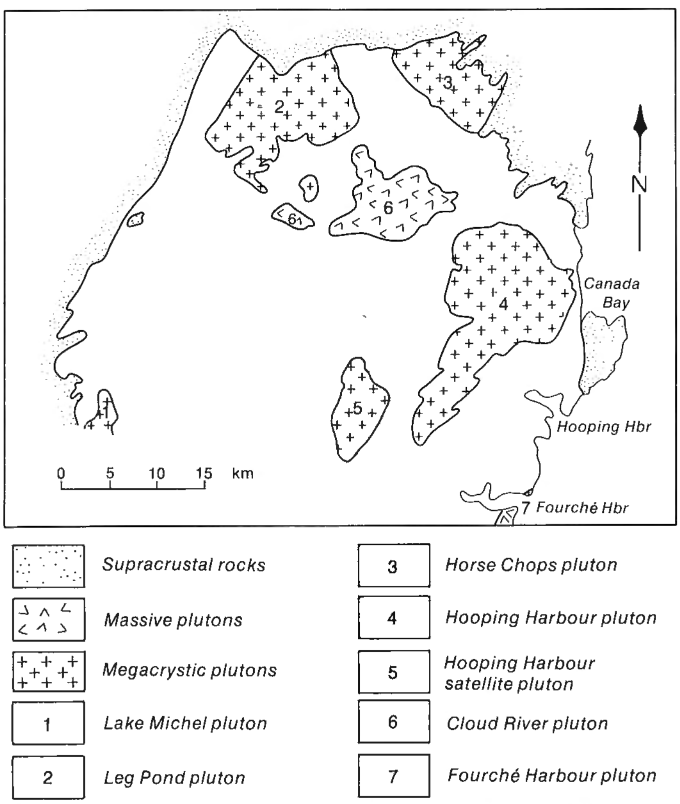


Figure 21. Names of granitic plutons in the northern Long Range, Newfoundland.

Contacts between megacrystic granitic rocks and the basement gneiss are nowhere known to be well exposed. At several localities bands of megacrystic granite are interleaved with gneiss or schist near contacts. Contacts between megacrystic plutonic rock and large gneiss inclusions may be sharp and wavey or they may be sheared. Foliation in the gneiss along the margins of the megacrystic granitic plutons with a few exceptions dips beneath the pluton, in many areas 45° or less.

Chemical analyses. Eleven samples (4.5 kg each) of megacrystic granitic rocks representing the Hooping Harbour, Leg Pond, and Lake Michel plutons were collected for chemical analysis (Table 5). The analyses demonstrate the wide range in composition evident in the megacrystic granitic rocks. Sample 159 with unusually high SiO_2 and normative quartz values may have been contaminated through assimilation of quartz-rich gneiss, which occurs along the contacts of the intrusion in the vicinity of the location sampled. As quartz-rich gneiss inclusions are probably the most common type of inclusion and some local areas of the megacrystic granitic plutons are obviously quartz-rich, this sample (159) has been included in the calculated mean (Table 5). Sample 477 from west of Lake Michel was obtained substantially beyond the map area and was not included in the calculated mean.

Age. Inclusion of metagabbro bodies within the megacrystic plutons (discussed in connection with description of the metagabbros) indicates that emplacement of the plutons postdates that of the metagabbro. Furthermore, evidence of early, dry, high temperature conditions in the gneiss (suggested by the local presence of hypersthene and mesoperthite) is absent in the megacrystic rocks. If high-grade metamorphism, which apparently affected all of the central and western parts of the northern Long Range Precambrian inlier and the granulite terrane in Labrador and Quebec, occurred everywhere at about the same time, then the megacrystic plutons of the Long Range postdate the mangerite and hornblende granite of the granulite terrane. Late andalusite found in sillimanite-bearing pelitic gneiss at one locality along the margin of the northwestern pluton may result from retrograde metamorphism accompanying emplacement of the megacrystic pluton subsequent to highgrade metamorphism. A Rb-Sr isochron for metacrystic granitic rocks of the western pluton by Pringle et al. (1971) gives an age of 1130 ± 20 Ma ($\lambda^{87}Rb = 1.39 \times 10^{-11} yr^{-1}$) which they interpreted to be the age of emplacement.

Mode of emplacement. Foliation inclined inward at moderate angles has been shown to be a common feature of gneiss that immediately surrounds the megacrystic plutons. (A clear exception to this generalization is one small megacrystic plug between Cloud River and Leg Pond plutons discussed under structure of the upper Cloud River region.) This configuration of country rocks suggests that the present surface is a section through the lower parts of the megacrystic plutons which, if viewed at a higher level, would be more extensive, or would be surrounded by outward dipping gneiss. Hamilton and Myers (1967) have quoted similar contact configuration about batholiths of the western United States, among other features, as evidence of downward warping of the country rocks that accompanied upward and outward spreading of diapirically emplaced granitic magmas. They further suggested that such intrusions may be accompanied by outward thrust faulting.

In the northern Long Range outward thrusting is suggested over local sectors of the megacrystic pluton contacts; elsewhere such movement of magma may have been absorbed through lateral injection of sills, or the resistance of country rock buttressed by rising plutons on either side may have forced the granite magma to slightly higher levels.

Alternatively movement of magma along some sectors of contact may have been relatively inactive compared with others, and in still other sectors, particularly where the granitic rocks are schistose and lineated, both gneiss and granitic rocks may have been folded during or after emplacement.

Sectors along which outward thrusting is suggested are found in association with the Torrent Cove assemblage, and along the northwest contact of the Horse Chops pluton. In the former case penetrative movement suggested by phyllonitization is directed at a high angle to the horizontal trace of the contact of the neighbouring pluton as shown by local pegmatite boudins. Furthermore, timing of phyllonitization so far as can be established, coincides with emplacement of the megacrystic plutons for it followed high-grade metamorphism, but preceded emplacement of the diabase swarm which intrudes the pluton. In the latter case the domal structure in the gneiss southwest of Horse Chops pluton pinches and becomes overturned southwestward where the northwestern part of the pluton contact bulges toward it.

Outward thrusting by injection of one or more sills is suggested along the south margin of the Leg Pond pluton. There, several isolated hills are capped by megacrystic granitic rocks with gently dipping gneiss outcrops between the hills, and the boundary of the main pluton is highly irregular in plan. Indeed, the absence of the aeromagnetic anomaly pattern characteristic of megacrystic granitic rocks in the amphibolite terrane elsewhere may suggest that the western half of this pluton is a thin, perhaps folded, sheet overlying the gneiss.

Outward thrusting may have been inhibited (by the presence of the adjacent Horse Chops pluton) along the northeast margin of the Leg Pond pluton where foliation in the bordering gneiss is locally vertical, thus forcing the magma to rise to higher levels or expand its margins elsewhere. This contact is followed by highly schistose melanocratic gneiss in which sheared inclusions of leucocratic gneiss are present. Quartz-rich gneiss, like that forming the core of an antiform some 5 km farther east, appears along the contact suggesting that here, in contrast to most megacrystic pluton contacts, country rocks from greater depth have been dragged upwards.

Folding of the megacrystic granitic rocks and enclosing gneiss on a large scale is thought to have affected the southern part of the Hooping Harbour pluton and its eastern satellite. As the supposed axis of folding is nearly parallel to the elongation of these granitic bodies its effect has been to locally steepen the dip of foliation in the bordering gneiss and to produce a parallel schistosity in the granitic rocks. This structure is discussed in the chapter on structural geology.

Hybrid rocks

The hybrid rocks underlie about 100 km^2 along the west margin of the Grenville inlier south of Leg Pond. They occur mainly in four areas (i) southeast of Leg Pond; (ii) southwest of the Leg Pond megacrystic pluton; (iii) in the core of Pikes Feeder antiform northeast of Pikes Feeder Pond; and (iv) at the north end of the Lake Michel pluton.

Southeast of Leg Pond, hybrid rocks consist chiefly of leucocratic to melanocratic gneiss interbanded in varying proportions with granitic rocks. In the northern part of the area scattered small outcrops of medium-grained equigranular quartz monzonite occur, whereas in the southern part megacrystic granitic rocks appear to predominate.

At the southwest end of the Leg Pond pluton hybrid rocks are formed from interbanded gneiss and megacrystic granitic rocks similar to those southeast of Leg Pond, and the granitic phase is locally slightly schistose. Farther to the southwest, however, the granitic phase is equigranular, leucocratic, mostly massive, and medium grained. Large irregular patches of this granitic phase occur within the gneiss and smaller veins and dykes intrude them. East of the large cross-shaped pond southwest of Leg Pond pluton, white, medium-grained granitic rocks are interlayered with meso- to melanocratic gneiss and pelitic schist. Parts of this latter body are megacrystic whereas other parts are massive, and some are garnet bearing.

At the southern end of Pikes Feeder Pond antiform hybrid rocks consist principally of massive megacrystic granite with a few scattered large inclusions of gneiss. Farther north leucocratic to melanocratic gneiss, quartz-rich gneiss, and calc-silicate rocks are interleaved with massive to weakly foliated megacrystic granitic rocks. Near the north end of the body, megacrysts in the granitic phase are smaller (less than 1 cm in diameter) and large patches of nearly equigranular, leucocratic to mesocratic granitic rock are present particularly along the margins of the hybrid rocks.

At the east margin of the Lake Michel pluton at the southwestern limit of mapping, hybrid rocks consist of megacrystic to massive granitic rocks that contain large lenses or bands of leucocratic to melanocratic gneiss and some quartz-rich gneiss.

Age relations and origin. The hybrid rocks are characterized by a granitic phase that is in large part megacrystic to submegacrystic. Furthermore the granitic phase is typically massive or only weakly foliated, a characteristic it shares with the neighbouring megacrystic plutons. These features suggest that the development of the hybrid rocks is related to emplacement of the megacrystic plutons. Whether the equigranular and megacrystic components that characterize the granitic phase in different areas are of the same age is uncertain, but the fact that there appears to be a gradation between them suggests that they are related.

Massive granodiorite, quartz monzonite, and granite

Two plutons and two smaller bodies of massive granite to granodiorite are present within the map area. Two minor bodies east of Pinware on the north coast of the Strait of Belle Isle, which may be parts of a larger body mostly covered by waters of the Strait, are included in this unit (for discussion see foliated granitic rocks, unit Hgdn). The largest body, the Cloud River pluton, and its associated satellite (Fritts, 1953) lie along the upper reaches of Cloud River in the northern Long Range (Fig. 19). A second pluton, present at Eastern Head and Granite Point on either side of the mouth of Fourché Harbour, is here referred to as the Fourché Harbour pluton. This pluton extends southward down the coast as far as the north shore of Robineau cove. Its contacts south of Fourché Harbour have not been mapped but their position can be inferred from the grain of the topography, which suggests that the pluton projects up to about 1.5 km inland. The remaining plug lies wholly in Quebec just north of Bradore Bay. Together these bodies underlie about 70 km^2.

The Cloud River pluton is characterized by rounded, tree-covered hills separated by broad valleys, the rock being for the most part more susceptible to erosion than the surrounding gneiss and megacrystic granite. Outcrops in the valley bottoms show prominent but widely spaced jointing whereas those at hilltops are commonly crumbly and poorly exposed. In hand specimens the rock is typically massive, medium grained, and grey-white with readily recognizable quartz, potash feldspar, plagioclase, biotite, and commonly sphene. Scattered large subhedral to anhedral potash feldspar crystals are present locally.

The rock is massive and medium to coarse grained (commonly 3 to 6 mm) with rare anhedral to subhedral microcline megacrysts up to 13 mm long. The principal minerals are quartz, microcline, intermediate oligoclase, biotite, and hornblende (x ochre, y brown-green, z dark blue-green). Small amounts of muscovite, chlorite, and epidote are present locally. Sphene and magnetite are major accessory minerals locally forming one per cent or more of the rock. Zircon, apatite, allanite, and pyrite are present locally.

Microcline and plagioclase are mostly present in about equal proportions along with 20 to 30 per cent quartz so that the pluton is chiefly made up of quartz monzonite. Potash feldspar is perthitic, commonly with perthite most abundant in crystal cores. Locally, in the northern part of the intrusion, patchy irregular submesoperthitic intergrowths are present and microcline crystals commonly contain large amoeboid patches of plagioclase. Plagioclase is locally zoned from calcic oligoclase cores to sodic oligoclase rims. Albite occurs along margins with microcline. In some crystals areas of sericitized albite are present within oligoclase.

Contacts of the Cloud River pluton are poorly exposed. Foliation in the surrounding gneiss along the northeast, south, and southwest margins of the pluton appears to dip gently to steeply toward the pluton, but in the northwest, where the contact between the pluton and surrounding gneiss is irregular, it may be deformed. Near the twin lakes on Cloud River the contact with melanocratic granodioritic gneiss is gradational. Inclusions of metagabbro, of leucocratic gneiss, and of pelitic gneiss have been observed, but are not abundant in parts of the intrusion examined during the present study (outcrops along the central part of Cloud River above the twin lakes, visited by Fritts (1953), were not examined).

The Bradore Bay plug is massive, medium grained, white weathering, and pinkish on fresh surfaces. Scattered pink subhedral megacrysts up to 1 cm long are locally present. The principal minerals are quartz, intermediate oligoclase, microcline, and biotite with chlorite and muscovite present in some specimens. Accessory minerals are magnetite, zircon and apatite. Carbonate is locally present along some feldspar cleavages. Microcline is very slightly perthitic. Plagioclase is distinctly more abundant than microcline and the rock is therefore granodiorite.

Gneiss at the northeastern and west margins of the Bradore Bay plug dips away from it at 20 to 36°. Elsewhere examination of air photographs suggests that the gneiss wraps around the plug forming a dome-like structure.

The Fourché Harbour pluton is pink to pale flesh-grey or slightly greenish, medium grained, and massive. It is composed of quartz, pink to pale flesh-coloured feldspar, chlorite, and epidote. Mafic minerals are typically very fine grained in anhedral patches. Locally carlsbad twins up to 2 or 3 mm long are recognizable but elsewhere the feldspar patches are finely polycrystalline. Rounded mafic inclusions are present in places and may be isolated or present in steeply dipping zones. The rock is commonly fractured or brecciated, locally severely so.

In thin section the rock consists of quartz, microcline, albite-oligoclase, chlorite, biotite or stilpnomelane, and epidote. Accessory minerals are apatite, zircon, magnetite, fluorite, sphene, and allanite. Fluorite is unusually common being present in two thirds of the specimens examined. Quartz is typically thoroughly strained and is serrated or shows mortar texture along its margins. Potash feldspar has irregular patches of microcline twinning. Plagioclase locally displays bent polysynthetic twin lamellae, and crystals commonly have serrated margins. In many specimens feldspars are to a greater or lesser degree recrystallized to a fine-grained mosaic. Chlorite or chlorite-biotite-epidote are fine grained, mostly somewhat altered in appearance, and occur in patches, or locally in log-like pseudomorphs after amphibole(?). The original mafic minerals have been thoroughly recrystallized but textures do not closely resemble those in the Fourché Point schist or gneiss to the west where patches of late, fine-grained clear biotite have grown in a biotite-bearing rock.

The Fourché Harbour pluton is locally intruded by pegmatite and granitic dykes. It is cut by diabase dykes of the Long Range swarm and by east-west dykes and irregular bodies of similar rock. One basic dyke striking 125° and dipping 55° southwest was found to be schistose parallel to its margins, and to contain lenses of granite elongated parallel to schistosity. The contact between granite and schist north of Eastern Head is a fault, striking west and dipping 75° south, which curves southwestward into the first cove west of Eastern Head and may follow the contact south of Fourché Harbour.

Chemical analyses

Two samples (2.3 kg each) of the Cloud River pluton and one (1.4 kg) of the Fourché Harbour pluton were analyzed. (Tables 6 and 7 respectively.) The samples from the Cloud River pluton show slight differences in silica and mafic contents, perhaps reflected in the appearance of minor hornblende in the more mafic sample (270 BK70). Because the Cloud River pluton is more homogeneous than the megacrystic plutons, samples from this pluton are more likely to be representative than a similar number of megacrystic granitic samples would be. The pluton falls within the quartz monzonite composition range. The sample from the Fourché Harbour pluton represents the northern extremity of the pluton, which may be somewhat more leucocratic and potassic than the main mass to the south represented by outcrop along the coast. The analysis nevertheless suggests that this pluton is more 'granitic' than the other plutons described here.

Age relations and origin

All of the massive granitic plutons intrude the basement gneiss complex. The Cloud River pluton furthermore is younger than the metagabbro and contains inclusions of it. Dykes of the Long Range dyke swarm intrude the Fourché Harbour pluton but the other plutons lie beyond the western margin of the swarm. The age limits for the massive plutons are thus the same as those for the megacrystic plutons. Contacts between the two are not known. In contrast to the megacrystic plutons, however, the massive granitic plutons are largely free of foliation (the isotropic fabric of the Cloud River pluton was emphasized by Clifford and Baird, 1962). Insofar as this may indicate that deformation, which affected parts of the megacrystic plutons, preceded emplacement of the massive granitic plutons it suggests that the latter plutons are the younger.

No contacts between the Bradore Bay granodiorite plug and the country rocks were seen. Because the surrounding gneiss typically has an imposed foliation, whereas the granodiorite is massive, there seems little doubt that the latter is the younger and has intruded the gneiss. Lack of foliation and lineation also suggests that it is younger than the schistose granitic bodies. No evidence of granulite facies metamorphism is evident within the plug and none was found in the immediately surrounding gneiss. Because the plug lies at the western limit of mapping, however, the gross pattern of metamorphism in its vicinity is not known and its emplacement might conceivably have preceded granulite facies metamorphism. This, nevertheless, is considered the less likely alternative and the plug is grouped tentatively with the massive plutons of the Long Range. Pervasive but variable cataclastic textures within the Fourché Harbour pluton may reflect its intrusion under different conditions than affected the other massive plutons (or these textures may be related to relatively deep-seated phases of later faulting). Faults along the northern contacts of the intrusion suggest that at least the northern part of the pluton has been brought to its final resting place by faulting. Diabase dykes

Table 6. Chemical analyses of the Cloud River granitic pluton*

Sample number 51 (BK70)				Sample number 270 (BK70)			
Oxide	Wt%	Norm Mineral	Wt%	Oxide	Wt%	Norm Mineral	Wt%
SiO_2	69.4	Quartz	36.15	SiO_2	65.6	Quartz	30.76
TiO_2	1.45	Corundum	1.05	TiO_2	1.70	Corundum	1.11
Al_2O_3	12.3	Orthoclase	21.98	Al_2O_3	12.9	Orthoclase	21.42
Fe_2O_3	3.1	Albite	20.39	Fe_2O_3	3.8	Albite	20.43
FeO	3.2	Anorthite	9.05	FeO	4.1	Anorthite	10.83
MgO	0.6	Enstatite	1.50	MgO	1.2	Enstatite	3.01
CaO	2.7	Forsterite	1.06	CaO	3.2	Forsterite	1.84
Na_2O	2.4	Magnetite	4.51	Na_2O	2.4	Magnetite	5.54
K_2O	3.7	Ilmenite	2.77	K_2O	3.6	Ilmenite	3.25
H_2O	0.5	Hematite	-	H_2O	0.7	Hematite	-
MnO	0.07	Apatite	1.56	MnO	0.13	Apatite	1.82
P_2O_5	0.67			P_2O_5	0.78		
CO_2	<0.1			CO_2	<0.1		
Total	100.1			Total	100.1		
Fe_2O_3 total	6.7			Fe_2O_3 total	8.4		
Sr	0.028			Sr	0.031		
Ba	0.14			Ba	0.14		
Zr	0.12			Zr	0.13		
Ce	NF			Ce	0.057		
La	0.0095			La	0.019		
Yb	0.00076			Yb	0.00088		
Cu	0.00077			Cu	0.0012		
Pb	0.0024			Pb	0.0019		
Zn	0.0082			Zn	0.0130		
Sn	0.00054			Sn	0.00060		

Approximate modal composition

Sample Number	51	270	Sample Number	51	270
Quartz	20	20	Epidote	-	tr
Microcline	35	35	Magnetite	tr	tr
Oligoclase	35	35	Sphene	tr	tr
Biotite	10	10	Apatite	tr	tr
Hornblende	-	m	Zircon	R	R
Chlorite	tr	tr	Pyrite	tr	-
Muscovite	tr	tr			

*Chemical analyses by the rapid method staff, and by the spectrochemical laboratory staff of the Geological Survey of Canada. Samples 51 and 270 represent the central and southern parts of the main body of the Cloud River pluton respectively.

tr trace; NF not found; R rare; m minor.

within the intrusion are altered, fractured or brecciated, but because some independent fault movements along the coast as well as some metamorphism are known to be later than the dykes elsewhere, the age of the fault along the northern part of the pluton relative to emplacement of the dykes is not known. Faulting of the northern part of the pluton may have accompanied folding of the megacrystic granitic rocks farther west, or it may have occurred later.

Basic Dykes of Labrador

Dark grey to olive basic dykes up to 3 m wide have intruded the gneiss along the north shore of the Strait of Belle Isle and similar dykes are probably present inland as well. These dykes strike north and nearly east, parallel to prominent joint sets in the gneiss. The dykes show massive to subtrachytic textures, some containing plagioclase laths up to 3 mm long. Biotitie and chlorite or amphibole are major constituents and indicate that the dykes are altered. Minor quartz and magnetite are also present.

Table 7. Chemical analysis of the Fourché Harbour granitic pluton*

Sample Number 35 (BK69)			
Oxide	Wt%	Norm Mineral	Wt%
SiO_2	76.2	Quartz	41.54
TiO_2	0.22	Corundum	2.46
Al_2O_3	12.6	Orthoclase	30.32
Fe_2O_3	1.3	Albite	20.41
FeO	0.9	Anorthite	1.86
MgO	0.3	Enstatite	0.75
CaO	0.4	Forsterite	0.29
Na_2O	2.4	Magnetite	1.90
K_2O	5.1	Ilmenite	0.42
H_2O	0.05	Hematite	-
MnO	0.04	Apatite	0.05
P_2O_5	0.02		
CO_2	<0.1		
Total	100.0		
Fe_2O_3 total	2.3		
Sr	0.0029		
Ba	0.022		
Zr	0.069		
Ce	<0.05		
La	0.015		
Yb	<0.0004		
Cu	0.00077		
Pb	0.0012		
Zn	0.0072		
Sn	0.00020		

Approximate modal composition

Sample Number	35	Sample Number	35
Quartz	>20	Sphene	tr
Microcline	>30	Apatite	tr
Albite	<30	Fluorite	tr
Chlorite	m		
Biotite	m		
Epidote	m		

*Chemical analyses by the rapid method staff, and by the spectrochemical laboratory staff of the Geological Survey of Canada. The sample is from the north shore of Fourché Harbour.

m minor; tr trace.

Two east-striking dykes in the gneiss north of the Strait of Belle Isle were dated by the K-Ar whole rock method at 920±32 Ma and 560±21 Ma (Wanless et al., 1974). The apparently older (western) dyke contains fine-grained amphibole in the matrix between feldspar laths whereas the eastern dyke, which yields the younger date, contains chlorite but is otherwise of similar composition. The dykes are probably of the same late Grenville age but have undergone different late Grenville or early Hadrynian alteration. A few dykes with similar trends have been observed in other parts of the Grenville Province (Mont Laurier map area, Wynne-Edwards, 1969; Cook-D'Audebourg area, Davies, 1965). The two ages fall within the range found by Grasty et al. (1969) for late dykes on the east coast of Labrador near the Grenville front.

Hadrynian-Lower Cambrian supracrustal rocks

Introduction

On Belle Isle the Precambrian gneiss at the northeast end of the island is overlain by feldspathic quartzite and conglomerate of the Bateau Formation. Both are intruded by dykes that fed sills and flows of the Lighthouse Cove Formation exposed on the south margin of the island. The Lighthouse Cove flows are in turn overlain by arkosic sediments correlated with the Bradore Formation.

In the Canada Bay area and at Henley Harbour up to 2 m of arkosic sediments typically lie on Grenville gneiss beneath basalt of the Lighthouse Cove Formation. Similar rocks occur in thin layers between flows, and in the overlying Bradore Formation. Farther west, and in places in Canada Bay as well, no basalt is present and a thicker section of arkosic sediments, previously correlated entirely with the Bradore Formation (equivalent to Cloud Mountains Formation of Betz, 1939), overlies weathered Grenville gneiss. Williams and Smyth (this memoir) have correlated conglomerate at Burnt Head (southeast Canada Bay) with the Bateau Formation, and have implied that conglomerate at Otter Cove (farther north on Canada Bay), and therefore perhaps all the sediments beneath the basalts, may also belong to the Bateau Formation. Alternatively it is possible that a significant unconformity exists between the Bateau and Bradore formations marked by the regolith present at many places on the Grenville gneiss where they are overlain by younger rocks, but that is not apparently present beneath the Bateau Formation on Belle Isle (and at Burnt Head?). In the present account, therefore, arkosic sediments along Canada Bay west of Burnt Head are grouped with the Bradore Formation, which is considered to intertongue with the Lighthouse Cove Basalts near its base. It should be recognized nevertheless that Bateau equivalents may exist beneath the basalt at Canada Bay.

The Bradore Formation is conformably overlain by the Forteau Formation, and by the Devils Cove Formation in the Canada Bay area. On Belle Isle the Forteau Formation is in fault contact with possibly younger rocks of the White Point Formation here included in the Forteau. A complete description of these formations has not been attempted. The structural and metamorphic histories of these beds were studied because they lie close to the Precambrian gneiss and might shed light on events recorded within the gneiss.

The Bateau Formation

The Bateau Formation (Williams and Stevens, 1969) consists of quartzite, boulder conglomerate, siltstone, slate, and minor volcanic rocks, which are well exposed along the northeast coast of Belle Isle. White quartzite of the White Islands and White Rocks is also assigned to the Bateau Formation as is an isolated coarse plutonic boulder conglomerate that occurs between the basement and volcanic rocks within a fault slice west of Burnt Point, Canada Bay (Willians and Smyth, this memoir). On Belle Isle the formation is intruded by abundant diabase dykes so that the sedimentary rocks are observed as screens and remnants between dykes (Fig. 22). Furthermore the top of the formation is not exposed and the remainder has been broken into several fault wedges. Estimates of the maximum thickness of exposed beds are thus conjectural but at least 200 m are present. Williams and Stevens (1969) have estimated that the exposed thickness ranges from 75 m at Greenham Bight to a maximum of 240 m north of Bateau Cove.

The type section for the Bateau Formation, as designated by Williams and Stevens (1969), comprises the exposures north of Bateau cove where these authors described some 90 m of conglomerate overlain by about

Figure 22. A screen of fine-grained Bateau conglomerate between basic dykes at Bateau Cove, Belle Isle. Both beds and cleavage in the conglomerate dip steeply to the right (east). (GSC 160167).

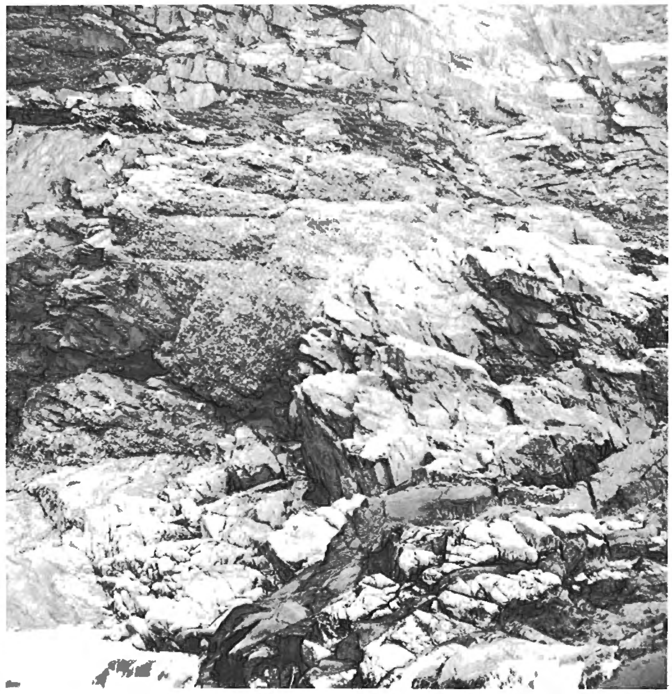


Figure 23. Conglomerate (grey in background) in fault contact with quartzite (light grey in foreground) of the Bateau Formation north of Bateau Cove. Basic dykes have intruded the quartzite which is dipping gently to the right (north). Beds in the conglomerate at Bateau Cove dip 78 degrees east (towards the observer). (GSC 160175).

150 m of quartzite. At Bateau Cove, purplish boulder conglomerate contains a few lenses of purplish to greenish arkosic sandstone up to 2 m thick. These lenses dip from 60 to 70° east and have a superimposed cleavage dipping about 80° east. Boulders and cobbles are abundant in the conglomerate, are up to 60 cm in diameter, and consist predominantly of gneissic, schistose and massive plutonic rocks, but vein quartz, quartzite, chert (Williams and Stevens, 1969), and some fine-grained, dark green siltstone are also represented. The matrix is gritty and chloritic to sericitic. At Bateau Cove, conglomerate overlies the gneiss to the west, but on the highlands to the north a wedge of massive white quartzite appears at the base of the formation and thickens northward. The contact between conglomerate and underlying quartzite is not exposed; its trace is irregular and has little topographic expression. A contact between conglomerate and overlying quartzite is exposed on the shore a few hundred metres north of Bateau Cove (Fig. 23) and may be projected northward to intersect the lower quartzite on the highlands less than 1 km farther north. Where exposed this contact is knife sharp, is well lithified, is smooth but undulating, and dips steeply east. Bedding in the conglomerate is not evident in the immediate vicinity but a pebble lens extending for 2 m parallel to faint, texturally defined lamellae in the quartzite dips 5° north and lies at a high angle (60 to 70°) to the contact. Boulders in the conglomerate along the contact were examined for truncation against it but evidence for this was inconclusive. A search for further bedding in the quartzite on the highlands to the north revealed one locality where topset beds of a complete crossbedded unit, marked by slight concentration of 'black sand', indicate a dip of 15° northeast. If bedding in the conglomerate at Bateau Cove represents the attitude of the conglomerate wedge as a whole (as seems likely in view of the excellent section provided by Bateau Cove), then the regional difference in attitude between quartzite and conglomerate appears too large to permit a conformable relationship between the two. The contact is therefore probably an unconformity or a fault, most likely the latter.

Rocks structurally overlying the conglomerate north of Bateau Cove consist mostly of massive white quartzite with pebble lenses near the lower contact and faintly defined crossbedded units, mostly up to 30 cm thick, observed locally. Grains of white- to pink-weathering feldspar are common locally. Farther north near Eagle Cove the quartzite dips east at 35° near the coast but is apparently nearly vertical along the western contact with gneiss. Up to 5 m of grey-green chloritic arkose is present at the base. This is succeeded gradationally by pale grey to white quartzite bearing a few per cent of feldspar grains, and scattered quartzite-pebbles within the first 9 to 12 m. Farther up in the section the quartzite is massive and irregularly jointed. Minor crossbedded units were seen only near the shoreline. Veins of white quartz cut the quartzite locally.

South of Bateau Cove on either side of Greenham Bight the Bateau Formation is more variable. At the north shore of Greenham Bight sandstone is interlayered with conglomerate in greater proportion than at Bateau Cove. Farther south a section was described by Williams and Stevens (1969) as follows:

> "At the southern exposed limit of the Bateau Formation south of Greenham Bight the sequence of units is more variable, although the exposed section is not more than 250 ft (76 m) thick. There the basal boulder conglomerate is very coarse, but only 10 ft (3 m) thick, and is followed in ascending order by 30 ft (9 m) of dark gray slate, pyritic siltstone, and quartzose sandstone; 100 ft (31 m) of gray siltstone and slate with cleavage parallel to bedding and locally displaying kink bands; and 50 ft (~15 m) of siltstone and slate with thin sandy beds. The latter is overlain by a 5-ft (1 1/2 m) green conglomerate bed that is sheared, and is composed mainly of discoidal quartz fragments from 1 inch to 8 inches (2 1/2-20 cm) in diameter and local sandstone and siltstone fragments up

to 2 ft (6 m) long. This conglomerate is overlain by about 50 ft (15 m) of sheared, epidote-rich, green volcanic rocks, which appear to have fragmental textures locally. The sequence of beds changes along strike, and at the northern Greenham Bight exposures it includes 50 ft (15 m) of slaty quartzite containing 1- to 4-ft (.3 to 1.2 m) beds of crossbedded quartzite."

Specimens of impure quartzite, conglomerate matrix, and slaty siltstone from the Bateau Formation between south Greenham Bight and Gull Island Cove were examined in thin section. These rocks typically consist of angular to subangular or lenticular clasts, commonly 1 mm or less in diameter, of quartz, microcline, and albite in a matrix composed of finer grains of these minerals together with one or more of muscovite, chlorite, epidote, and late metamorphic biotite. Accessory minerals are magnetite, apatite, sphene-leucoxene, and zircon. Quartz is commonly strained but albite fragments are not more altered than those in the basement gneiss (in contrast to plagioclase in the Bradore Formation which is unidentifiable except locally near the base). Metamorphic biotite, where it occurs, is in fine grained, clear, randomly oriented crystals, locally in 'nests' in the shadow of larger quartz grains, whereas muscovite and chlorite, and locally lenticular quartz define a moderate to strong foliation.

The contact between the Bateau Formation and the gneiss is probably mainly faulted, but fault movement appears to decrease southward, and the zone of movement may move into the gneiss to the west of the contact so that the unconformity is preserved on the shoreline south of Greenham Bight. Gneiss between Black Joke Cove and Eagle Cove on the north and extending south to Bateau Cove has been finely fractured and most mafic minerals have been altered to chlorite, and epidote. In contrast to the gneiss along the west coast south of Wreck Bay, little oxidation is evident. Similar altered gneiss is present in a poorly exposed wedge that projects along faults through the Bateau Formation from the southwest to within 200 m of the shore near Gull Island Cove (Fig. 24). On the north point of Greenham Bight discrete faults appear to cut the gneiss at the base of the Bateau Formation, but at the south point the conglomerate at one locality clearly lies unconformably upon the gneiss. Although this latter gneiss may be separated from gneiss farther west by a continuation of the faults seen to the north, it appears that no extensive zone of fine fracturing is present west of the Bateau Formation at Greenham Bight.

Age relations and correlation. The Bateau Formation is younger than the gneiss that it unconformably overlies, and is older than the latest dykes of the Long Range swarm that have intruded it. It is thus older than the Lighthouse Cove Formation (minimum age greater than about 540 Ma based on fossils in the overlying formations), but whether there is an essentially conformable sequence from Bateau Cove through Lighthouse Cove to Bradore Formations (as suggested by Williams and Stevens, 1969), or whether the Bateau Formation is distinctly older than the Bradore Formation depends on the validity of correlation of minor volcanic rocks within the upper part of the Bateau Formation with the Lighthouse Cove Formation, and on the ability of postdyke deformation to produce the present disposition of Bateau fault wedges without recognizably disturbing either the structure of the dyke swarm or that of the gneiss. The problem is rendered more difficult because no section exposing both Bateau and Bradore beds has been recognized. The persistence of the dyke swarm through gneiss and the various fault wedges of the Bateau Formation appears to support the conclusion that faulting of the Bateau Formation took place before emplacement of the dykes, and hence that a period of deformation intervened between deposition of Bateau and Bradore formations. A more detailed examination of the strikes and dips of both dykes and their foliation in the gneiss and in the various fault wedges of the Bateau Formation is necessary to establish this relation.

The Precambrian-Bradore unconformity

The contact between Precambrian plutonic rocks and the overlying Bradore Formation is exposed locally along the western and northeastern margins of the Long Range, in the Blanc Sablon area, and in the Henley Harbour area (Fig. 25, 28, and 29). At many of these localities weathering of the Precambrian surface is evident, the thickness of the weathered zone (regolith) being up to 2 m or more. At the western and northern exposures the regolith is friable and less resistant to erosion than either the underlying Precambrian or overlying arkosic rocks, but in the Canada Bay region it is indurated and the resistance of the altered gneiss at the contact is equivalent to those below.

The Precambrian-Bradore regolith on the west side of the Long Range is well exposed on the tributary creek that enters Torrent River from the north at the entrance to Eastern Blue Pond. There, massive medium-grained leucocratic gneiss is exposed in the creek bed, and the canyon wall to the south is formed of 1.5 m of nearly flat-lying, well-cemented Bradore arkose that overlies 1.5 m of red-brown friable granitic regolith containing 60 per cent or more weathered feldspar. No conglomerate was observed at the contact. A covered interval separates the regolith from fresh basement gneiss in the creek bed.

Near Blanc Sablon the gneiss at the contact with overlying Bradore arkose is argillized. Pebble lenses lie along the contact and are overlain by thin lenticular argillaceous and arkosic sandstone interbeds. A specimen of bright green argillized gneiss (derived from the tide zone) proved to contain a clay fraction entirely composed of disordered mica (illite). Farther west, near Bradore Bay, quartzite within the

Figure 24. View looking north toward Gull Island Cove, Belle Isle. Quartzite forms a screen between schistose dykes in the foreground. To the west beyond the nearest dyke a long poorly exposed wedge of fractured gneiss extends down the valley to within a few tens of metres of the cove. The hill in the background is of Bateau quartzite and dykes. (GSC 160181).

basement gneiss complex displays colour banding, which transects bedding and is similar to that present locally in the lower part of the Bradore Formation (Fig. 26 and 27).

At Henley Harbour about 1 m of pink Bradore arkose is exposed at the top of the bluffs below and east of the basalt (Fig. 28). The gneiss below is throughly fractured and stained red, but this deformation does not extend into the Bradore arkose. Similar fracturing and alteration is common in the surrounding gneiss over most of the eastern and southern part of the island.

Along the shoreline of Canada Bay the unconformity between basement gneiss and the Bradore Formation is well exposed at several places between Otter Cove and Fly Point. On the east shore near the head of Otter Cove more than 0.5 m of yellow-green altered gneiss containing quartz eyes about 1.5 mm in diameter is overlain by 15 cm of conglomerate (overlying rocks have been eroded in the immediate vicinity). On the west shore of Otter Cove southwest of Green Island typical leucocratic gneiss at the water's edge is overlain locally by 2.5 to 3 m of pale yellowish green rock in which pegmatite lenses are still recognizable. This is overlain in succession by a 20 cm bed of conglomerate containing some angular quartz fragments up to 20 cm in diameter, a 1 m bed of arkose with quartz-feldspar fragments up to 3 cm in diameter, and finally by 30 cm of dark grey thoroughly indurated laminated siltstone that lies against a basic volcanic rock. Farther south both the gneiss at the unconformity and a large gneiss fragment some 4.5 m long included within the Bradore sediments appear to be much less altered than the regolith to the north.

At the Bradore-gneiss contact north of Fly Point, pebbly arkose overlies yellow-green altered gneiss and the contact is throughly indurated (Fig. 29). This alteration extends into the gneiss along shore for about 30 m beyond the contact but the true thickness of altered gneiss is not apparent. Amphibolite bands within the gneiss in the altered zone are stained deep red.

The granitic leucocratic gneiss below the unconformity near Pikes Feeder Pond consists principally of quartz and submesoptherthitic microcline. Sodic plagioclase is present chiefly as irregular patches in potash feldspar. Mafic minerals are hornblende, chlorite with a few remaining shreds of biotite, and dark opaque minerals. Accessory minerals comprise scattered large crystals of sphene, zircon, and apatite. In the overlying friable regolith immediately below the Bradore Formation, microline shows the same abundance and the same submesoperthitic texture evident in the unweathered rocks below, but the plagioclase phase of the perthite is altered to a dusty red-brown material. Plagioclase grains are completely altered to fine-grained, sericite-like material and similar material, in part mixed with hematite, chlorite, and biotite shreds, surrounds microcline grains. Sphene is altered to a white opaque mineral, probably leucoxene.

The yellow-green altered gneiss near the unconformity at Canada Bay contains as its main constituents quartz, microcline, and a fine-grained, sericite-like mineral. Mafic minerals except magnetite are largely altered to hematite. Accessory minerals are zicron, apatite, and sphene-

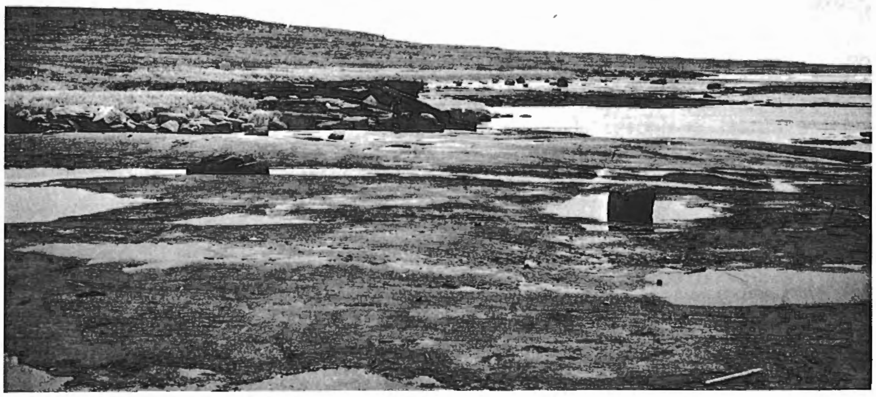


Figure 25. Bradore arkose unconformably overlies a nearly horizontal surface of gneiss east of Lourdes du Blanc Sablon. (GSC 160160).

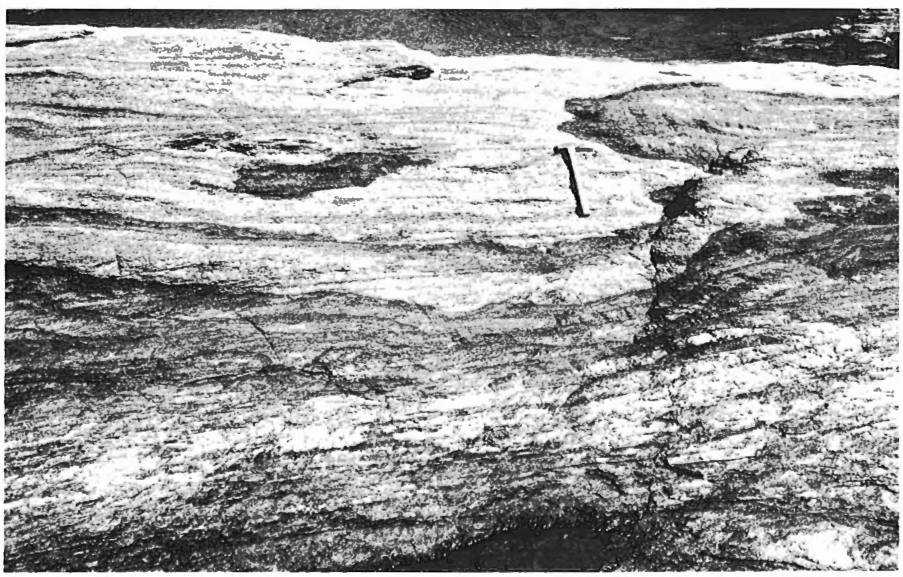


Figure 26. Gneissosity and colour banding in Precambrian quartzite north of Bradore Bay. (GSC 160109).

leucoxene. In some specimens a few remnants of early muscovite characterized by impurities along cleavage traces are present; whereas in others fresh poikilitic muscovite of later origin appears within the fine sericite-like matrix. In the least altered gneiss, fine sericite-like alteration is restricted to grain boundaries and plagioclase crystals; in more altered gneiss this fine-grained alteration extends along fractures in microcline grains and in places isolated remnants remain in optical continuity. In extreme cases only quartz remains.

Comparison of X-ray powder patterns made of the regolith near Pikes Feeder Pond and at Canada Bay indicated that the mica fraction in the former regolith is relatively disordered, polymorph lines being weak or absent as might be expected in a weathered material. In the patterns of the regolith at Canada Bay, however, lines of the $2M_1$ muscovite polymorph are well defined as would be consistent with metamorphic recrystallization.

Lighthouse Cove Formation

The name of the Lighthouse Cove Formation was proposed by Williams and Stevens (1969) for the basic volcanic rocks that overlie basement gneiss and underlie arkosic sandstone of the Bradore Formation on Belle Isle. The name was later extended to include similar rocks in the vicinity of Cloud Hills, Newfoundland, and in the Henley Harbour-Table Head region in Labrador (Strong and Williams, 1972). The type section for the formation is at Lighthouse Cove at the south end of Belle Isle where the volcanics are approximately 90 m thick (Williams and Stevens, 1969). The formation is probably as much as 305 m thick at Barbers Cove (Williams and Stevens, 1969). A single basalt flow about 20 m thick is present at Henley Harbour. At least two basic flows separated by arkosic sandstone, siltstone, and conglomerate are present between Cloud River and Northwest Brook. These flows probably aggregate less than 75 m in thickness.

Diabase and altered diabase dykes at least in part if not entirely of the same age as the Lighthouse Cove flows occur in a northeasterly trending swarm (Long Range swarm) that extends from Belle Isle southwest at least as far as Little Harbour Deep. A few northeasterly trending dykes are known to be present in the gneiss farther south (Fritts, 1953; Bostock and Cumming, 1973) and it is possible therefore that the swarm extends the full length of the two Precambrian inliers, a distance of over 320 km. The width of the swarm is approximately 27 km west of Little Harbour Deep-Great Harbour Deep, but because it extends to the coastline north

Figure 27. Colour banding in Bradore arkose on Greenly Island. (GSC 160113).

Figure 28. Bradore arkose forms a resistant undeformed ledge overlying shattered gneiss on the east side of Henley Island. The top of the overlying Lighthouse Cove basalt is visible in the left background. (GSC 160133).

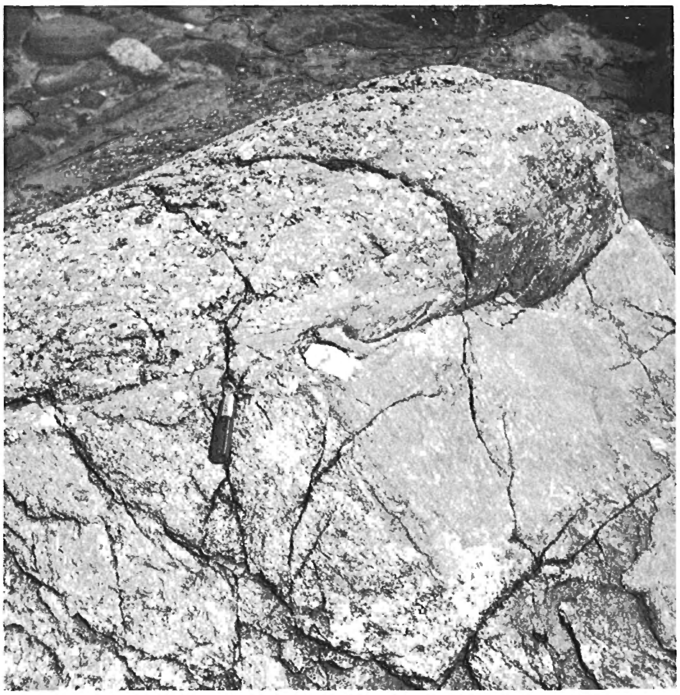


Figure 29. Pebbly arkose of the Bradore Formation unconformably overlies weathered leucocratic gneiss near Fly Point, Canada Bay. The weathered gneiss (regolith) and the Bradore arkose have both been subjected to greenschist facies metamorphism. (GSC 157985).

of Great Harbour Deep its full width is not exposed within the map area. Most dykes trend northeasterly parallel to the swarm as a whole, and are steeply dipping to vertical. They are commonly up to 60 m thick, the thickest known being over 105 m thick as reported by Foley (1937) in the valley of Soufflets River.

Flows. In the Cloud Hills region, the Lighthouse Cove Formation forms an imposing but discontinously exposed basalt capping at the edge of the Precambrian granite-gneiss highland. From the thickest section (less than 75 m aggregated in two or possibly three flows) between Northwest Brook and Cloud River, the formation apparently thins in both directions parallel to strike and across the trend of diabase dykes of the Long Range swarm in the underlying basement. At Beaver River where the flows pass beneath the Bradore Formation a window of megacrystic granite projects through a columnar flow. South of Cloud River, for the most part, only one flow is evident and the basalt thins toward Otter Cove, where the last outcrops are represented by isolated flow cappings which overlie 2 m of Bradore arkose. Absence of flows along much of the Bradore-gneiss contact in this region suggests that the flows never were extensive near and south of Otter Cove.

In Cloud River canyon at least two flows are present, the lower flow lying on basement gneiss and containing a few quartz cobbles along its basal contact. The upper flow is separated from the lower by a metre or two of Bradore arkose, suggesting an intertonguing relation. A similar succession of two flows, separated by an unknown thickness of arkose and conglomerate of the Bradore Formation, is present along the upper, northeast margin of exposure of the Lighthouse Cove Formation northwest of the lower reaches of Cloud River. There, the upper flow is clearly thinner than the lower, although the upper contact is concealed in the creek valley. Scattered occurrences of Bradore arkose and conglomerate up to about 6 m thick are known between gneiss and basalt as far north as Cloud River, but they appear to be absent at this horizon farther north. It is possible that this distribution results from the upper flow having extended southeastward beyond the limit of the lower one so that the arkose beds, which lie between flows northwest of Cloud River, lie directly on basement to the southeast of the river.

The basalts are fine grained to aphanitic with plagioclase in phenocrysts or aggregates of phenocrysts up to 5 mm in diameter. They are dark grey, blue-grey, olive-green, or pale grey-green. Pale green to red-purple cherty fracture fillings are common. Chlorite-filled amygdules from 1 to 5 mm in diameter are present locally and in places these are rimmed by carbonate or hematite. The rock near the lower-flow contacts (the only contacts seen) is commonly slightly vesicular.

Columns from 20 to 60 cm in diameter are evident in the basalt from Beaver River south to Cloud River beyond which they are less common. Clifford (1965) reported that the columnar zone, which is well developed in the lower central part of the flow near Cloud River, thins southward and is absent near Otter Cove. At one locality between Cloud River and Otter Cove, where a dyke of the Long Range swarm appears to form a feeder to the overlying flows, the attitude of the columns in the flow directly over the dyke appears to be variable perhaps as a result of late movement of magma in a feeder dyke below the flow. Rare small bodies of laminated sediment showing soft sediment deformation are inclusions within the basalt. Breccias composed of basalt fragments in a basalt matrix of slightly different colour are present but not common.

The basalt typically is slightly porphyritic with plagioclase phenocrysts or glomerocrysts or both up to 5 mm commonly accompanied by slightly smaller pyroxene phenocrysts, in a microlitic matrix. In the southern exposures near Otter Cove and as far north as the cirque-like valley north of Cloud River, epidote and saussurite are commonly abundant, whereas farther north the rocks are less altered and interstitial chlorite is present. In one specimen interstitial material resembling finely devitrified glass was observed, and it is likely that such glass has altered to chlorite elsewhere. Some interstitial quartz is common; muscovite is rare. Locally, pigeonite and carbonate are present. Magnetite-ilmenite is ubiquitous.

At Henley Harbour the Lighthouse Cove Formation consists of a single basalt flow about 20 m thick exposed immediately west of Henley Harbour and at the north end of Castle Island (Fig. 1). No rocks overlie the basalt. Similar basalt is exposed at St. Peter Islands and Table Head (not examined during the present study). No flows were observed within or beneath the Bradore Formation in Labrador and Quebec west of Henley Harbour and it is likely that they never did extend to this area. The upper half of the flow at Henley Harbour is hackly fractured and the lower half is columnar. The rock is fine grained and dark blue-grey with patches of hematitic stain. Both top and base are nearly horizontal. Seen in thin section this basalt is fine grained and porphyritic to glomerocrystic. The principal minerals are labradorite, augite, and magnetite-ilmenite, with minor quartz, muscovite, chlorite, and amphibole. The size distribution of plagioclase in the basalt is locally polymodal with coarsest crystals up to 4 mm long, severely sericitized intermediate laths 1 mm long, and matrix laths 0.1 mm long.

On Castle Island gneiss can be seen to within about 30 cm of the lower contact. The base of the flow immediately west of Henley Harbour is obscured by talus, but at the southwestern end of the island about 12 m of red-brown arkosic sandstone, horizontally bedded, overlies fractured gneiss and may project beneath the flow.

On Belle Isle the Lighthouse Cove Formation consists chiefly of dark green basalt which, near the base of the formation, is typically massive, medium to fine grained, and largely structureless save for a few vesicles along contacts and abundant irregular fractures. Higher within the formation reddish oxidized flow contacts and yellow-green, epidote-rich agglomerate zones locally mark flows up to 20 m thick. These flows are dark green, or more rarely partly blue-grey to purplish, and fine grained. Locally they contain vesicles, or amygdules filled with carbonate, epidote, or chlorite. In places thin, pillowed horizons containing pillows up to about 75 cm long, similar to those at Barbers Cove, are present (Fig. 30 and 31). West of Blandfords Cove a zone of small rounded fragments up to 15 cm in diameter, containing amydules in concentric patterns, may be pahoehoe clasts. At Barbers Cove a bed or lens of sandy to silty arkosic rock, up to about 60 m thick and resembling similar rocks within the Bradore Formation, lies between the flows. The absence or poor development of columnar structure, and the abundance of oxidized flow contacts and breccia zones within the volcanic rocks on Belle Isle contrasts with the Lighthouse Cove flows at Henley Harbour and Cloud Hills.

The basalts on Belle Isle lie mostly on Precambrian gneiss, but at Lark Harbour they are separated from the gneiss to the east by a metre or more of steeply dipping chloritic schist-bearing lenses of pebble conglomerate, which are tentatively correlated with the Bradore Formation. A diabase dyke has intruded the contact between the schist and gneiss; the contact between the schist and basalt appears conformable and gradational. South of Lark Harbour, where the eastern limit of the basalt is again exposed on the coast, the volcanic rocks appear to be faulted against the gneiss. On the point between Green Cove and White Point Cove dark green massive basalt lies directly upon a moderately southward dipping surface of well banded gneiss. The contact is parallel to banding in the underlying gneiss, but steps abruptly downward by about 1 m at intervals up the bluff face to the north. No sediment or flow banding was observed along the contact, but scattered vesicles are present in the volcanic rock. No weathered zone is evident in the gneiss at the contact and no oxidized flow contacts were seen in the bluff 100 m or more high. At Gull Battery Cove similar massive dark green basalt lies on a highly irregular but southeastward dipping gneiss surface, again without evidence of weathering at the contact. At this locality a diabase dyke merges with the overlying basalt and no oxidized zones are visible in the bluffs above. The absence of primary structures in the lower massive basalt, in contrast to their relative abundance in the overlying flows, together with the crosscutting relationship and apparent plucking out of the gneiss locally parallel to their banding, may suggest (although they do not prove) that the lower massive basaltic rock is a sill. If true, this interpretation may explain the absence of a weathered horizon at the basalt-gneiss contact. The sill (if it exists) was probably not intruded everywhere beneath the flows, because an agglomerate zone was found within 2 m of the contact with gneiss on the highlands northeast of Barbers Cove.

Joints and fracture zones in the basalt of Belle Isle are abundant, commonly irregular, and in many places are coated with epidote. At Lark Harbour, joints cut the basalt into huge subrectangular blocks, but do not appear to penetrate the gneiss to the east. Along the southwest coast steeply dipping fracture zones parallel faults marginal to the basalt, and a metre or more of schistose greenstone is present locally along these faults. North of Barbers Cove the basalt is cut by zones of vertical to steeply southeast-dipping schistosity parallel to that in the dykes to the northeast, indicating that this deformation is later than both.

The basalt of the Lighthouse Cove Formation forms homoclinal sequences in each part of the map area where the unit is exposed. At Henley Harbour the rocks are nearly horizontal but of restricted extent. On Belle Isle, flow attitudes are bimodal; most flows along the west margin of the island and about Lighthouse Cove dip south to southwest, whereas those on the promontory west of Blandfords Cove and at Barbers Cove dip southeast. The alternate appearance of southeast-dipping flow remnants on promontories underlain by gneiss, separated by downdropped blocks in which younger south- to southwest-dipping sediments overlie the basalt, suggests that the basalt is tilted as well as graben faulted. In the Cloud Hills region the basalt lies on a surface that dips northeasterly at about 10° near Horse Chops ridge. Farther southeast across faults at Northwest Brook and Cloud River, the dip increases to 20 - 25° near Otter Cove.

Figure 30. Pillowed basalt overlying an agglomerate horizon in the Lighthouse Cove Formation at Barbers Cove, Belle Isle. (GSC 160176).

Figure 31. Close-up of pillows showing chilled margins and epidote-rich interstices in Lighthouse Cove Formation at Barbers Cove, Belle Isle. (GSC 160177).

Diabase dykes. Basic dykes of the Long Range dyke swarm form prominent northeast-trending ridges, which are the dominant physiographic feature (Fig. 32) over much of the northeastern Long Range and on Belle Isle. Some dykes are massive, some have chilled margins up to 2 m thick about a medium-grained core; others are polygonally jointed with joints more or less perpendicular to their contacts, and still others are hackly fractured or slightly schistose. On Belle Isle, dykes commonly show some schistosity or cleavage; schistosity is particularly well developed in dykes in the northern and eastern parts of the island. In some dykes isolated angular fragments of country rock up to 3 m long are present. In the eastern part of the northern Long Range, dykes are altered and are green to dark green, but near the west margin of the swarm they are less altered and grey. A similar change is evident in dykes on Belle Isle, where the more schistose, greener, and altered dykes occur in the north and along the east coast. In many dykes the principal mineral components are approximately equigranular, but dykes in which slightly porphyritic plagioclase crystals up to 4 mm long occur in a matrix with a 1 or 2 mm grain size, are common.

In thin section the least altered, western, diabase dykes are composed principally of augite and andesine-labradorite with accessory magnetite-ilmenite. Small amounts of quartz were observed but olivine was not. Alteration of the western dykes is typically restricted to local interstitial patches of epidote-chlorite-serpentine and minor sericite. On Belle Isle the least affected western dykes contain augite that is nearly unaltered, but plagioclase is typically more altered than it is in the western dykes of the Long Range. In the more altered dykes the principal minerals are amphibole and saussuritized plagioclase with variable proportions of chlorite and epidote. Opaque minerals are altered to slatted intergrowths of magnetite-leucoxene. Small amounts of quartz, biotite, and apatite are typically evident, and carbonate, prehnite (?), and a mineral of low birefringence and refractive index (possibly chabazite) are present in some specimens. Two prismatic phenocrysts composed of chlorite-serpentine-carbonate rimmed by pyroxene were observed in one moderately altered specimen. Where pyroxene is preserved it is commonly rimmed by fine-grained, colourless to brownish amphibole, which is in turn rimmed by blue-green amphibole. Plagioclase may be thoroughly saussuritized (fine-grained epidote-albite identified by x-ray) or, particularly in dykes west of Great Harbour Deep, it may be partly dusty brown andesine-labradorite with clear cores and rims. All degrees of alteration from almost unaltered western dykes to completely recystallized eastern dykes are known.

The dykes in the Long Range typically strike 35 to 55° with most dykes striking within a few degrees of 45°. Two altered diabase dykes in the Canada Bay area strike 100° and 85°, and similar dykes were seen within the Fourché Harbour pluton. Most dykes for which dips were determined are close to vertical, but a few dips range from 50° southeast to 65° northwest. Although the number of dykes for which dips were measured is small (30), and exceptions are known, there is a suggestion that northwesterly dipping dykes are slightly more numerous on the southeast side of the swarm and that southeasterly dipping dykes are concentrated on the northwest side.

On Belle Isle most dykes trend 12° ± 10°, and dip more than 65°. Northwest-dipping dykes are most abundant on the east and northwest coasts of the island (Fig. 33), but in the basement between Wreck Cove and White Point Cove both northwest- and southeast-dipping dykes are common.

A slightly fan-shaped structure of the dyke swarm seems probable both in the Long Range and on Belle Isle, but the trend of dykes is more northerly on Belle Isle than it is in the adjacent part of the Long Range.

Breccia dykes. A total of eight breccia dykes of different types were observed within the map area, seven near Canada Bay (Fig. 34), and one farther south. These dykes are 12 m or less wide, commonly have irregular contacts, and locally pinch and swell to an extreme degree. Although their strike is thus difficult to measure, estimates along the west shore of Canada Bay suggest that most dykes strike between 50 and 75°, or about parallel to the Long Range dyke swarm.

Some breccia dykes (Fig. 34) are altered basic dykes containing variable proportions of country rock fragments, and others include fragments of earlier basic dyke rock as well. Internal contacts separating inclusion-free and inclusion-rich altered diabase were observed locally. One such dyke contains an older breccia core in which diabase

Figure 32. View looking south along a ridge-forming diabase dyke of the Long Range dyke swarm near Hooping Harbour. (GSC 153327).

Figure 33. Northwest-dipping diabase dykes of the Long Range dyke swarm on the east coast of Belle Isle. (GSC 160166).

fragments are schistose, surrounded and intruded by more massive younger diabase. In other breccia dykes a diabase phase is absent and the dyke consists of variable proportions of country-rock in a finely comminuted, purplish stained matrix. The proportions of matrix varies from abundant to scarce and in one tabular breccia body the country rock fragments appear to be essentially in place.

In thin section the matrix of one dyke was found to consist of small fragments of saussuritized porphyritic, microlitic altered diabase, and unaltered phyllite fragments in a matrix of carbonate-epidote-chlorite. The matrix of breccia dykes free of diabase fragments consists of comminuted fragments of country rock down to single mineral grains without obviously added material except for hematite and scattered to clustered octahedra of magnetite.

The breccia dykes are clearly younger than the Torrent Cove phyllite they intrude. On the other hand, their age relations with the dykes and flows of the Lighthouse Cove Formation appear contradictory, for in one case, a diabase dyke has apparently intruded a breccia dyke although the two dykes have followed the same fracture, whereas in another, a breccia dyke appears to intrude the gneiss within a metre of overlying basic flows of the Lighthouse Cove Formation which lie across its trend. These relations, and the fact that the breccia dykes appear to follow about the same trend as the diabase dykes may indicate that the two are related and formed at about the same time. Country rocks are warped along the margins of some dykes and this suggests that their emplacement may have been accompanied by fault movement. Repeated fault movements might account for the complex relations observed within the breccia dykes.

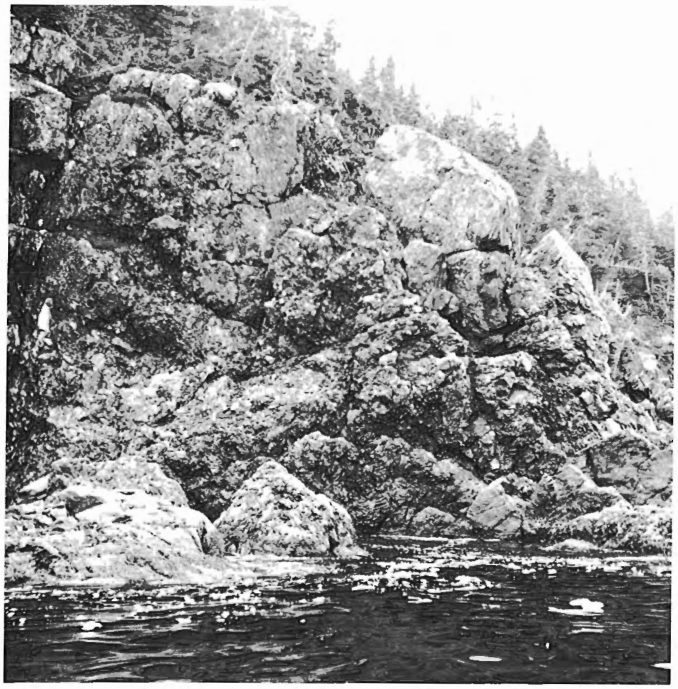


Figure 34. Diabase breccia dyke intrusive into the Torrent Cove assemblage at Canada Bay. (GSC 158036).

Two small breccia bodies resembling those at Canada Bay without a diabase phase, were found in the gneiss near Henley Harbour. One near the south end of Castle Island is a poorly exposed breccia zone a few centimetres wide of uncertain strike. The other, about 15 cm wide, cuts across leucocratic gneiss along the coast southwest of Henley Harbour (Fig. 35). Both of these occurrences are probably fault breccias.

Age. Dykes of the Long Range swarm are nowhere known to penetrate rocks younger than the Lighthouse Cove flows. Locally both on Belle Isle and in the Cloud Hills the lowermost unit of the Lighthouse Cove Formation is fed by dykes of this swarm. Thus at least some dykes in the Long Range dyke swarm are of the same age as the flows. A few small, younger basic dykes were reported by Betz (1939) within the allochthonous rocks of the Ordovician klippe south of Canada Bay, but these are probably not related to the Long Range swarm.

Flows of the Lighthouse Cove Formation overlie the basement gneiss complex and intertongue with the basal part of the Bradore Formation (Fig. 36). The basal carbonate beds of the Forteau Formation, which conformably overlies the Bradore Formation, contain archeocyathid reefs, and trilobite fragments of late Early Cambrian age (Fritz in Williams and Stevens, 1969). Fossil evidence therefore shows that the flows and the dykes of the Long Range swarm are of pre-late Early Cambrian age (older than approximately 540 Ma).

Diabase dykes near the western margin of the Long Range dyke swarm have been collected by Clifford (1965), Fahrig (in Wanless et al., 1968), and by Pringle (in Pringle et al., 1971) for K-Ar whole rock age determination. Ages obtained by Fahrig, 751 ± 100 Ma, and Pringle et al., 805 ± 35 Ma, are similar and have been interpreted as minimum ages of intrusion for dykes at the western margin of

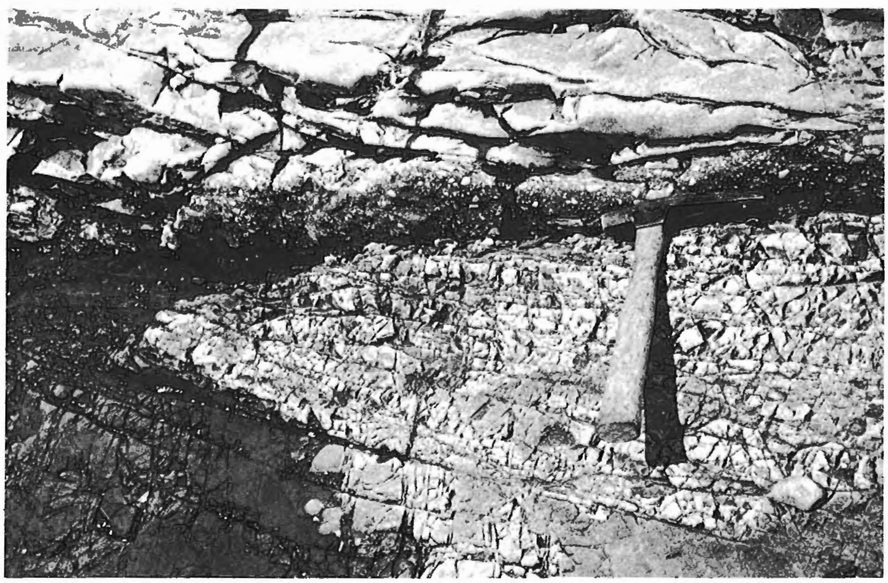


Figure 35. Fault breccia (breccia dyke?) southwest of Henley Harbour. (GSC 160123).

the Long Range swarm by these authors. The date obtained by Clifford (334±100 Ma) is from a dyke at the margin of the greenschist facies metamorphic terrane. Because of the spread of ages and the likelihood that the dykes were feeders to Lighthouse Cove flows lying conformably below Lower Cambrian strata, Stukas and Reynolds (1974) treated 11 whole rock specimens collected from dykes by the writer west and southwest of Fourché Harbour by the ^{40}Ar-^{39}Ar dating method. They showed that the dykes do contain excess argon, thus explaining the "old" 751 and 805 Ma ages of Fahrig and Pringle et al., and they concluded that the primary age is probably 605±10 Ma, i.e. about latest Hadrynian.

This age for the dykes and the Lighthouse Cove flows fits well with the stratigraphy. The flows lie conformably below Bradore sandstone, which in turn grades upward into Forteau shale and limestone bearing late Early Cambrian fossils (about 540 Ma).

Five K-Ar whole rock ages have been obtained from flows of the Lighthouse Cove Formation: from the Table Head flow, 375 Ma (Wanless et al., 1965); from the Henley Harbour flow, 411±17 Ma and 421±17 Ma (Wanless et al., 1974); and from the northern exposures near Canada Bay beyond the greenschist facies overprint, 413±16 Ma and 427±17 Ma (Wanless et al., 1974). The mean age of the flows, as derived from the four more recent determinations (in which isotope dilution techniques were used in potassium analyses), is 418 Ma. This age is over 100 Ma too young to be the age of volcanism. Furthermore, this age is essentially the same as a K-Ar biotite age (434±18 Ma) of the basement gneiss complex near Fourché Point, which is thought to reflect uplift presumably at the end of Ordovician orogeny. Thus, even though the flows dated are from areas west of visible late metamorphism in the gneiss and supracrustal rocks, it seems likely that these five dates reflect uplift rather than the age of volcanism.

The consistent northeasterly trend and tholeiitic composition (Strong and Williams, 1972) of the Long Range dyke swarm suggest that the dykes were all intruded under similar crustal conditions, and are therefore perhaps of the same age. On the other hand cross-cutting relations (Fig. 38), variation in style of deformation of dyke and country rock from place to place (Fig. 37), and inclusion of schistose dyke fragments within nonschistose dykes (see discussion of breccia dykes), suggest that dykes of significantly different ages may be present. Nevertheless each instance of complex age relations might in itself be explained by faulting during emplacement, by local variation in condition of country rocks during emplacement of dykes, or by relatively small intervals between emplacement of successive dykes. Further evidence is required before the existence of dykes of significantly different ages can be accepted.

Bradore Formation

During the present study the Bradore Formation has been examined only locally in conjunction with mapping of the Precambrian basement. As a result of the reconnaissance nature of mapping in the Labrador and Quebec sector, virtually no examination of the Bradore was made in these areas. For more complete description of these and succeeding Paleozoic formations the reader is referred to the accompanying reports by L.M. Cumming and by H. Williams and W.R. Smyth as well as Williams and Stevens (1969).

The Bradore Formation was defined by Schuchert and Dunbar (1934) with the type section east of Bradore Bay, Labrador. The formation is exposed along the northern periphery of the Long Range, outliers reaching as far south as St. Pauls Inlet on the west and as far south as Sugarloaf hill on the east. It also appears along the coast of Labrador and Quebec from Bradore Bay to West St. Modeste, and again in the Henley Harbour region and on Belle Isle. Thickness estimates of the formation in the Long Range are from 45 to 60 m at Blue Mountain (Nelson, 1955), and 55 m (present work) at Canada Bay; at Forteau Bay in Labrador 85 m are present (Cumming, 1970). On Belle Isle the thickness is probably between 60 and 90 m.

About the northern Long Range the lower beds of the Bradore Formation consist of pink, purple, grey, brown, green, or dark grey arkosic sandstone containing pebble beds and lenses. Locally, as along the west margin of the Long Range, silty to shaly beds are prominent, and in the region about Wild Cove fine-grained, greenish subgreywacke is

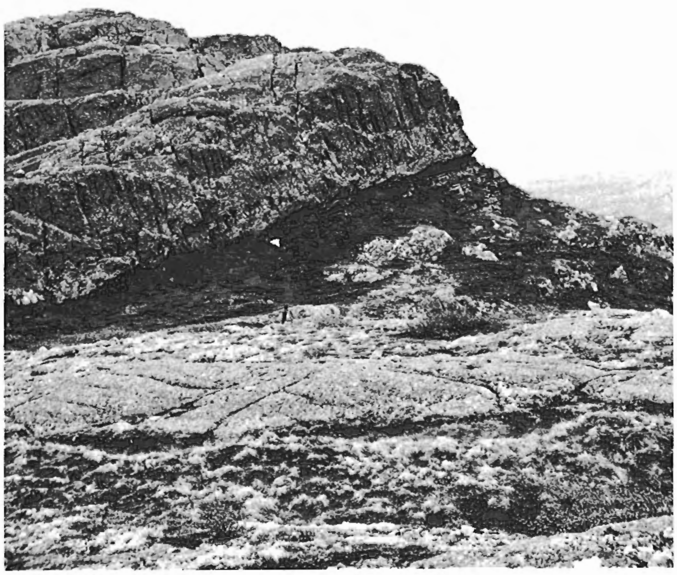


Figure 36. A basalt flow of the Lighthouse Cove Formation near Canada Bay overlies a metre or more of jointed Bradore arkose and conglomerate that rest on massive ledges of leucocratic gneiss. (GSC 157979).

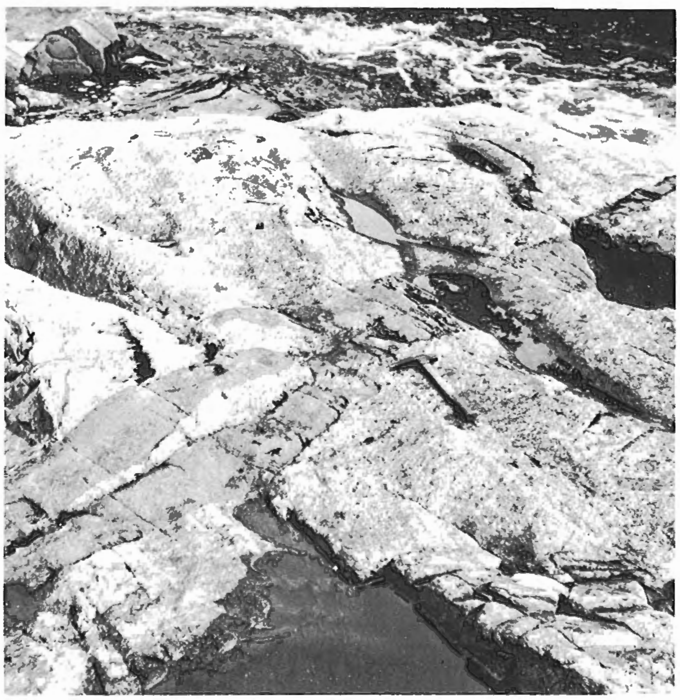


Figure 37. Disrupted basic dyke with discontinuity 'healed' by flowage of granitic material into the breach, near Northeast Arm, Hooping Harbour. (GSC 158020).

interbedded. Glauconite, chiefly as granules, is a common to abundant constituent of the lower Bradore beds along the west margin of the Long Range and has been found as far northeast as Boony Lake. Locally, within the glauconite-bearing beds, chips of collophanite (optically isotropic sedimentary apatite identified by x-ray) in part showing colloform lamination are present. These chips are largely free of inclusions and may form up to several per cent of individual thin beds in which they occur. The principal localities at which they were found (in thin section) are along the creek valley south of Leg Pond and in the Bradore inlier on the highlands farther south. Traces of possibly similar material were observed near Boony Lake. Dietz et al. (1942) suggest that recent phosphate nodules off the coast of California appear to be forming in association with coarse-grained glauconitic sands on topographic highs on the seafloor, where the waters are well oxygenated organic matter is relatively scarce, and the sedimentation rate is generally slow; furthermore, such nodules occur neither too close to shore, nor beyond the continental slope. This type of environment provides a likely model for the origin of the source formation from which the phosphatic-glauconitic Bradore beds were derived. Such source beds, as indicated by the immature condition of the Bradore Formation, were likely close at hand, and primary phosphate-bearing beds may persist locally (as yet undiscovered) at the Bradore-gneiss unconformity. Along the northeast margin of the Long Range, and particularly near Wild Cove, disseminated magnetite is more abundant in the basal beds of the Bradore Formation than elsewhere. Just north of Wild Cove where the Bradore beds are exposed along a late northeast-trending fault, magnetite content in a pebble conglomerate locally exceeds 10 per cent. Magnetite, present as tiny octahedra, forms concentrically banded, flattened, pisolite-like structures up to 5 cm in diameter. Individual pisolite-like structures consist of poorly defined lamellae of iron-rich material interbanded with fine-grained silica containing a little biotite or stilpnomelane. Magnetite is also abundant in the rock matrix between quartz pebbles.

The Bradore sandstone consists chiefly of angular to rounded quartz of variable grain size with up to 20 per cent of slightly altered microcline grains. Plagiocalse clasts are rarely identifiable and where they are evident they are typically intensely argillized. Detrital biotite, muscovite, chlorite, and amphibole are locally evident, the first two being more common than the latter two. Glauconite granules and collophanite chips are peculiar to the basal beds in the northwest, collophanite being concentrated in a more restricted area about Leg Pond. Accessory magnetite-hematite and zircon are typical, and apatite, sphene-leucoxene, and tourmaline were observed locally.

About 1 km north of Otter Cove, 55 m of clastic rocks of the Bradore Formation overlie gneiss and are overlain by grey carbonate beds. Betz (1939) reported 175 m (Cloud Mountains Formation) on the north shore of Otter Cove. This higher figure is probably a typographical error because the contacts and dip of the formation as shown by Betz in this area are correct, but when combined with the contours shown on his map do not permit more than half of this thickness to be present. The section north of Otter Cove is composed of 46 m of grey to purplish or greenish arkosic sandstone with local fissile zones toward the top, overlain by about 7 m of black dense to slaty argillaceous siltstone. Similar dark beds near the top of the formation at Fly Point contain a higher proportion of somewhat coarser quartz and feldspar grains, and are about 12 m thick.

On Belle Isle the Bradore Formation consists of about 3 m of purplish argillite at the base overlain by a thicker sequence of buff, pink, and reddish purple arkosic sandstone, commonly crossbedded, with local pebble and siltstone lenses. The top of the formation is marked by a dark grey to dark purple sandstone that thickens westward toward Round Head and, locally near the southern tip of the island, is overlain by 60 cm of brownish weathering sandstone containing a 10-cm bed of hematitic iron ore (Williams and Stevens, 1969). These authors also reported the presence locally of basalt clasts in conglomerate lenses, which has not been observed either in the Henley Harbour or Cloud Hills regions where basalts intertongue with the Bradore Formation.

A single thin section from the middle part of the Bradore Formation east of Lighthouse Cove consists mostly of angular to subangular quartz with about 15 per cent of microcline clasts, a small amount of very fine grained sericitic matrix, and accessory magnetite and zircon. It is notably free of plagiocalse clasts and in this respect resembles the Bradore Formation elsewhere. A thin section of arkose from a 60-cm lens of silty to sandy arkosic beds between flows at Barbers Cove consists of fragments of quartz, microcline, and finer grained plagioclase (albite) in a matrix consisting in large part of fine-grained epidote. Sheaths of a fibrous mineral of positive elongation, probably acicular amphibole, are abundant within many quartz clasts.

A metre or more of green to greyish, chloritic schist containing lenses of arkosic quartz pebble conglomerate and a few carbonate-rich lenses are in conformable contact with

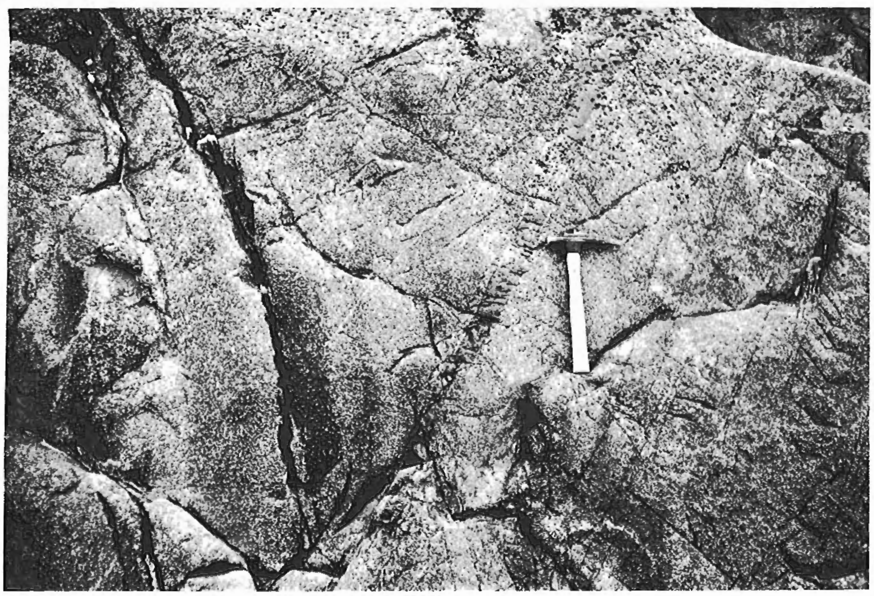


Figure 38. A large, medium-grained, diabase dyke of the Long Range dyke swarm intruded by a later fine-grained diabase dykelet on the west shore, Northeast Arm. Hooping Harbour (GSC 160083).

basalt to the west at Lark Harbour. To the east they are separated from a near-vertical diabase dyke by a short covered interval. Farther south along the coast the basalt is faulted against the gneiss basement. Matrix from a conglomerate lens beneath the basalt is composed of brecciated quartz, microcline, and plagioclase clasts in a fine grained but recrystallized sericitic matrix. Because the basalt does not appear to be intruded by diabase dykes it seems likely that both basalt and underlying sediments are related to the Bradore Formation. On the other hand, the possibility that both are related to the Bateau Formation cannot be excluded.

Contacts between the Bradore Formation and the basement gneiss complex on Belle Isle are not exposed, but the between-flow lens of arkose at Barbers Cove shows that arkosic sedimentation comparable to that in the Bradore had begun before extrusion of the Lighthouse Cove flows had ceased. By analogy with other parts of the map area it is likely that local pockets of Bradore sediments may lie below the flows as is possible at Lark Harbour. Contacts between the upper flows and the Bradore Formation are locally exposed, the best being on the shoreline 200 m east of Lighthouse Cove. There, a thin flow is overlain by a metre or more of conglomerate containing chips of dark, fine-grained volcanic rock followed by slaty siltstone of the Bradore Formation. Williams and Stevens (1969) report a "discontinuous 3-ft (.9 m) thick purple to green amygdaloidal lava flow about 20 ft (approximately 6 m) above the basal contact" of the Bradore Formation in a small fault-bounded remnant east of Lighthouse Cove.

The above relationships lead to the problem of whether arkosic sedimentation was continuous from Bateau to Bradore time as the late Precambrian sea spread westward or whether a significant interruption occurred in this pattern. A more detailed comparison of Bateau and Bradore formations is desirable to examine this question before their limits with respect to the Lighthouse Cove basalt are defined. In the present paper the Lighthouse Cove basalt is considered to intertongue with the Bradore so that all exposures of arkosic sediment on Canada Bay west of Burnt Head are assigned to the Bradore. Elsewhere the terminology of earlier workers is applied.

A complete, gently dipping section of the Bradore Formation is exposed in a cliff at Round Head Cove. This section starts with the uppermost Lighthouse Cove flows at sea level and terminates with the conformably overlying basal carbonate-bearing beds of the Forteau Formation exposed a short distance south of the top of the cliff. Contours (50-foot interval) on the 1:50 000 scale topographic map of Belle Isle indicate that the cliff is about 52 m high and suggest that the Bradore Formation is about 60 m thick at Round Head Cove. Williams and Stevens (1969) estimated that the Bradore section east of Lighthouse Cove may be between 90 and 120 m thick, but did not measure the section. The Round Head Cove estimate is probably the more closely controlled and falls within limits set by estimates to the southwest (45 to 60 m) and to the west (85 m).

The Bradore Formation in the three large central fault blocks at the south end of Belle Isle dips gently southward from 5 to 30°. No minor folds were seen, except for one in the lens between flows at Barbers Cove. The fold is overturned to the northwest and plunges southwest. Locally the beds are fractured and cut by minor faults. Smaller remnants of Bradore arkose are preserved along major faults above White Point and north of White Point Cove. These are severely fractured and locally veined by white quartz.

Age relations. The lower beds of the Forteau Formation (Devils Cove Formation at Canada Bay), which overlie the Bradore Formation, are fossiliferous and are known to be Early Cambrian on both sides of the Strait of Belle Isle and on both east and west flanks of the northern Long Range

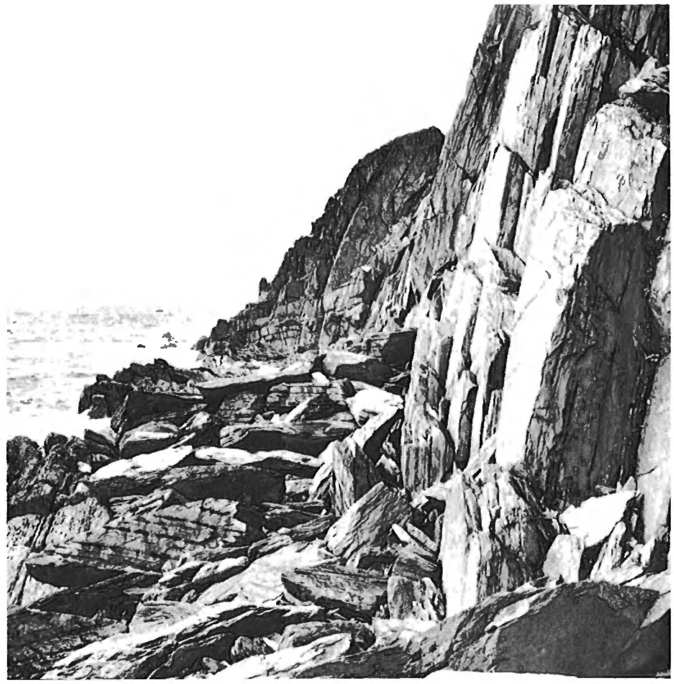


Figure 39. View looking north showing cleavage and bedding in the Forteau Formation just north of White Point, Belle Isle. In the foreground cleavage and bedding are approximately vertical but bedding bends over to the east near top of the outcrop. (GSC 160190).

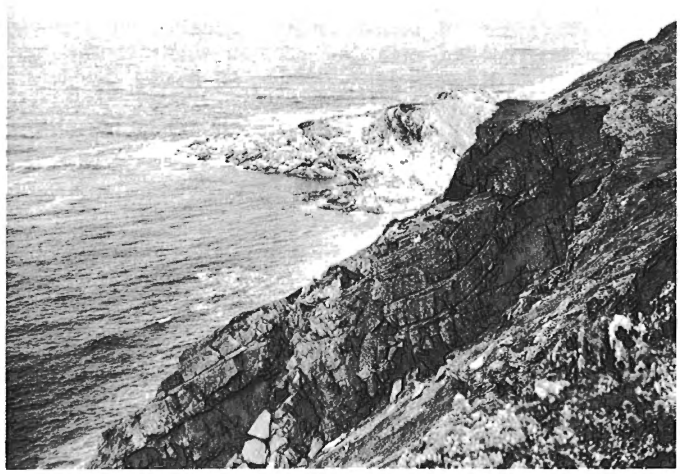


Figure 40. View looking north at White Point, Belle Isle. Contorted beds of the White Point Formation are visible in the background. Note the eastward dipping cleavage on the west limb of the anticline in the Forteau shale in the foreground. (GSC 160188).

(North, 1971). Locally these beds are late Early Cambrian (Otter Cove, Palmer in North (1971); Belle Isle, Fritz in Williams and Stevens, 1969). The Bradore Formation is therefore pre-late Early Cambrian. Vertical tube-like structures (*Scolithus linearis*) are abundant within parts of the Bradore Formation, but no other fossil remains are known.

Forteau Formation

The Forteau Formation was examined during the present work only on Belle Isle, and locally in the Canada Bay area in conjunction with mapping of the Precambrian rocks. The formation was defined by Schuchert and Dunbar (1934) with a type section near Forteau, Labrador, where a maximum of 55 m of variegated shale, archaeocyathid reefs, and sandy and oolitic limestone were reported. For more complete descriptions the reader is referred to the accompanying reports by L.M. Cumming, and by H. Williams and W.R. Smyth as well as Williams and Stevens (1969).

On Belle Isle the Forteau Formation consists chiefly of grey shale which, where deformed, displays a well developed parting parallel to bedding but elsewhere has a strong slaty cleavage. The uppermost arkosic beds of the Bradore Formaton and the lowermost shaly beds of the Forteau Formation contain minor beds of brown-weathering carbonate, forming a conformable sequence. These pass upwards into nodular shale and eventually into grey shale free of carbonate. The upper beds of the formation have been removed by erosion and faulting.

The Forteau Formation, which comprises the least competent of the supracrustal rocks on Belle Isle, has been strongly deformed where subjected to compression or shearing. In the fault blocks between Round Head and Scotswood Cove it is largely undisturbed and dips gently south in the east and southwest in the west. Near the east margin of each block, however, the formation is contorted and has particularly well developed slaty cleavage. South-plunging minor folds are evident near the east contact east of Round Head. At White Point the Forteau Formation forms a complex north-northeast-trending, nearly horizontal, westward overturned anticline (Fig. 39). Bedding defined by carbonate nodules is everywhere accompanied by, and in most places cut by, well developed east-dipping to vertical axial plane cleavage. At the south end of Lark Harbour, folds of some 12 m wave length have well developed axial plane cleavage dipping east at 60°. At the northeast end of Lark Island similar folds plunge 40° east-northeast, whereas on the outer northwestern island they plunge only 15° northeast. The dip of easterly dipping cleavage increases slightly as the plunge of the folds steepens northward.

White Point Formation

The White Point Formation (Williams and Stevens, 1969) is exposed only at the southwest of Belle Isle where it occupies a few thousand square metres on a small fault-bounded peninsula (Fig. 40). The formation consists mostly of white - to cream- or grey-weathering, pale grey siltstone, chert, and fragmental limestone. The strata lie against a major fault to the east and are deformed and internally faulted. Bedding in the southern part of the exposure appears to have an overall south dip of about 30°. The White Point Formation is believed to be a younger part of the Forteau Formation.

METAMORPHISM

Introduction

The Precambrian metamorphic rocks within the map area have been divided into three metamorphic terranes of different metamorphic grade, each terrane reflecting different textures and mineral assemblages. These terranes are defined as follows (see also Fig. 45 and 56):

(1) The granulite terrane of metamorphism comprises a region north of the Strait of Belle Isle between Bradore Bay and Red Bay, and extends north-northwest beyond the northern limit of the map area. The south margin of the terrane is close to the coast.

(2) The greenschist terrane of metamorphism comprises a region along the east coast of the Long Range extending inland some 27 km from the coast at Fourché Harbour. Farther north the western boundary is less well known but appears to veer northeastward near Cloud River. Belle Isle is included within this terrane.

(3) The amphibolite terrane of metamorphism comprises the remaining regions where basement gneiss is exposed.

Early granulite facies metamorphism is evident in gneiss of the western part of the map area. It is clearly imprinted about the anorthosite suite intrusions in the granulite terrane, but in the Long Range amphibolite terrane only traces of granulite facies metamorphism are present and these cannot be related spatially to any group of major intrusions. Nevertheless minor metagabbro intrusions, thought to form part of the anorthosite suite, are present throughout the area where this metamorphism is evident. The difference between the two terranes is thought to have arisen because the gneiss in the western Long Range (here included in the amphibolite terrane) only barely reached granulite facies conditions.

Amphibolite facies metamorphism is evident in the eastern part of the map area in the amphibolite terrane of Labrador and Quebec where it is presumably of the same age as granulite facies metamphism to the west. The metamorphic grade attained is lower than in the granulite terrane because the amphibolite terrane lies beyond the zone of major mangerite and hornblende granite intrusions, and metagabbro bodies may be less abundant than they are farther south and west. In the eastern Long Range and on Belle Isle evidence of the metamorphic grade attained during the period corresponding to high-grade metamorphism farther west has been obscured by later metamorphic overprints.

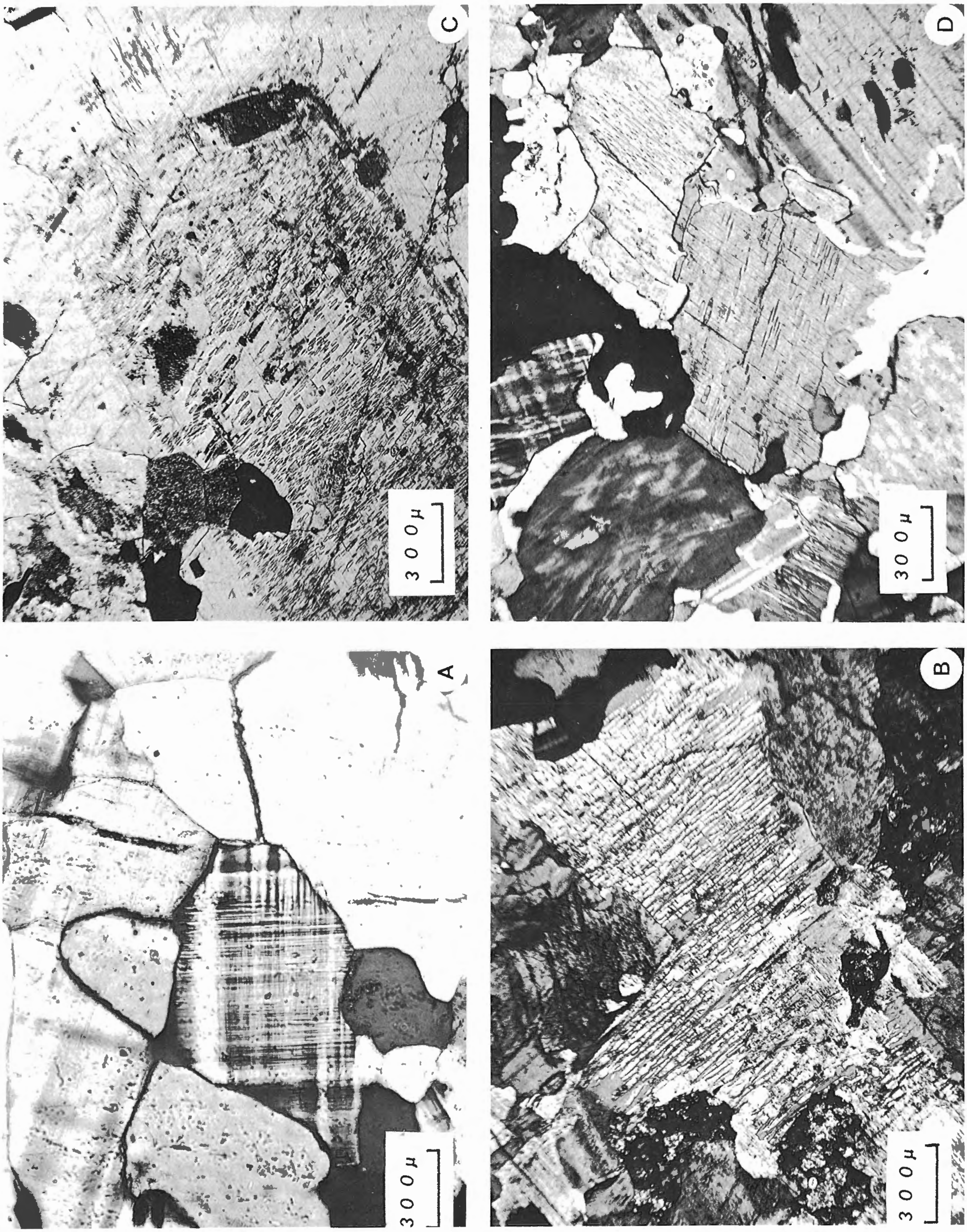
Local later lower amphibolite facies metamorphism is evident about the massive and megacrystic biotite granite plutons that postdate the highest grade metamorphism. These intrusions are much more extensive in the Long Range than in Labrador and Quebec and it is possible that the sporadic preservation of evidence of earlier granulite facies metamorphism is partly due to retrogression following their emplacement.

Retrograde greenschist facies metamorphism predominates in the eastern part of the map area, where it occurred in at least two phases. The first preceded emplacement of the Long Range dyke swarm, and in the Long Range may have accompanied folding of the southern part of the Hooping Harbour megacrystic pluton during the latter stages of its emplacement. The second postdated the Lighthouse Cove basalt and probably the Forteau (Devils Cove) Formation (age of fossils about 540 Ma) as well. The youngest K-Ar biotite age from the gneiss along the east coast 434±18 Ma) indicates that the latest metamorphism took place before the end of the Ordovician.

Granulite terrane of metamorphism

Mineral assemblages observed in the granulite terrane of Labrador and Quebec are listed in Table 8. The presence of two pyroxenes, and the absence of hornblende-garnet assemblages in the metagabbro, together with the scattered occurrence of hypersthene in the gneiss are characteristic of granulite facies metamorphism (Wynne-Edwards, 1971). The microcline-sillimanite-cordierite-pyralspite assemblages of the pelitic rocks are to be expected in either upper amphibolite or lower granulite subfacies. These data and the absence of clinopyroxene-pyralspite assemblages in the gneiss suggest that they belong to the lower part of the granulite facies.

The granulite terrane was not delineated in the field because of the relative scarcity of megascopic criteria upon which rocks in this facies could be recognized, true granulite texture, and the greenish to buff-brownish colour typical of many regions of granulite facies metamorphism being only locally apparent. Thin section study revealed that characteristic mineral assemblages (containing hypersthene) are sporadic particularly in the more leucocratic gneiss. It was found, however, that almost all gneiss within areas where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independent criteria of granulite facies metamorphism were available contained alkali feldspar characterized by irregular patchy submesoperthitic to mesoperthitic intergrowths of albite-oligoclase and microcline (or apparently untwinned potash feldspar) (Fig. 41). Furthermore, the distribution of such intergrowths was found to define a region about the mangerite and hornblende granite plutons beyond which such intergrowths were absent. Hence, although the position of the granulite facies isograd shown on the map may not exactly correspond to that representing the first appearance of hypersthene, it is probably parallel to the hypersthene isograd and not far separated from it.

Amphibolite terrane of metamorphism of Labrador and Quebec

The amphibolite terrane of Labrador and Quebec lies chiefly east of the granulite terrane (east of Red Bay), but the gneiss along the southwest margin of the map area in Quebec, north of Bradore Bay, is also free of hypersthene and mesoperthitic to submesoperthitic alkali feldspar, and has been included in the amphibolite terrane. The metagabbro at Lourdes du Blanc Sablon contains both hypersthene and clinopyroxene, but these are heavily armoured by hornblende containing local garnet. Quartz-rich pelitic bands at Bradore Bay contain sillimanite-biotite and sillimanite-microcline-garnet-biotite but no hypersthene or mesoperthitic to submesoperthitic intergrowths. This region therefore may have reached granulite facies conditions, and has undergone later retrogression more severe than that effecting the rocks to the northeast. Such retrogression may be related to plutonism that accompanied emplacement of the granodiorite stock west of Bradore Bay. Except perhaps for conditions responsible for the development of garnet in amphibole coronas about hypersthene in the metagabbro, the gneiss of this region may have had a metamorphic history similar to gneiss of the amphibolite terrane of the western Long Range. A fringe of gneiss along the coast between Pinware and Red Bay includes some rocks in which the alkali feldspar is only perthitic, intermixed with rocks in which submesoperthite is present. This gneiss may therefore be marginal between granulite and amphibolite facies terranes.

East of the granulite terrane the gneiss is free of hypersthene and submesoperthitic to mesoperthitic alkali feldspar. In the metagabbro hornblende-plagioclase is commonly accompanied by garnet, anthophyllite, or cummingtonite, assemblages characteristic of the amphibolite facies (Winkler, 1967). Pelitic rocks were not observed in this part of the amphibolite terrane, but a single calc-silicate gneiss occurrence contains prominent diopiside-plagioclase-hornblende presumably consistent with amphibolite facies metamorphism, although later alteration has resulted in crystallization of epidote-oligoclase. Minerals common in greenschist terranes such as albite, epidote, muscovite, and chlorite are widespread, but are generally not abundant. Albite is present only in the leucocratic gneiss where hornblende, epidote, or other common calcium-bearing minerals are typically absent. Muscovite, chlorite, and particularly epidote are, however, apparently in equilibrium with the surrounding minerals and locally are well crystallized in contrast to the prevailing anhedral ragged form in the Long Range. Thus the rocks of the eastern amphibolite terrane appear to have reached conditions of amphibolite facies metamorphism but no higher. Locally only greenschist facies conditions may have been reached or, more likely, the greenschist facies mineral assemblages may be due to more complete retrogression of rocks that had reached amphibolite facies conditions but had not been desiccated by earlier granulite facies metamorphism. The development of weakly perthitic rims on submesoperthitic cores in microcline along the east margin of the granulite terrane (Fig. 41C) suggests that lower grade conditions followed granulite facies metamorphism.

Table 8. Common mineral assemblages of the granulite terrane

Leucocratic	Mesocratic	Melanocratic	Metagabbro	Quartz-rich	Pelitic	Calc-silicate
qtz-olig-Ks	*qtz-olig-Ks*	*qtz-olig-Ks*	*hbl-and*	*qtz-olig-Ks*	*qtz-Ks-sill*	*and-qtz-Ks*
-bio	-bio	-hbl-bio	-opx-cpx-bio	bio-hbl	cord-gar-(spin)	cpx
-hbl	-bio-hbl	-bio-cpx	-opx-cpx-bio-Ks		gar-bio-(spin)	
-bio-hbl	-bio-gar		-opx-cpx	*qtz-ab-Ks*		*qtz-Ks-lab*
-bio-opx	-bio-hbl-cpx		-cpx			gar-sill-ves
	-bio-opx	*qtz-olig*		*qtz-Ks*		
		hbl-bio-opx-cpx		bio		
	olig-Ks					
	bio-hbl					

qtz: quartz; ab: albite; olig: oligoclase; and: andesine; lab: labradorite; Ks: potassic feldspar; bio: biotite; hbl: hornblende; opx: hypersthene; cpx: clinopyroxene; sill: sillimanite; cord: cordierite; gar: garnet; spin: spinel (not in contact with quartz); ves: vesuvianite.

Figure 41. Perthitic textures in gneiss from Labrador and Quebec.
A) Clear microcline with only traces of fine perthite, characteristic of the amphibolite terrane. Crossed nicols (GSC 201902-Z).
B) Irregular submesoperthitic intergrowth within the granulite terrane. Crossed nicols (GSC 201901-F).
C) Submesoperthitic euhedral core within a nonperthitic anhedral rim in gneiss near the east margin of the granulite terrane. Polarized light (GSC 201902-N).
D) Patchy perthite in gneiss near the eastern margin of the granulite terrane. Crossed nicols (GSC 201901-G).

Amphibolite terrane of metamorphism of the Long Range

Pelitic rocks, though occurring in small bodies, are widely distributed through the northern Long Range. In the amphibolite terrane these rocks are characterized by microcline and sillimanite with either cordierite or pyralspite garnet present at most localities. Rarely do both cordierite and pyralspite occur together. These assemblages suggest that the pelitic rocks have almost everywhere reached the sillimanite-cordierite-orthoclase-almandine subfacies of the amphibolite facies of Winkler (1967). They do not however preclude the attainment of conditions of the lower part of the granulite facies.

The lower thermal limit of the granulite facies is defined as occurring at the first appearance of hypersthene in basic rocks (Wynne-Edwards, 1971). Thus the presence of hypersthene remnants in two slightly calcareous mesocratic gneissic units in the amphibolite terrane of the Long Range indicates that these rocks did reach temperatures corresponding to the granulite facies. Remnants of hypersthene in amphibolite at widely scattered localities (see Fig. 15) are further suggestive of granulite facies conditions, provided that such hypersthene was not formed during crystallization of the original basic magma. One occurrence of true mesoperthite and scattered occurrences of irregular submesoperthitic intergrowths in alkali feldspar, by comparison with the granulite terrane in Labrador and Quebec, provide additional indications of early granulite facies metamorphism. The combined evidence of remnant mineral assemblages and textures from rocks of different compositions thus suggests that metamorphic conditions similar to those of the granulite terrane of Labrador and Quebec were attained in the Long Range. However, the greater abundance of amphibole in amphibolite, the absence of hypersthene remnants in all gneiss except for a few calcareous ones, and the sporadic occurrence of the submesoperthitic intergrowths in the Long Range suggests that the metamorphic history of the two areas was not quite the same. These features suggest either that granulite facies metamorphism in the Long Range was less severe than it was north of the Strait of Belle Isle, thus permitting the rocks to remain in a more hydrated condition facilitating the appearance of retrogression, or some retrogressive event such as widespread intrusion of the massive and megacrystic biotite granite has been effective in partly rehydrating the gneiss. The former alternative is likely the most important because:

(1) In the granulite terrane, intrusions of the anorthosite suite, to which granulite facies metamorphism is spatially related, are large and the suite includes both basic and acidic members. In the Long Range amphibolite terrane, on the other hand, members of this intrusive suite are confined to small metagabbro bodies, and it seems reasonable to expect that metamorphism related to this suite would be less severe.

(2) In the granulite terrane, hypersthene is locally present in leucocratic to melanocratic gneiss as well as in the metagabbro, whereas in the Long Range it is present only in certain low-potash calcareous gneiss and in metagabbro. Wynne-Edwards (1971) suggested that, at the onset of granulite facies conditions, orthopyroxene fails to appear in biotite-bearing quartzofeldspathic gneiss for some considerable interval of the metamorphic progression after it has appeared in the accompanying basic rocks. The distribution of hypersthene in the amphibolite terrane of the Long Range therefore suggests that conditions of granulite facies metamorphism have only barely been reached.

(3) In the granulite terrane, hypersthene in metagabbro is relatively abundant whereas in the Long Range amphibolite terrane it is mostly absent, and where present is typically scarce and corroded. On the other hand sillimanite in the microcline-bearing pelitic rocks of either terrane is only locally extensively altered to muscovite. This suggests that late rehydration has been of only local significance and that some other reason must be sought for the relative scarcity of hypersthene in the Long Range amphibolite terrane.

(4) At least one small body of anorthosite is present in the southern part of the Long Range (Baird, 1960a), south of the map area, and gneiss containing abundant unaltered hypersthene was observed not far from the contact of the Lake Michel megacrystic pluton. These features suggest that gneiss both north and south of the Long Range amphibolite terrane reached granulite facies conditions and that emplacement of the megacrystic granitic plutons did not cause pervasive regional amphibolite facies retrogression.

(5) Cummingtonite and anthophyllite are typically absent in the granulite facies (Winkler, 1967). The rare occurrence of these minerals in the gneiss of the Long Range might have resulted from retrogression after granulite facies metamorphism, perhaps during emplacement of the biotite-bearing granitic plutons; however, as no relicts of granulite facies mineral assemblages were found in gneiss displaying these minerals in the Long Range, it is more likely that such rocks represent a few areas that never quite reached granulite facies conditions.

Retrogression of the pelitic gneiss in the amphibolite terrane is indicated by the occurrence of andalusite (Fig. 42) in these rocks locally, where they approach the massive and megacrystic biotite granite plutons. Its occurrence near the Leg Pond pluton in patches with late, green biotite isolated from microcline, and surrounded by sillimanite interspersed with brown biotite, suggests that equilibrium was not attained even over small volumes. Nevertheless, the presence of andalusite indicates that retrogressive metamorphism was of the Abukuma type, and may have occurred at lower pressures than are normally thought to be attained during granulite facies metamorphism (distinctive high-pressure assemblages of the granulite facies including kyanite have not been found in the Long Range). The occurrence of andalusite-microcline in contact with one another in an inclusion within the Cloud River pluton suggests even lower pressure, high temperature (contact) metamorphism.

Some of the retrogression evident in gneiss more remote from the biotite-bearing granite plutons may also be related to their emplacement. Such retrogression likely includes some alteration of sillimanite to muscovite, cordierite to chlorite or sericite, corrosion of hypersthene, crystallization of blue-green hornblende, and sausseritization of calcic plagioclase.

The absence of garnet (except at one locality on Canada Bay) in the many amphibolites examined from the Long Range is of interest because garnet is common in similar rocks of the amphibolite terrane in Labrador and Quebec. Garnet is common in rocks of Barovian (high pressure) metamorphic facies series at grades above middle greenschist facies, but it is more restricted in rocks of the Abukuma (low pressure) facies series where it occurs typically in rocks of the middle and upper amphibolite facies (Winkler, 1967). Thus the nonappearance of garnet in the metagabbro and amphibolite of the Long Range is consistent with the presence of andalusite in pelitic rocks close to the Leg Pond pluton because both suggest that overprint metamorphism was of the Abukuma facies series. The andalusite suggests that the maximum grade reached was lower amphibolite facies and the absence of garnet in the metagabbro is consistent with any lower grade (assuming that garnets were not formed during early granulite facies metamorphism, an assumption which seems to be true in the granulite terrane).

Greenschist terrane of metamorphism

Greenschist terrane within the map area comprises the gneiss on Belle Isle and the gneiss and granite in the belt that extends along the eastern margin of the Long Range reaching some 27 km inland from Fourché Harbour. The precise western limit of alteration is difficult to determine because, south of Cloud River, this limit lies within the area of drift cover associated with the ice divide, and north of that pluton, alteration of the gneiss is patchy and apparently decreases in intensity. Farther north the boundary lies to the west of Belle Isle and beneath the Strait of Belle Isle. Diabase dykes intrusive into the gneiss and granite of the greenschist terrane are also affected by greenschist facies metamorphism, but in the dykes the intensity of alteration, as suggested by the proportion of remaining unaltered pyroxene, increases toward the east coast, and the western limit of severe alteration lies to the east of the western limit of overprint in the gneiss.

Greenschist facies metamorphism of gneiss and granite in northern Long Range is recognizable chiefly by the increase in frequency and proportion of epidote and, in the southern part of the greenschist terrane, by recrystallization of biotite. Early medium-grained biotite is typically oriented parallel to schistosity, and commonly contains acicular rutile (?) inclusions in trigonal patterns. Fine-grained epidote or other alteration products commonly cluster at the ends of biotite crystals, along cleavage or form zig-zag patterns across cleavage. Late biotite is fine grained, clean, and in clusters of randomly oriented crystals (Fig. 43).

Pelitic rocks in the greenschist terrane are evident along the northern and western margins of the satellite pluton. They occur also as thin bands (too thin to be shown on the map) associated with bands of quartz-rich gneiss along the western and northern margins of the Hooping Harbour pluton. These rocks are predominantly biotite-oligoclase-quartz gneiss in which sillimanite-microcline persist. Sillimanite is commonly partly altered to muscovite and muscovite is present as porphyroblasts or as interleaves in biotite. No trace of cordierite remains, but garnet (apparently in equilibrium with the surrounding mineral assemblages) was observed in two thin sections. No andalusite was observed. North of the Hooping Harbour pluton, where thin bands of mesocratic to melanocratic micaceous gneiss (probably pelitic gneiss) and quartz-rich gneiss are associated near the pluton contact, no trace of sillimanite was found and chlorite and muscovite are more abundant. Similar lithological associations were not observed on the east side of the pluton, and thin bands of pelitic rocks, if they occur, were not recognized.

The abundance of alteration products in the granite and gneiss is difficult to measure because much of this alteration is fine grained and the abundance may vary with the composition of the host rock. Estimates of frequency of the minerals involved are readily made and suffer less from these restrictions. Such estimates were made for epidote, chlorite, and muscovite for some 390 thin sections studied from the Long Range by plotting the location of each section together with a symbol to represent whether it contained the mineral in question or not. A point frequency map was then constructed showing the percentage of specimens carrying the specified mineral in successive 6.4-km squares, each square overlapping its neighbour by 50 per cent. The maps were then contoured at 20 per cent intervals (Fig. 44). Although the distribution of thin section data is irregular and the contours influenced to some extent by arbitrary placing of the point grid, the resultant maps may be expected to reveal approximate regional differences in the frequency of the three minerals concerned.

The frequency distribution of epidote, as might be anticipated, outlines the boundaries of the greenschist facies metamorphic terrane. Muscovite and chlorite, however, do not show a parallel distribution possibly because the greenschist overprint was superimposed upon rocks that had presumably mostly already undergone granulite facies metamorphism and were consequently relatively dry. Under such conditions crystallization of the least hydrous alteration product (epidote) would be favoured.

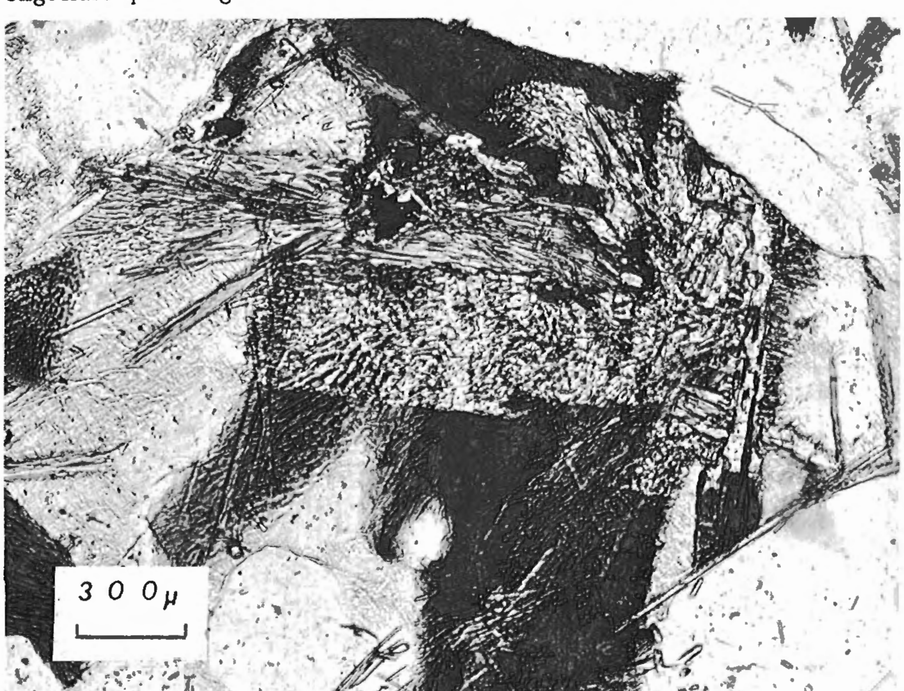
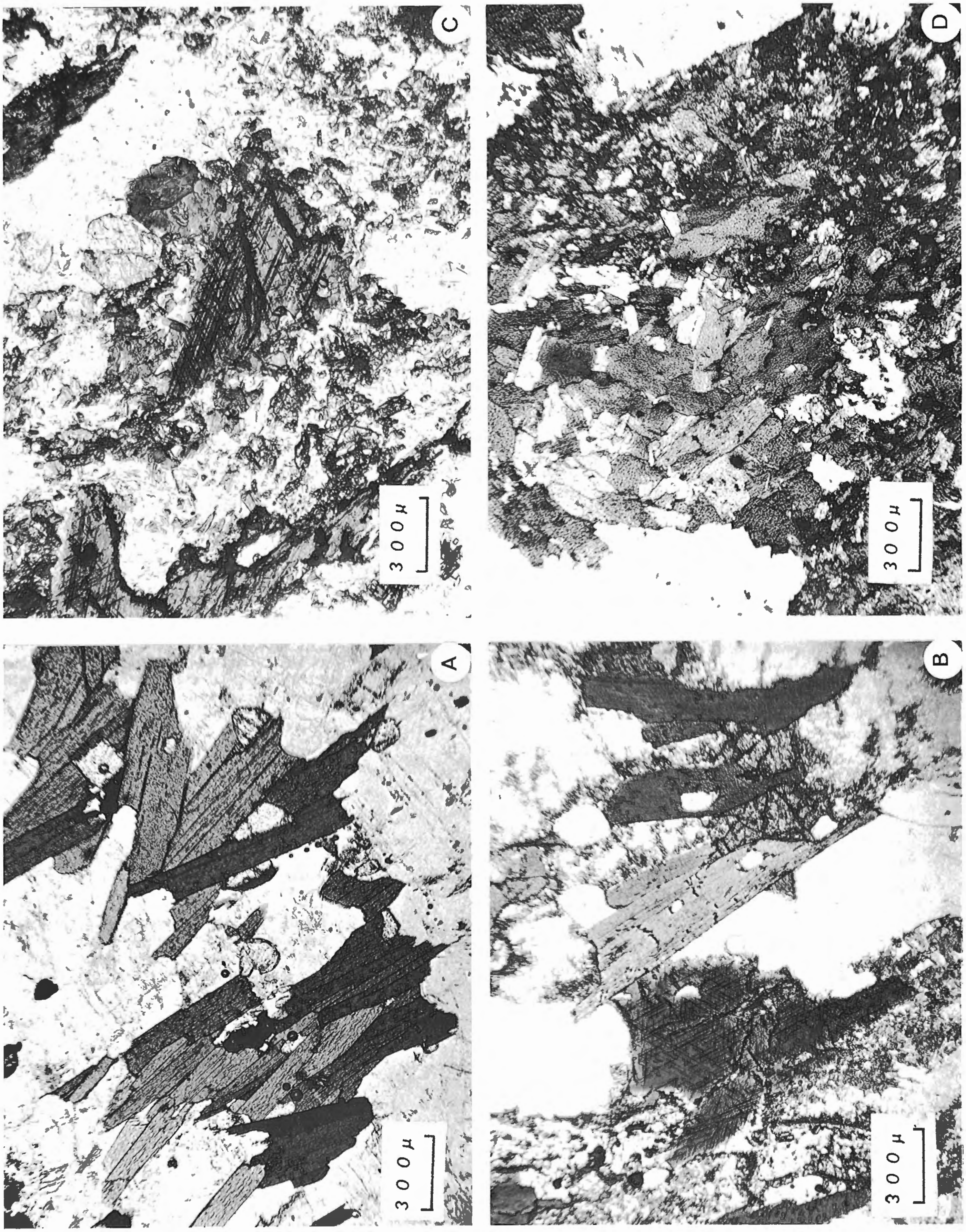


Figure 42. Late andalusite (centre) associated with greenish biotite and early sillimanite in pelitic gneiss near Leg Pond pluton. Polarized light (GSC 201901-U).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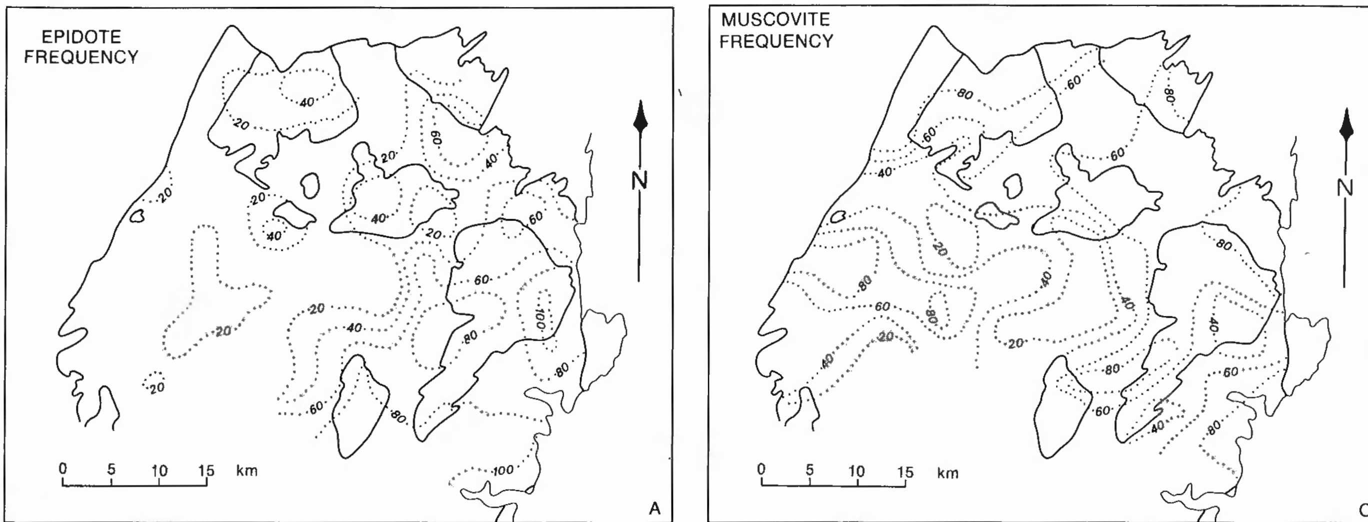


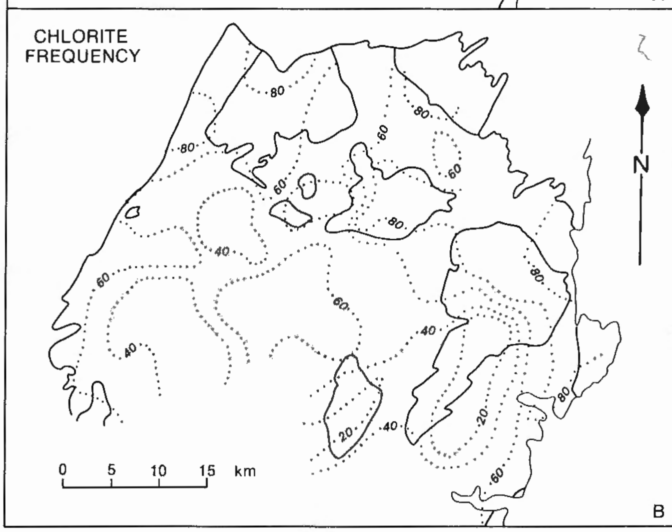


Figure 44. Sketch-maps showing epidote, chlorite, and muscovite frequency in thin sections of granites and gneisses in the northern Long Range. Contours express an estimate of the percentage of samples from a given 6.4-km-square area that can be expected to contain the mineral concerned. In computing muscovite frequency only thin sections containing muscovite or potash feldspar were considered. The outlines of the granitic plutons and of the Grenville inlier are also shown.

Epidote, the least hydrous of the alteration products, shows a frequency distribution clearly related to greenschist facies overprint in the eastern part of the map area. The pattern shown by the more hydrous alteration products, chlorite and muscovite, is less distinct perhaps partly because earlier high-grade metamorphism had reduced the water content of the rocks to the point where crystallization of these minerals was restricted.

The frequency distributions of muscovite and chlorite are somewhat similar, with highest values concentrated about the margins of the Grenville inlier and with lower values in the interior. This pattern may be related to emplacement of the megacrystic granitic rocks, to introduction of water into the gneiss near the east coast during late metamorphism that followed emplacement of the diabase dykes, to the proximity of the rocks to the surface at the time of last metamorphism, and/or to the late Hadrynian unconformity.

Alteration of the diabase dykes of the Long Range dyke swarm intrusive into the gneiss of the greenschist terrane increases toward the coast (this alteration pattern was first reported by Foley (1937) for dykes near Great Harbour Deep). Severely altered dykes are in a belt along the east coast, the western limit of which extends approximately southwest from Otter Cove to a point some 12 km inland from Hooping Harbour. Farther south the belt expands and extends some 20 km inland from the coastline at Fourché Harbour, and some 25 km inland at Great Harbour Deep. Alteration of dykes west of this belt decreases to very slight at the west margin of the dyke swarm. Variation in intensity of alteration of the dykes based on estimation of alteration in thin sections is shown in Figures 45 and 46.

Alteration of the diabase dykes is characterized by retrogression of plagioclase (labradorite) to albite and saussurite or to albite, saussurite and sericite, and of pyroxene to amphibole and chlorite. At the same time minor biotite is formed along with a fine-grained mineral characterized by radiating crystallites of negative elongation (possibly prehnite). In the most severely altered dykes the original mineralogy is entirely converted to these alteration products.

Alteration of the dykes clearly postdates at least part of the greenschist facies metamorphism of the gneiss because recrystallization of biotite in the gneiss exends well to the west of severe alteration of the dykes into an area where the dykes are slightly altered to unaltered.

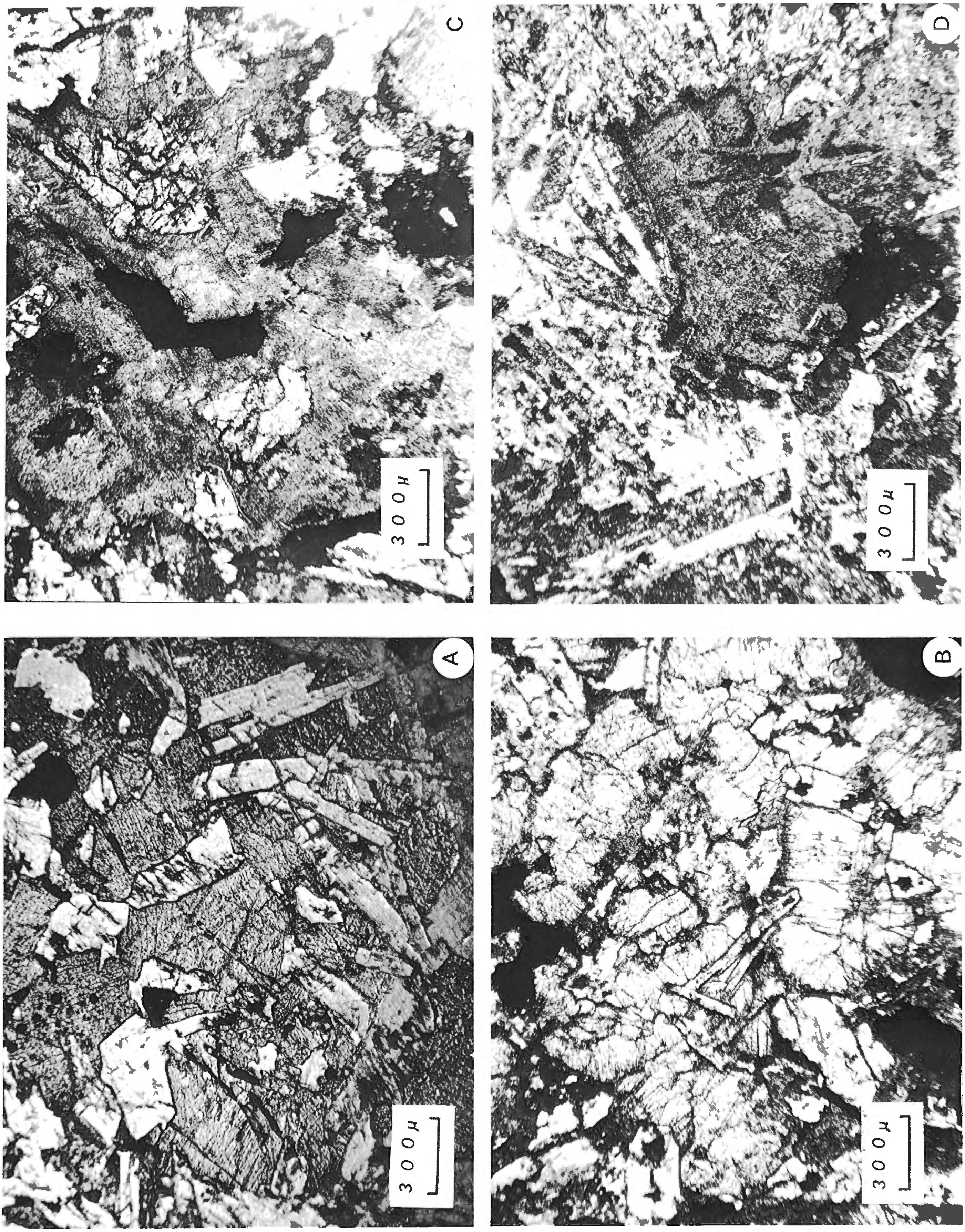
Figure 43. Alteration and secondary growth-textures in biotite in the greenschist terrane.

A) Unaltered biotite common in the amphibolite terrane. Polarized light (GSC 201901-K).

B) Biotite showing concentration of alteration products along cleavage and in zig-zag discontinuities across cleavage. Minimal secondary biotite is present. Polarized light (GSC 201901-M).

C) Early, coarse-grained, inclusion-filled biotite surrounded by late fine-grained clear biotite. Polarized light (GSC 201901-S).

D) Early biotite completely recrystallized and outlined by epidote with associated patches of clear late biotite. Polarized light (GSC 201901-J).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Thus it might be assumed that alteration of the dykes is of deuteric origin and their pattern of alteration due to progressive increase in depth of burial toward the coast at the time the dykes were emplaced. Other evidence however, that late greenschist facies metamorphism was restricted to the eastern part of the inlier exists. The presence of secondary muscovite in the regolith at Canada Bay suggests that at least the basal part of the Bradore Formation in this region has undergone low-grade metamorphism. The absence of similar muscovite and the friable nature of the regolith west of the highlands and elsewhere, suggests that this metamorphism was confined to the northeastern periphery of the Long Range. Effects of metamorphism within the Bradore Formation are difficult to recognize because fine-grained argillaceous material is rarely concentrated and detrital muscovite is common in the formation about Canada Bay. Siliceous argillaceous limestone (late Lower Cambrian), which overlies the Bradore Formation north of Otter Cove, shows no evident metamorphism, but markedly similar sheared siliceous argillaceous limestone immediately above the Bradore arkose at Wild Cove contains disseminated secondary biotite and chlorite. In both these limestone units scattered grains of clear albite were observed. It is therefore possible that the late greenschist facies metamorphic front evident between altered and unaltered diabase southwest of Canada Bay, lies within the Bradore Formation at Otter Cove and rises above the base of the Devils Cove Formation somewhere between Otter Cove and Wild Cove.

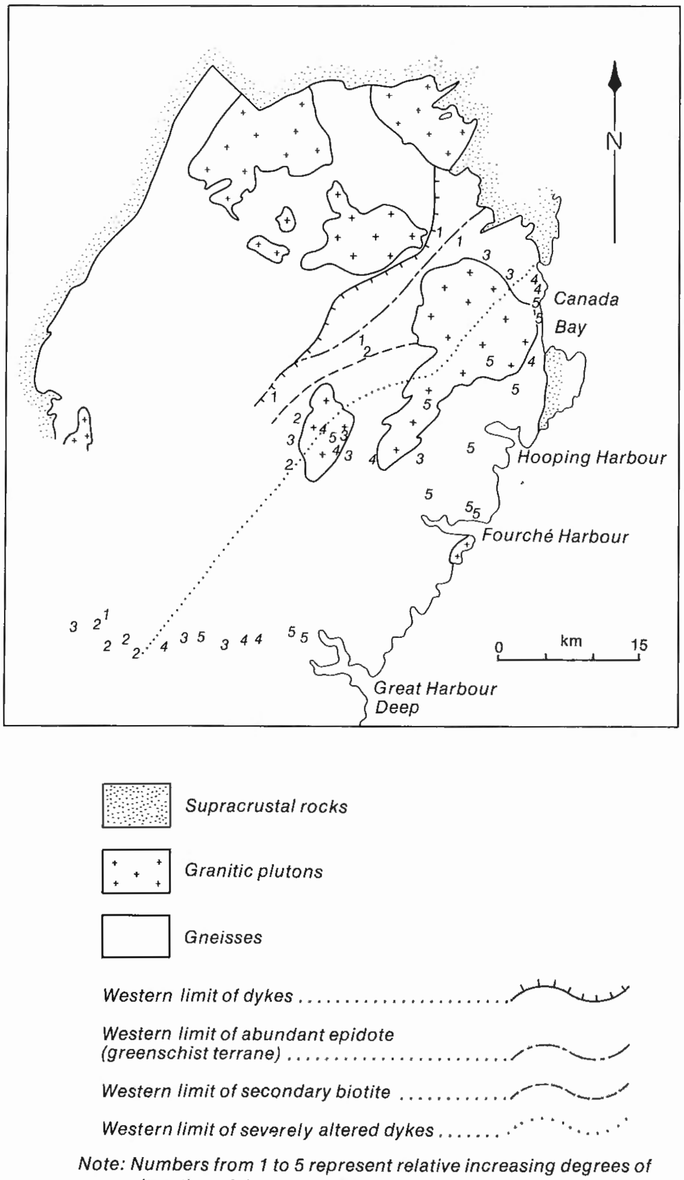


Figure 45. ***Map showing the outline of regions containing secondary epidote, biotite and severely altered diabase dykes.***

The gneiss on Belle Isle may be divided into two metamorphic zones (both in the greenschist facies) by an irregular line running northward roughly from near Beauty Cove to east of Wreck Bay. West of this line the predominant mineral assemblage is albite-chlorite-epidote (quartz-microcline), but remnants of partly chloritized olive-brown biotite exist and are prominent locally. To the east the predominant mineral assembalge is albite-biotite-epidote (quartz-microcline). Biotite is greenish, unaltered and in clusters of fine flakes similar to late biotite in the greenschist terrane of the Long Range. The predominance of albite in association with biotite or chlorite and epidote in both domains suggests that both have been subjected to greenschist facies metamorphic overprints. The textures of the mafic minerals indicate that the eastern zone represents the latest overprint.

Diabase dykes of the Long Range dyke swarm on Belle Isle show intense greenschist facies alteration within the eastern zone defined by late biotite in the gneiss. Pyroxene in dykes along the east coast has been completely altered to amphibole-chlorite, and remnants of pyroxene were observed locally inland. West of the biotite zone within the chlorite zone pyroxene is moderately to slightly altered except in those dykes that appear to be close to the late Hadrynian unconformity, where the pyroxene is intensely altered. Because all dykes are altered in the eastern part of the island it appears that the eastern biotite-zone overprint is younger than the Lighthouse Cove Formation. This is consistent with development of cleavage within the flows east of Barbers Cove. Alteration of the western dykes on the other hand is variable possibly because they had access to meteoric waters at the time of emplacement.

Summary and interpretation of radiometric dates

The age of emplacement of the Lake Michel megacrystic pluton has been interpreted as 1130 ± 90 Ma (λ^{87}Rb$=1.39\times10^{-11}$ yr^{-1}) by Pringle et al. (1971) on the basis of an Rb-Sr isochron. High-grade metamorphism of the gneiss surrounding the megacrystic plutons occurred before their emplacement, and hence deposition of the sediments from which the basement gneiss complex was derived occurred

Figure 46. Degrees of alteration of diabase dykes.

A) Unaltered diabase, equivalent to (1) in Figure 45. Polarized light (GSC 201902-A).

B) Slightly altered diabase shows selvages of amphibole about pyroxene, equivalent to (2) in Figure 45. Polarized light (GSC 201902).

C) Moderately altered diabase shows about equal proportions of pyroxene and amphibole alteration, equivalent to (3) in Figure 45. Polarized light (GSC 201902-Y).

D) Severely altered diabase shows complete alteration of pyroxene to amphibole, equivalent to (4 and 5) in Figure 45. Polarized light (GSC 201901-X).

before this date. As pointed out by Pringle et al. (1971), the date of emplacement approximates the age of the younger of two periods of plutonic activity (1125 Ma) suggested by Silver and Lumbers (1965) for the western part of the Grenville Province. This correlation remains appropriate in the light of the present report because the megacrystic granitic plutons postdate high-grade metamorphism and no granitic intrusions within the map area have been shown to be substantially younger than they are.

In view of the extensive intrusion of the metacrystic granite it is not surprising the K-Ar dates obtained from hornblende, muscovite, biotite, and whole rock samples are younger than the determined age of intrusion, (1130 Ma Pringle et al. 1971), of the megacrystic plutons, for regional metamorphism approaching amphibolite grade would be sufficient to expel earlier formed radiogenic argon. The K-Ar dates are listed in Table 9.

The hornblende K-Ar dates represent respectively an amphibolite band within the basement gneiss complex at Fourché Harbour, and a metagabbro body bearing a small percentage of remnant hypersthene from near the western margin of greenschist facies overprint (west of the severely altered diabase dykes). The dates are almost identical and suggest that postdyke metamorphism has had little effect upon argon retentivity in hornblende. The hornblende dates are both rather close to the biotite K-Ar dates west of the greenschist facies overprint. In view of the higher argon blocking temperature known to exist for hornblende than for biotite, this suggests either that cooling of the gneiss between argon blocking temperatures for these two minerals was rapid in the western part of the map area, or that gneiss in the eastern part of the area reached the hornblende argon blocking temperature later than those in the west. In view of the evidence for a very early (pre-dyke) greenschist facies metamorphic overprint in the eastern gneiss, the latter possibility should receive further investigation.

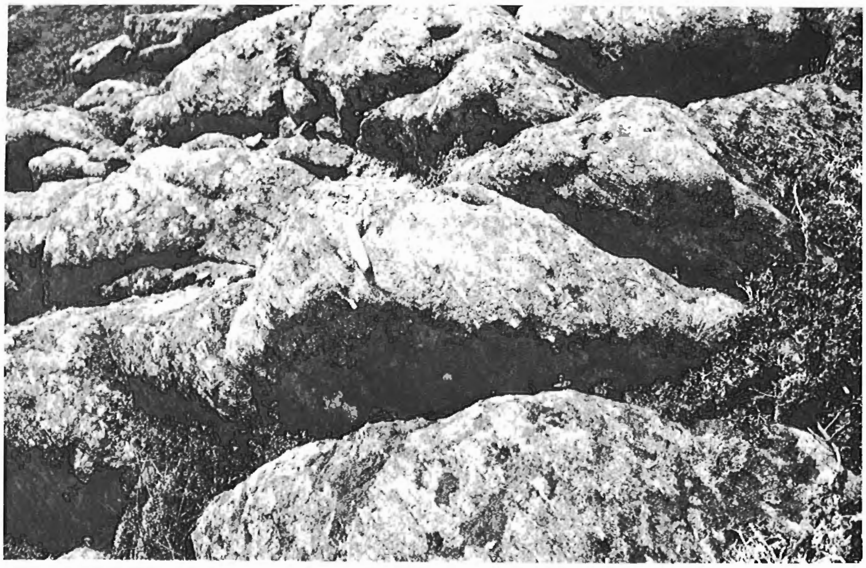


Figure 47. Sillimanite porphroblasts elongated parallel to the regional mineral lineation trend in pelitic gneiss northeast of Pikes Feeder Pond. (GSC 160141).

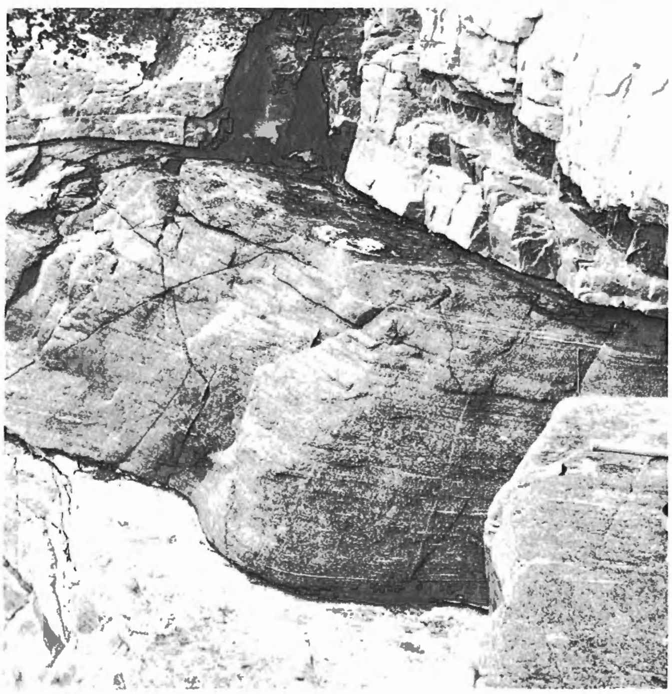


Figure 48. Feldspathic lenticles elongate parallel to regional lineation in a melanocratic gneiss band north of Hooping Harbour. (GSC 153341).

Figure 49. Clasts elongate parallel to regional mineral lineation in a thin melanocratic gneiss band (meta-conglomerate ?) west of Hooping Harbour. (GSC 153335).

Table 9. Summary of radiometric age determinations of rock samples from the Strait of Belle Isle area

	Rb-Sr isochron	K-Ar hornblende	K-Ar muscovite	K-Ar biotite	K-Ar whole rock
Henley Harbour Region					*Dykes* 920 ± 32 Ma 560 ± 21 Ma *Flows* (L.C.F.) 375 ± 100 Ma 411 ± 17 Ma 421 ± 17 Ma
Western Long Range	*Megacrystic granitic rocks* 1130 ± 90 Ma[1] $(\lambda 87_{Rb}=1.39\times10^{11\ \ 1})$			*Megacrystic granitic rocks* 960 ± 65 Ma 840 ± 20 Ma[1] *Massive granitic rocks* 945 ± 65 Ma	*Dykes* 805 ± 35 Ma[1] 751 ± 100 Ma 605 ± 10 Ma[2] *Flows* (L.C.F.) 413 ± 16 Ma 427 ± 17 Ma
Eastern Long Range		*Metagabbro* 903 ± 37 Ma 903 ± 38 Ma	*Pegmatite within Torrent Cove assemblage* 843 ± 24 Ma	*Gneiss complex* 434 ± 18 Ma 512 ± 20 Ma	*Dykes* 334 ± 100 Ma

[1] Pringle et al. (1971). [2] Stukas and Reynolds (1974), $^{40}Ar-^{39}Ar$ method. L.C.F. Lighthouse Cove Formation.

Other age determinations are by the geochronology laboratory of the Geological Survey of Canada, in Lowdon, J.A. et al., 1963, p. 118; Wanless, R.K., et al., 1965, p. 111; Wanless, R.K., et al., 1966, p. 96; Wanless, R.K., et al., 1968, p. 136; Wanless, R.K., et al., 1973, and Wanless, R.K., et al., 1974.

A single muscovite K-Ar date, 843 ± 24 Ma, is from a pegmatite intrusive into the Torrent Cove assemblage south of Torrent Cove. Feldspars are fractured and 5-cm muscovite crystals are locally bent indicating that the pegmatite has been deformed. Allowing for some difference in argon blocking temperatures for muscovite and hornblende, the date is similar to the dates obtained from hornblende within the greenschist terrane and suggests that the late (postdyke) metamorphism has not severely affected the radiometric age of the muscovite.

Two biotite K-Ar dates, 434 ± 18 Ma and 512 ± 20 Ma (Wanless et al., 1973), are from gneiss near the coast east of Williamsport, and from amphibolite at Squally Point, Fourché Harbour, respectively. The former date is from fine-grained, fresh, late biotite in a mesocratic gneiss band, whereas the latter is from schistose, medium-grained biotite from the same amphibolite band from which the Fourché Harbour hornblende date was obtained. The difference in these biotite dates may represent a difference in grain size and parent minerals from which the biotite was formed, or it may in part represent a real difference in the time of uplift of the rocks from which the samples were obtained. Both dates taken together are clearly younger than those obtained from the western part of the Grenville inlier (840 ± 20 Ma, Pringle et al. 1971; 960 ± 65 Ma and 945 ± 65 Ma, Lowdon et al., 1963) and they therefore suggest that postmetamorphism uplift along the east coast was much later than in the west. The radiometric dates yielded by biotite in the eastern gneiss correspond approximately to late Ordovician and late Cambrian. They provide a minimum age for the low-grade metamorphic event in the eastern diabase dykes, in the post-Grenville pre-Bradore regolith, and in the Forteau (Devils Cove) Formation at Wild Cove.

The results of K-Ar whole rock dating of the diabase dykes and flows of the Lighthouse Cove Formation (Table 9) suggest that the flows and some of the dykes were degassed about 420 Ma ago during the Ordovician orogeny of Newfoundland.

STRUCTURAL GEOLOGY

Structural Elements

Foliation

Banding from a millimetre to a metre or more wide is evident in the basement gneiss complex at many places throughout the map area. Bands of leucocratic gneiss up to 9-m thick are suggested locally by regularly repeated, joint-defined cuestas. Where well exposed, thin bands are commonly lenticular and discontinous. Locally, where quartz-rich gneiss or conglomerate bands are present and banding is continuous, it appears that original bedding is present. In no instance, however, have sedimentary structures such as crossbedding been recognized with confidence, and no stratigraphic top determinations were made. In many instances, banding has been completely disrupted by penetrative deformation and a more or less homogeneous schistose rock has been formed. Such intensely schistose rocks are most commonly mesocratic or melanocratic but may contain large lenses of schistose leucocratic or granitic gneiss.

Lineation

Mineral lineations, most commonly due to elongate platy patches of biotite, or to rodding of quartz grains, are well developed over extensive areas, but in some they are apparently absent, difficult to detect, or of patchy occurrence. Lineations in lineated pelitic gneiss are commonly expressed by sillimanite, and locally such sillimanite lineation is spectacular (Fig. 47). In places, feldspathic lenticles in melanocratic gneiss bands are elongate parallel to the regional lineation (Fig. 48). At one exposure west of Hooping Harbour fragments or cobbles of

Figure 50. 'Sine-wave' folding at Chateau Bay with three sets of folds, each differing in size by approximately an order of magnitude or more. The smallest folds are tiny crenulations visible on bedding surfaces but not obvious in the picture. Intermediate folds form linear features about a decimetre across in the vicinity of the hammer. Only one large scale fold is evident. (GSC 160125).

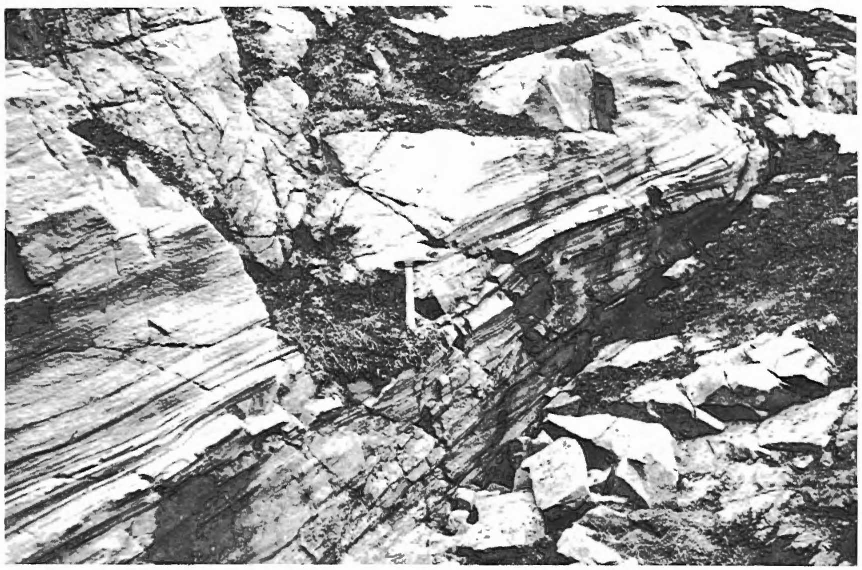


Figure 51. Steeply dipping gneiss bands show 'sine-wave' folding near Back Cove. (GSC 160092).

Figure 52. Asymetrical minor folds at the contact between a leucocratic gneiss band and melanocratic gneiss near Black Bay. (GSC 160119).

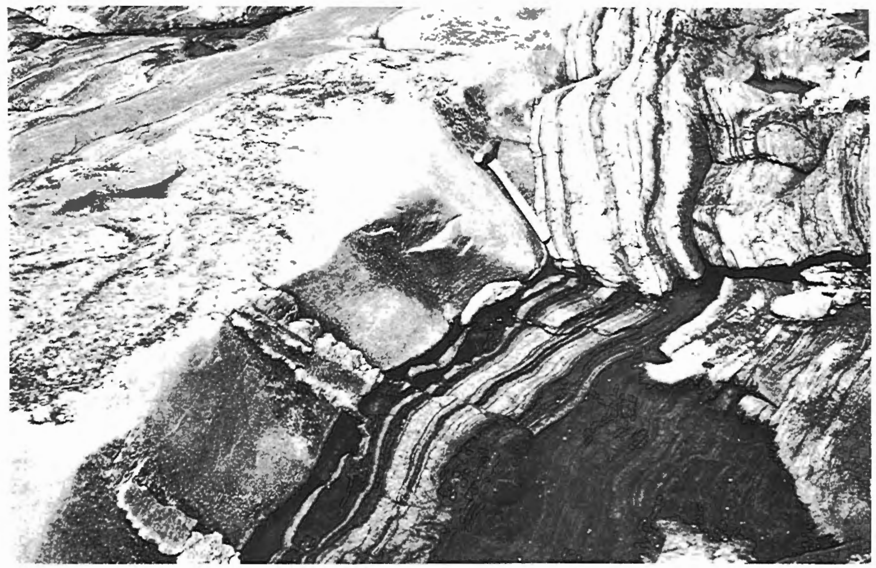


Figure 53. Steeply plunging 'sine-wave' folding near Black Bay. (GSC 160118).

fine-grained leucocratic rock in a melanocratic gneiss band 10 cm or more thick are highly elongate parallel to the mineral lineation in the surrounding leucocratic gneiss (Fig. 49). Mullion-like linear structures consisting of ribbons of biotite defining a somewhat sinuous plane are present in leucocratic gneiss in the core of the dome-like structure north of upper Torrent River. These biotite ribbons are oriented parallel or nearly parallel to quartz rodding and biotite lineations in the surrounding rock. In places minor quartz, or quartzofeldspathic veins display a fine crenulation either parallel to, or slightly divergent from the mineral lineation in the enclosing rock. Where both mineral lineation and foliation are present, the lineation normally lies within the foliation plane. However, locally, lineation observed at one point is not coplanar with foliation measured at a different point in the same outcrop. In some places, particularly along the east coast of the Great Northern Peninsula, penetrative mineral lineation is not obvious but a fine crenulation is present at contacts between leucocratic and melanocratic gneiss bands along which amphibole crystals show preferred orientation parallel to the crenulation axis.

Mineral lineation is thought to have been widespread and well developed in early folding. In later folds this lineation is apparently partly to wholly destroyed and new linear fabrics are only locally strongly developed.

Minor Folds

Minor folds are locally prominent within the basement gneiss complex, the most common type (particularly in Labrador and Quebec) being gentle sine-like or asymmetrical waves visible at contacts between leucocratic and melanocratic layers (Fig. 50, 51, 52 and 53). These are typically several centimetres or more in wavelength and may be only a few millimetres in amplitude. Similar folds are locally present in two or rarely three parallel series each differing in size by an order of magnitude or more (Fig. 50). Other minor folds are commonly fragmentary and in many parts of the map area are apparently scarce. In most minor folds axial plane foliation was not observed but locally, as in well exposed isoclinal minor folds at Wreck Bay, schistosity parallel to the limbs can be seen to continue uninterrupted through the fold crests. Where both minor folds and mineral lineation are present together they may be either nearly parallel or divergent. In the latter case the mineral lineation is warped about the fold axis indicating that the minor folds are the younger.

Joints

The attitudes of prominent joints in the Precambrian rocks were recorded. In Labrador and Quebec, note was taken of joints along which the country rocks were conspicuously oxidized. These joints, however, were found to fall within the general distribution of joints showing no alteration and without evident concentration. It is therefore assumed that alteration was late and took place along joints that happened to be locally favourable. Joints characterized by this alteration are thus not treated separately.

Joint measurements (Fig. 54) have been divided for comparative purposes into four geographic areas: (i) central Long Range; (ii) eastern Long Range lying within about 9 km of the east coast in a belt parallel to the Long Range dyke swarm; (iii) Labrador and Quebec; and (iv) Belle Isle.

Comparison of joints from the four areas shows that in each case there are concentrations of steeply dipping joints that strike northeasterly, whereas only the central Long Range and possibly Belle Isle have prominent northwesterly striking sets.

In the northern Long Range (Fig. 54A, B) the principal joint set strikes about 45°, parallel to the Long Range dyke swarm. There seems to be a slightly greater number of southeast-dipping joints in the central Long Range than of northwest-dipping joints in the eastern Long Range. The joint data are in harmony with measurements of dip made on the dykes and suggest that both joints and dykes may have a slightly fan-shaped distribution about a northeast-trending line some 8 to 11 km inland from the east coast.

In Labrador and Quebec, (Fig. 54C) there appears to be two major sets of steeply dipping joints, one striking slightly east of north, and a second nearly east.

The latter set of joints is dispersed toward a more northeasterly trend and may in fact represent two near parallel subsets. A few joints parallel to the Long Range dyke swarm may represent a weakly developed third set. Late basic dykes, though few in number, follow the two major joint sets suggesting that these joints predate the Long Range dyke swarm and are of Grenville age.

On Belle Isle (Fig. 54D), although the rocks are commonly highly fractured, regular jointing is rarely well developed. The principal joint set trends north-northeasterly roughly parallel to the dyke strike, but both northwesterly and southeasterly dipping joints are present on both sides of the island. The data are too few to indicate whether there is a slightly fan-shaped distribution as suggested in the northern Long Range.

Structural interpretation

Structural subdivisions of the basement complex

The basement gneisses are clearly severly deformed, but may show large areas within which foliation is shallowly to moderately dipping. Between these areas are zones of steeper dip that in part are due to later deformation and in part are thought to follow the boundaries of major fold structures. Subdivision of the map area into different structural regions (Fig. 55) is based on zones of steeply dipping foliation, granitic plutons, faults and areas of sparse outcrop. In the northern Long Range, where more detailed information is available, five regions are delineated; in Labrador and Quebec, two regions; and on Belle Isle, one region, although it includes parts of two structural domains. These regions are listed in Table 10, and their boundaries are shown in Figures 55 and 56. Structural data for each region are summarized in stereogram Figures 57A and 57B.

Table 10. Structural regions within the Precambrian rocks

Northern Long Range	Labrador and Quebec	Strait of Belle Isle
Upper Cloud River region Leg Pond region Northwest Brook region Hooping Harbour region East coast region	Henley Harbour region Red Bay region	Belle Isle region

Summary of folding

Folds in the map area trend predominantly northeast, and have developed during at least three periods of deformation. Northwest-trending folds occur locally and are different ages in different regions of the map area. In part they may have formed adjacent to rising granitic plutons.

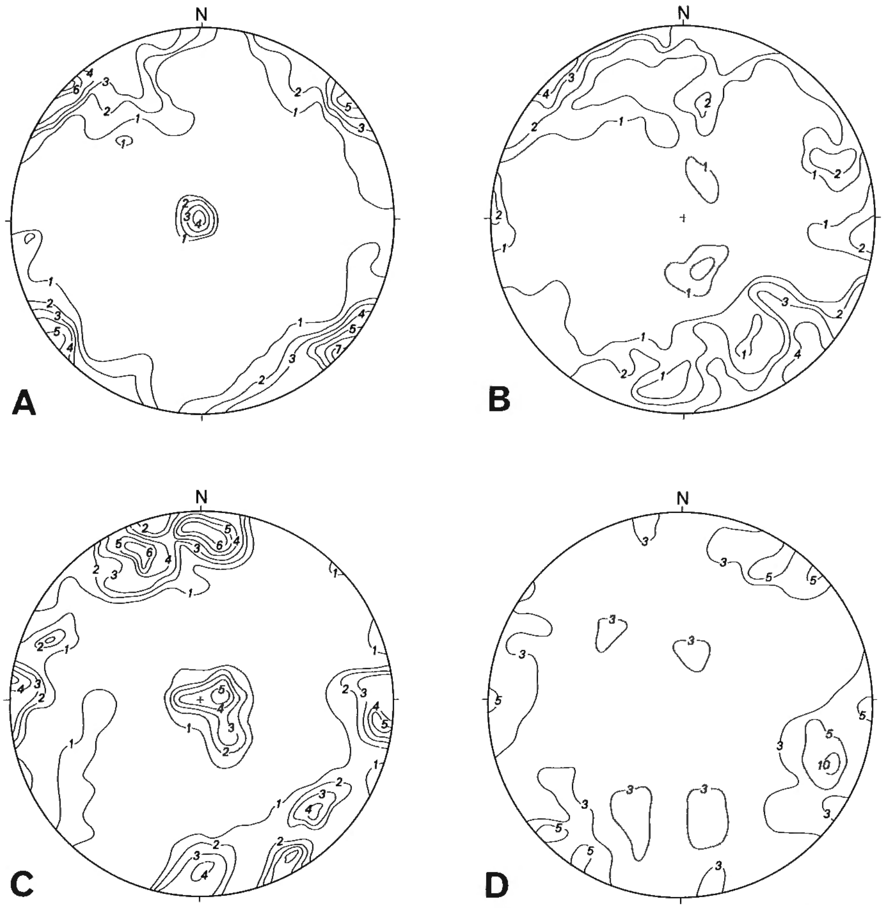


Figure 54. Joints of the northern Long Range, Labrador and Quebec, and Belle Isle.
A) Joints from the central Long Range (220 measurements).
B) Joints from the eastern of the northern Long Range (108 measurements).
C) Joints from Labrador and Quebec (63 measurements).
D) Joints from Belle Isle (33 measurements).

Early, northeast-trending, northwest-overturned, gently plunging, tight folds accompanied by strong lineation are present in the Henley Harbour region; similar folds have produced the regional lineation pattern in the Long Range (best developed in the upper Cloud River and Northwest Brook regions). The Henley Harbour folds predate more steeply plunging smaller scale folds in the Red Bay region that characterize the gneiss in and around the granulite terrane. It is possible however that both these large scale folds in the Long Range and Henley Harbour regions, which presumably formed during the metamorphic maximum, and the folds in the Red Bay region, which may have formed during emplacement of the large hornblende granite plutons, are related to different episodes of the same orogenic phase, that of emplacement of the anorthosite suite of intrusions. This is consistent with evidence that the metamorphic maximum of the granulite terrane is of about the same age as the mangerite plutons and preceded emplacement of the hornblende granite.

A second period of more local northeast-trending, northwest-overturned, gently plunging, large scale folds, which developed late in the Grenville orogeny, is evident in the Precambrian rocks on Belle Isle, in the southeast part of the Hooping Harbour and east coast regions, and about Pikes Feeder antiform. This folding is thought to have overturned the southern half of the Hooping Harbour pluton so that the inverted roof of the pluton is now exposed to the west of its root. Folding was probably accompanied by greenschist facies metamorphism and occurred during or soon after emplacement of the megacrystic plutons.

Later, northeast-trending, northwest-overturned folds are evident in the eastern part of the map area where they involve supracrustal rocks of both Belle Isle and the Long Range. On Belle Isle it is possible that this folding occurred in two distinct episodes, the earlier being of Hadrynian age and involving the Bateau Formation, and the later being of Paleozoic age and involving the Lighthouse Cove and younger formations.

Northwesterly trending folds with differing styles in differing areas are more restricted than the northeasterly trending folds, and their age relations are in large part unknown. Early northwest-trending, isoclinal folds of unknown vergence are evident on southwest Belle Isle where they represent the earliest deformation recognized on the island. Early folds with a similar trend are suggested in an isolated area in the southeast corner of the upper Cloud River region. In the Red Bay region northwest-trending foliation is evident locally within the hornblende granite plutons and surrounding gneiss. Upright to southwest-overturned northwest-trending folds are present in the Northwest Brook region where they postdate the regional northeast-trending lineation and are thought to have formed in the gneiss during diapiric emplacement of the surrounding biotite granite plutons. Late, west-northwest- to northwest-trending open warps are suggested by plunge reversals in northeast-trending lineations along the eastern seaboard.

Folding in the northern Long Range

Upper Cloud River region

The upper Cloud River region (Fig. 55) is bordered on the north by the Leg Pond pluton and the hybrid rocks to the west of it; on the east by the Cloud River pluton and an area of little outcrop (near the ice divide) that extends southward from that pluton; on the west by the contact with supracrustal rocks; and on the south by Torrent River near the limit of present mapping. The upper Cloud River region is characterized by rocks in which mineral lineation, though not everywhere, is more widely present and better developed than in any of the other regions. Three linear patterns were discerned: (1) a southwest-plunging pattern in the central and northeastern part of the region; (2) a southeast-plunging pattern isolated in the southeast corner of the region; and (3) a south-southwest-plunging pattern along the western margin of the region. Patterns 1 and 3 converge within an elongate area of hybrid rocks where lineations are scarce. Pattern 2 is isolated in the core of a dome-like structure surrounded on the north and west by pattern 1.

Near upper Cloud River the structure is characterized by homoclinal schistosity with local gneissosity and a nearly homogenous linear pattern of type 1. These are illustrated in stereogram Figure 57A-1. Poles to foliation define a sector of a great circle that is perpendicular to the lineation maximum, suggesting that the latter are "B" lineations parallel to the axis of folding. A single pole maximum is present suggesting that folding may be nearly isoclinal. Although a similar arrangement of lineation and poles to foliation might arise from a gently undulating but unfolded monoclinal sequence, this alternative appears unlikely when the distribution of lithlogies on the map is considered. The contact between leucocratic and mesocratic-melanocratic gneiss appears to bend sharply across the upper Cloud River valley about 8 km below the drainage divide. Foliation in this region in the valley bottom dips west near the sharpest part of the bend whereas that on either side of the valley dips south, with steepest dips on the south slope. The data therefore suggest the presence of a tight northwestward overturned synform in this region. The girdle formed by poles to foliation planes (see Fig. 57A-1) in part reflects a slight steepening of dips south of Cloud River valley and defines the fold axis, which is parallel to the maximum of mineral lineation. The ridge crest south of upper Cloud River, which is occupied chiefly by leucocratic gneiss, is flanked farther south by mesocratic gneiss and may therefore be interpreted to form the core of a parallel overturned antiform. For convenience of reference these folds are called the upper Cloud River synform and antiform respectively.

East of where upper Cloud River crosses leucocratic gneiss, outcrop is scarce and the gneiss is mesocratic. Foliation and a single fold axis do not conform to the pattern farther west and extension of the structure inferred to the west is not possible. Rather, foliations and fold axis may conform with those in the western part of the Hooping Harbour region to the southeast.

Leucocratic gneiss, which forms the lower megastratum of the upper Cloud River folds, was traced westward along the hill crest and into an area of submegacrystic hybrid granitic rocks in which leucocratic gneiss forms a substantial but variable component. These hybrid rocks form an elongate oval shaped area enclosed by mesocratic to melanocratic gneiss. Although mineral lineation of pattern 3 is present within the west margin of this body of hybrid rocks, the more southwesterly trending lineations of pattern 1 do not appear to persist within it. Most mineral lineations and minor fold axes within the surrounding mesocratic to melanocratic rocks plunge south and suggest that the hybrid rocks plunge south as well. An open south-plunging synform defined by

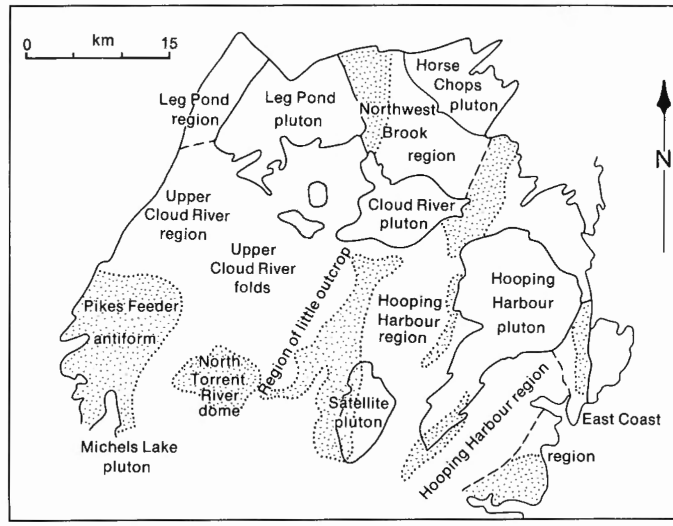


Figure 55. Structural Regions of the northern Long Range, Newfoundland. Stippled areas are characterized by foliation dipping at angles steeper than 45 degrees.

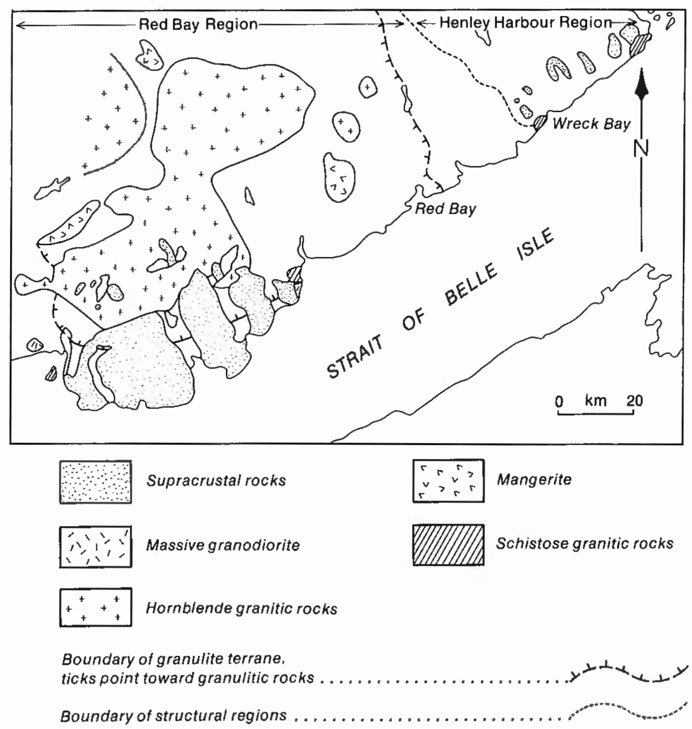
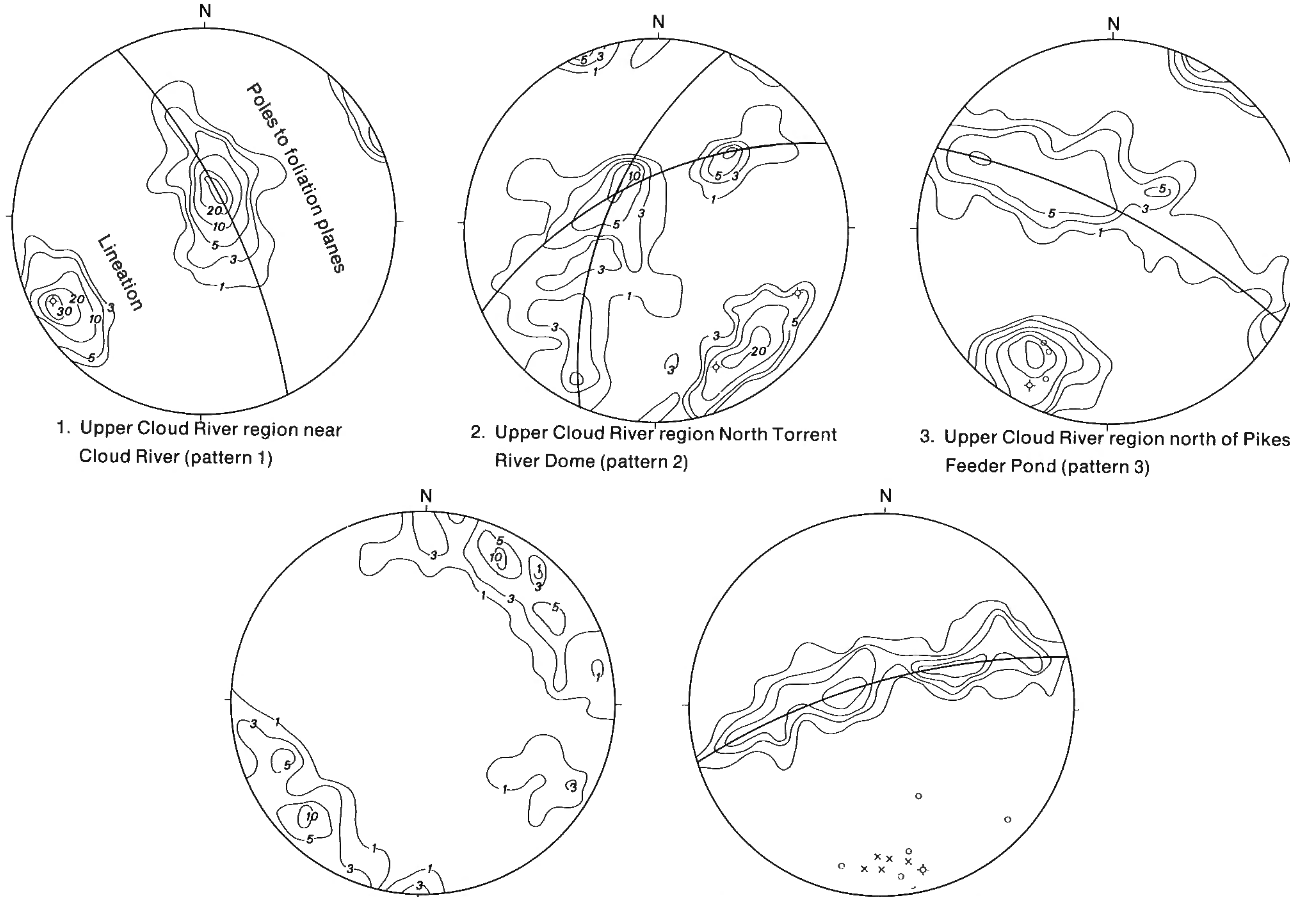


Figure 56. Metamorphic terranes and structural regions of the Labrador and Quebec area.

schistosity and gneissosity, and clearly visible on air photographs, is evident in the northern part of the hybrid rocks, but farther south, dips on the limbs of this synform steepen and the structure is pinched. Still farther south the hybrid rocks nose out belying a south-plunging synformal interpretation. Maintenance of the south plunge in the larger structure suggests that the south-plunging synform, evident in the north, reverses plunge in its southern pinched region and is eliminated before the southern nose of the hybrid rocks is reached. The structure is thus interpreted to be a compound antiform, pinched at the south and opening northward, with a doubly plunging synform at its northern crest (see structural sections, Figure 58). For convenience of reference it will be called the Pikes Feeder (compound) antiform.



1. Upper Cloud River region near Cloud River (pattern 1)

2. Upper Cloud River region North Torrent River Dome (pattern 2)

3. Upper Cloud River region north of Pikes Feeder Pond (pattern 3)

4. Northwest Brook region south part, contours for mineral lineation only

5. Northwest Brook region (upright antiform with quartz-rich gneiss at core)

Figure 57A. Stereograms showing lower hemisphere projection of linear features and contoured poles to foliation (with corresponding great circles) for structural domains in the Upper Cloud River and Northwest Brook regions. Contour intervals in per cent per one per cent area (intervals are 1, 3, 5, 10, 20, 30, 40, except in 1 where 1 per cent contour is omitted). In (3) a 7 per cent contour is shown in dotted outline. Stereogram (4) shows lineations only. For each stereogram the best fitting (visually adjusted) circle for each set of foliation poles is shown together with the pole to this circle plotted as (o). Individual mineral lineations are shown with an (x) and fold axes with an (o).

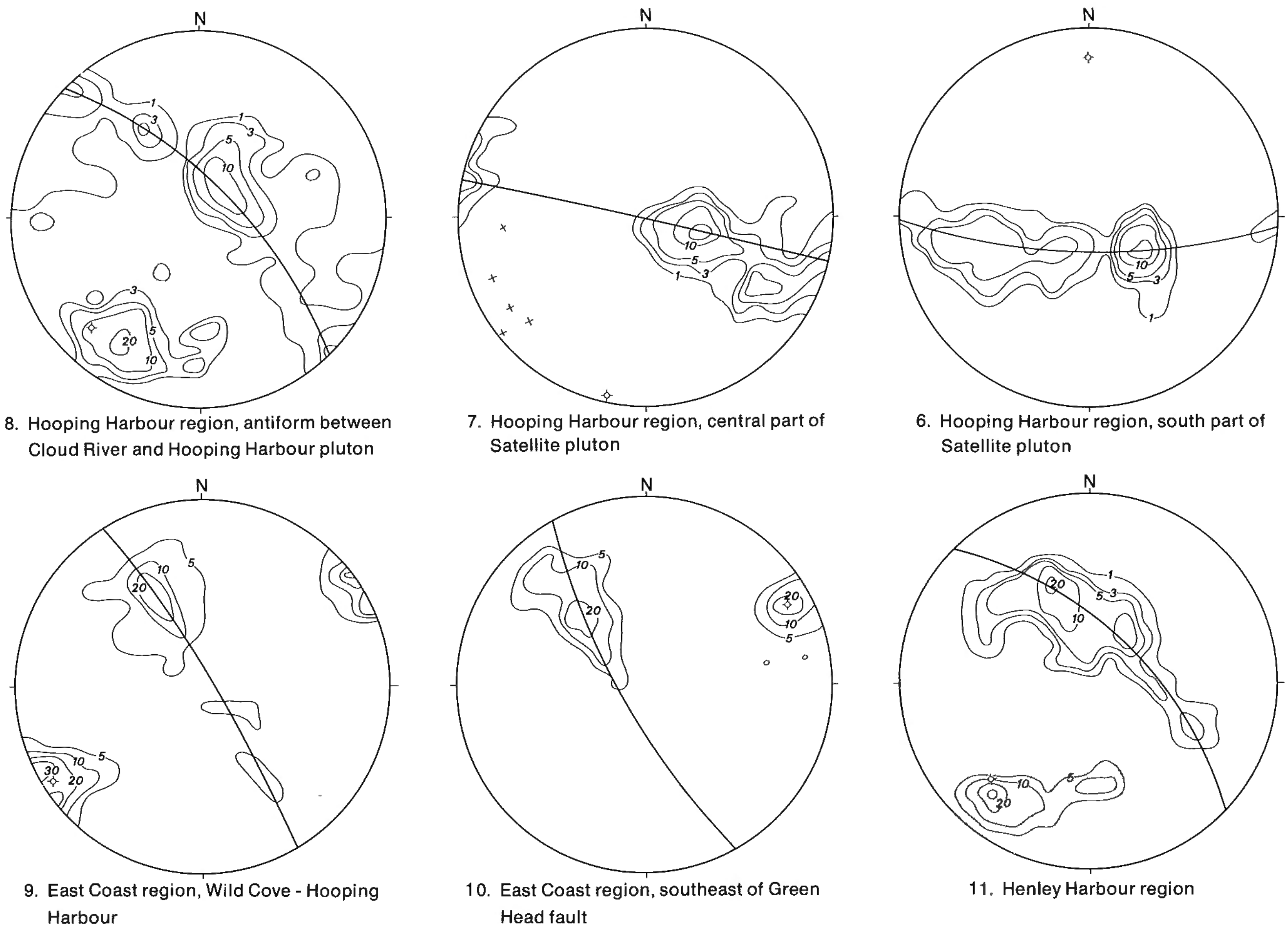
The Pikes Feeder antiform is considered younger than the upper Cloud River folds because it appears to deform these folds, and the prominent mineral lineations by which they are characterized have apparently been destroyed in its core. To the extent that the megacrystic component of the antiform is mostly weakly foliated or is unfoliated and yet conforms with the outline of the major structure, it seems likely that development of Pikes Feeder antiform is related to emplacement of the megacrystic granitic rocks.

Some 6 km east of the Pikes Feeder antiform foliation in the gneiss defines a dome-like structure that is well exposed only on its western flank. The core of this dome-like structure is characterized by pronounced southeast-plunging mineral lineation (pattern 2), very weak schistosity, and sporadic mullion-like structures. Surrounding the leucocratic gneiss core is a carapace of predominantly foliated mesocratic gneiss, although leucocratic gneiss extends well into the carapace in the northern part of the structure. Mineral lineation of pattern 2 is not present in the carapace and that of pattern 1 is present only in the outer, northern and western parts.

Foliation and mineral lineations from the core of the dome-like structure are plotted in stereogram Figure 57A-2. The figure suggests two possible girdles representing southeasterly plunging folds with corresponding axes at either end of an elongate maximum of mineral lineation and mullion-like structures.

Structure in the carapace rocks is also complex. At several localities in the northern part of the carapace small bluffs display zones of steeply dipping schistosity which truncate more gently dipping schistosity, and local alteration of shallower with steeper dips suggest that this feature may be present on a larger scale as well. Arching of the pattern 2 lineations between the northwestern and southeastern margins of the dome-like structure may reflect a slight doming and insofar as this is true, doming is later than the pattern 2 lineations. Folds and linear fabric in the core of the dome-like structure thus probably predate doming, but the southeast trend of pattern 2 lineations is unique in the map area and its relationship to structures in other parts of the map area is unknown. For convenience of reference this dome-like structure will be called the north Torrent River dome.

Pikes Feeder antiform and the north Torrent River dome are separated by a band of mesocratic gneiss in which foliation defines an open, nearly horizontal synform at the northeastern end. Near Torrent River melanocratic gneiss appears at the core of a southwest-plunging synform within



8. Hooping Harbour region, antiform between Cloud River and Hooping Harbour pluton

7. Hooping Harbour region, central part of Satellite pluton

6. Hooping Harbour region, south part of Satellite pluton

9. East Coast region, Wild Cove - Hooping Harbour

10. East Coast region, southeast of Green Head fault

11. Henley Harbour region

Figure 57B. Stereograms showing lower hemisphere projection of linear features and contoured poles to foliation (with corresponding great circles) for structural domains in the Hooping Harbour satellite pluton, and in the east coast and Henley Harbour regions. Contour intervals in per cent per one per cent area (intervals are 1, 3, 5, 10, 20, 30, except in 9 and 10 where 1 and 3 per cent intervals are omitted). In (11) these contours are omitted for lineations only. For each stereogram the best fitting (visually adjusted) circle for each set of foliation poles is shown together with the pole to this circle plotted as (o). Individual mineral lineations are shown with an (x) and fold axes with an (o).

the mesocratic gneiss, which bends westward around the southern nose of Pikes Feeder antiform. The latter synform includes remnants of pelitic gneiss on both flanks. These remnants appear to form a dismembered marker horizon somewhere near the contact zone between mesocratic and melanocratic gneiss. A prominent remnant of the pelitic gneiss also occurs within the mesocratic gneiss to the north of the synform as do patches of undivided melanocratic gneiss. Foliation of the northwest limb of the synform southeast of Pikes Feeder antiform reverses dip from steeply southeast through vertical to northwest as it is traced northward through this area, suggesting that the synform becomes tightened and overturned to the east as it is traced northward. Pelitic and melanocratic gneissic units are restricted by plunge to the southern part of the structure and are discontinuous, presumably because of disruption accompanying the northward increasing tightness of folding.

The above postulated synform, perhaps complicated by later faulting, may thus extend the full length of the mesocratic gneiss belt between the Pikes Feeder antiform and the upper Torrent River dome. This mesocratic gneiss belt however, possesses a strong linear fabric continuous with pattern 2 lineation, which apparently penetrates through both limbs of the postulated synform. It seems unlikely that this linear fabric can be entirely related to the Cloud River folds, but it is perhaps possible that both Cloud River folds and the later synform have fold axes that are roughly parallel (although overturned in the opposite direction) and that the later fold is responsible for the linear fabric in the tightly folded part of the structure.

Minor folds, mostly observed in the southern part of the area near Torrent River, are either subparallel to the mineral lineations or plunge outward from Pikes Feeder antiform. The minor folds can locally be seen to deform the linear pattern and are therefore at least partly younger.

North of upper Cloud River the linear fabric characteristic of the upper Cloud River folds continues, but at the margins of the region along the contacts with Leg Pond and Cloud River plutons, and in the pinched area between these plutons it is missing. In the central part of the region near the small megacrystic granitic plug the linear fabric persists but reverses plunge northwest of the plug, and is deformed by northwest-trending minor folds. Foliation about the plug dips gently outward unlike that about the larger megacrystic plutons. It is possible that the plug is actually a deformed sill related to those inferred along the margin of the Leg Pond pluton, but the outward dipping foliation rather suggests that it is the upper part of a small megacrystic diapir that did not rise as far up into the gneiss as did the surrounding plutons.

Near the eastern margin of the region a thin but remarkably continuous band of leucocratic gneiss separates lineated melanocratic and pelitic gneiss to the west from unlineated melanocratic and mesocratic gneiss to the east. Although the origin of this leucocratic band is not known, it appears to mark a structural discontinuity possibly limiting deformation related to emplacement of the Cloud River pluton that forms the east margin of the region. Along the south margin of the Leg Pond pluton, where a similar discontinuity is not evident, the regional linear fabric is also poorly defined.

West of Pikes Feeder antiform the gneiss are characterized by mineral lineations of pattern 3 which plunge mostly southerly (with some reversals), and are more southerly directed than the southwest-trending lineations of pattern 1. These mineral lineations are widely more prominent than foliation within the area covered by pattern 3 and are commonly the only texture evident in the rock. Although the lineation trend is fairly constant, the plunge varies (with some reversals) chiefly from 0 to 35° southward except near Torrent River, where plunge is steeper and the trend bends southwestward. Minor fold axes are few and those seen have axes subparallel to the mineral lineation. Poles to foliation within areas of southward lineation plunge (see Fig. 57A-3) suggest that westward overturned folds may be present, with axes trending approximately parallel to the lineation.

The predominant lithology within the area characterized by pattern 3 lineations is mesocratic to melanocratic gneiss. Large bodies of leucocratic gneiss, lenticular in plan, lie within darker gneiss, but data are too sparse to define their structure. In the southern part of the area apparently disconnected remnants of pelitic schist lie within the darker gneiss on either side of the large leucocratic lens northeast of Pikes Feeder Pond. West of that lens the pelitic remnants extend in decreasing abundance for at least 10 km north of Pikes Feeder Pond. The distribution of lithologies is thus highly complex and provides little basis for the evaluation of westward overturned folding suggested by the overall distribution of poles to foliation.

The distribution of rock types about the Cloud River folds and between the north Torrent River and Pikes Feeder domes suggests that a lower megastratum of leucocratic gneiss underlies a series of mesocratic and melanocratic gneiss layers within which a thin layer of pelitic gneiss is present. Application of this sequence in the area characterized by lineation pattern 3 (west of Pikes Feeder antiform) would suggest that the lens-like bodies of leucocratic gneiss are fault wedges or antiformal fold culminations in which the leucocratic gneiss has been brought up from below. North of upper Cloud River, where leucocratic gneiss lies predominantly on the flanks and northeastern extremity of the regional structure, the sequence may be the same if the overall plunge is southwestward. Local mineral lineations and minor fold axes, however, plunge gently northeast in the northeastern part and southwest in the southwestern part of this area.

The oldest structures recognized in the upper Cloud River region are those related to the regional pattern of mineral lineation, though it is clear that this pattern was superimposed on rocks that were already deformed. The upper Cloud River folds are thought to comprise major isoclinal fold structures related to this pattern. The regional pattern of lineation west of Pikes Feeder antiform appears to be slightly divergent from that to the east, and where the two come together south of the antiform the former appears to cut across the trend of the latter. This might indicate that the western lineation pattern is the younger, or it may represent a slight rotation of the two areas with respect to one another during development of Pikes Feeder antiform. Age relations between the linear fabric and structures in the north Torrent River dome and the linear fabric in other parts of the region are unknown.

Emplacement of the megacrystic plutons followed development of structures related to the regional lineation pattern because:

(1) The megacrystic granitic rocks of the region are only locally weakly foliated in contrast to the gneiss.
(2) Sillimanite in the pelitic gneiss is commonly oriented parallel to the regional lineation, and as high-grade metamorphism preceded emplacement of the megacrystic granitic rocks (and the massive granitic rocks) the lineation, pattern preceded emplacement of the plutons.
(3) In the northern part of the region the regional lineation pattern appears to be obliterated near the major pluton contacts, and is warped about the megacrystic granitic stock.

Development of Pikes Feeder antiform is thought to have accompanied emplacement of the megacrystic granitic rocks. If the granitic phase were substantially younger it might be expected to show extensive cross-cutting relations, whereas if it were older, it might be expected to show more extensive foliation formed in association with development of the antiform. Pikes Feeder antiform is younger than the regional lineation pattern because the pattern is obliterated in the gneiss phase in the core of the antiform, and is preserved only on its western margin.

Northwest Brook region

The Northwest Brook region lies between Boony Lake on the northwest and a line through Northwest Brook and the twin lakes on Cloud River on the southeast. In the northeast it is bordered by the Horse Chops pluton and in the southwest by the Leg Pond and Cloud River plutons.

In the southeast part of the region foliation defines a northwesterly elongate antiform of leucocratic gneiss with smaller, poorly defined areas of mesocratic gneiss and quartz-rich gneiss at its crest. Towards the northwest this structure appears to pinch, overturn to the southwest, and plunge out northwestward. To the northwest it is joined by a south-plunging complex antiform with quartz-rich gneiss at its core. The melanocratic gneiss on the west limb of the latter antiform is highly schistose, steeply dipping along its western margin, and locally is separated from the Leg Pond pluton by lenses of quartz-rich gneiss with interbanded pelitic rocks.

Mineral lineations in the large antiformal structure that dominates the central and southern part of the region trend northeasterly and plunge northeast and southwest away from the crest of this fold (Fig. 57A-4). At the northwest margin of the antiform these lineations are locally deformed by northwestward plunging minor folds that trend parallel to the larger structure. The northeasterly trending regional lineation pattern therefore represents an earlier deformation that has been deformed by folding along northwest- trending axes. In the northern part of region the mineral lineations trend northerly. Minor folds are apparently more numerous than in other parts of the map area, and axes roughly parallel the mineral lineation and the axis of the large south-plunging complex antiform with quartz-rich gneiss at its core. These relations are illustrated in stereogram Figure 57A-5. Because both minor folds and major south-plunging antiform are upright, this folding is thought to be later than the northeasterly trending regional lineation pattern to the south, which by analogy with the upper Cloud River region, may represent isoclinal recumbent folding. By this hypothesis it is supposed that the earlier northeast-trending mineral lineation has been destroyed with northward increasing tightness of the later structure and development of a new northerly trending mineral lineation.

The Northwest Brook region is unique within the map area in comprising a narrow belt of gneiss bordered by three major unfoliated granitic plutons. Because the plutons are unfoliated it is inferred that major deformation of the gneiss preceded and perhaps accompanied but did not follow emplacement of the plutons. The later component of major deformation in the gneiss, comprising the large southern doubly plunging antiform and the south-plunging northern antiform, is restricted at the south end of the region to the area closely bounded by the granitic plutons, and there is therefore some suggestion that this deformation is related to diapiric spreading of the granitic magmas late in their emplacement. On the other hand, local late northwesterly trending folds are also evident in other parts of the area where the granitic plutons are not so closely opposed.

The region is also unique because of the abundance of quartz-rich gneiss present. This gneiss is exposed chiefly in the core of the above described south-plunging antiform where leucocratic gneiss might be anticipated by comparison with other parts of the map area. The northern part of the region is characterized by a negative aeromagnetic anomaly that joins with an even more pronounced northeast-trending negative anomalous belt near the Paleozoic-Precambrian contact. By comparison with other parts of the map area this area of negative aeromagnetic anomalies may suggest a thickened gneiss section. It is possible that the thick quartz-rich gneiss unit is one that does not occur in the gneiss section elsewhere either as a result of nondeposition or of complex folding. The region is relatively well exposed and offers a potentially fruitful area for more detailed structural study of these problems.

Leg Pond region

The Leg Pond region lies north of the upper Cloud River region between the western scarp and the Leg Pond pluton. Of all the regions it is the least well exposed and most poorly known.

The gneiss dips for the most part moderately southeast beneath the Leg Pond pluton. Mineral lineation is only sporadically preserved and does not conform to a single pattern, suggesting that the gneiss is complexly folded. Two southeast-plunging minor folds were observed crossing the trend of nearby mineral lineation and they therefore probably represent later folding.

Hooping Harbour region

The Hooping Harbour region is extensive, lying between the east coast region on the east, and the Northwest Brook region, Cloud River pluton, and the drift-covered ice divide on the west. Its western boundary follows a zone of steeply dipping foliation that may represent a structural discontinuity (Fig. 55). It includes the Hooping Harbour pluton and its southeastern satellite, which are in part foliated and are thought to have been intimately involved in deformation within the region.

The regional northeast-trending pattern of mineral lineation, prominent in other parts of the map area, is locally absent or only sporadically preserved in the Hooping Harbour region, and is complicated by local development of subparallel to slightly more northerly or easterly trending mineral lineations and fold axes that may be younger, but for which relative age criteria could not be ascertained. Early folding by analogy with other regions of the map area is probably isoclinal recumbent, and took place before emplacement of the Hooping Harbour megacrystic pluton.

The Hooping Harbour pluton is uniquely shaped among megacrystic plutons of the northern Long Range in having a large laterally expanded 'head' with a narrow elongate 'tail' to the southwest. Although the combination of Leg Pond pluton and hybrid rocks to the southwest provides a somewhat similar pattern, the resemblance is less close than appearance would suggest, because the proportion of the granitic phase in the hybrid rocks is much lower than in the 'tail' of the Hooping Harbour pluton, and the structure expressed by foliation in the gneiss cuts across the trend outlined by distribution of the hybrid rocks. As with the other megacrystic plutons the gneiss bordering the Hooping Harbour pluton dips mostly toward the pluton, commonly at rather shallow angles (mostly 45 per cent or less). The 'tail' and satellite of the pluton, unlike the other major plutons, are foliated and have a locally prominent gently plunging mineral lineation that trends mostly parallel to their elongation. This lineation probably developed during emplacement of the pluton rather than at a distinctly later date, because parallel-trending lineation, and pegmatite boudins in the Torrent Cove assemblage that developed during emplacement of the pluton, suggest that the direction of minimum stress was northeasterly at this time, whereas later folding developed along the northwest-trending axes.

The configuration of the pluton has some resemblance to that of the steep-walled (tadpole) plutons described by Hutchison (1969) in the Central Coast Mountains of British Columbia. These plutons are believed to have been diapirically emplaced and are now seen in oblique section with prominent massive 'heads' and heterogeneous foliated 'tails'. Contacts are sharp around the heads where foliation within the pluton is steeply dipping. They are surrounded by rocks characterized by Barovian facies series (deep seated) metamorphism. On the other hand, although the Hooping Harbour pluton was probably also diapirically emplaced, there appears to be important differences between it and the western tadpole plutons:

(1) The metamorphism of the gneiss surrounding the megacrystic plutons that can be attributed to their emplacement has produced an andalusite-bearing mineral assemblage. Earlier metamorphism in the surrounding gneiss is of the lower pressure type, cordierite being locally present and kyanite absent in the pelitic rocks. Thus the depth of emplacement of the Hooping Harbour pluton is likely shallower than that of the tadpole plutons of the west coast.

(2) Structural dips in the gneiss north of the 'head' of the Hooping Harbour pluton are shallow, whereas foliation within the heads of the tadpole plutons is steep. This contrast is consistent with the hypothesis that the tadpole plutons are exposed in oblique section and were moving headwards just before solidification; and it supports the view that the Hooping Harbour pluton may be exposed in a section approximately perpendicular to the direction from which it rose and has a 'head' that was expanding laterally at the time of solidification.

(3) Although less distinctly defined and bearing more inclusions than the 'head' of the pluton, the 'tail' of the Hooping Harbour pluton is more homogenous than those described for the tadpole plutons and it has a locally prominent gently plunging linear fabric parallel to its elongation. Although this contrast might result from development of 'flow' lineation combined with different depths of pluton emplacement, it might also be due to pinching of the southern half of a laterally expanding pluton, both parts of which had reached approximately the same level.

(4) The Hooping Harbour pluton is unique in shape among megacrystic plutons in the Long Range, whereas the tadpole plutons belong to a group all of which have their head to the northwest. This suggests that the

terrane into which the tadpole plutons were emplaced has been regionally tilted, whereas tilting of the Hooping Harbour pluton, if it exists, must be local. Local tilting extensive enough to involve the whole of the pluton seems unlikely because such tilting would produce a northward regional plunge in the surrounding gneiss. If any regional plunge is evident however, it is gently to the south, as suggested by the southward decrease in the area of leucocratic gneiss exposed at the crest of the north-trending antiforms east and west of the pluton.

This last contrast particularly suggests that the shape of the Hooping Harbour pluton, which is the main point of similarity between it and the tadpole plutons of the Central Coast Mountains of British Columbia, is related to the particular environment of this pluton and is different in origin from the shape of the tadpole plutons described by Hutchison (1969).

Southwest of the 'tail' of the Hooping Harbour pluton and separated from it by about 3 km, is a small unusually heterogeneous, megacrystic pluton, the (Hooping Harbour) satellite pluton. Foliation in this pluton defines a doubly plunging synformal structure concordant with foliation in the surrounding gneiss. The structure in the southern and central portions of this synform is summarized in stereogram Figures 57B-6 and 57B-7, but in the north the structure has been bent and pinched to the point that it cannot be readily visualized on a stereogram consistent with the present scale of mapping. In the south section (Fig. 57B-6), the west limb of the structure steepens northward until eventually it is overturned, and is represented by an elongate pole maximum on the west side of the diagram. The east limb is nearly monoclinal and dips gently west, therefore producing a less dispersed pole maximum. The two limbs together define a nearly complete girdle with an axis very roughly parallel to one measured lineation in schistose megacrystic granodiorite. Farther north (Fig. 57B-7) the central part of the structure is overturned to the east with a southward plunge. Mineral lineations are dispersed, largely to the southwest, about the axis to the partial girdle formed by the limbs of the overturned structure. Three modes of emplacement of the satellite body may be considered: (i) it was emplaced as an independent stock; (ii) it was emplaced as a west-dipping sill, derived from the Hooping Harbour pluton, and was later folded with the enclosing rocks; or (iii) it represents the upper part of the 'tail' of the Hooping Harbour pluton which has been overturned and inverted by folding.

The first alternative appears improbable because, although pinched at the north end, foliation within the satellite pluton otherwise conforms remarkably well to a synformal fold pattern that exends into the gneiss well beyond its margins. The second alternative resembles the third in providing a ready explanation for the fold structure that is apparent in the rocks. Both suggest that whereas the northern head of the Hooping Harbour pluton spread laterally, the southern part of the pluton was squeezed and flowed westward into the satellite body. The second hypothesis suggests that flowage occurred parallel to the foliation in the surrounding gneiss, separated the roof from the floor, and resulted in a normal sequence of gneiss containing a layer of plutonic rock. On the other hand, the third hypothesis suggests that the magma chamber itself was folded and overturned to the west, and that both the satellite pluton and a part of the surrounding gneiss sequence are inverted. In the former case the gneiss may have been deformed in place producing an eastward overturned synform that bears no necessary relation to the Hooping Harbour pluton. In the latter, the gneiss around the satellite pluton represents the roof of the Hooping Harbour pluton, and formation of a continuous fold structure (now deeply dissected by erosion) between the 'tail' of the Hooping Harbour pluton and the

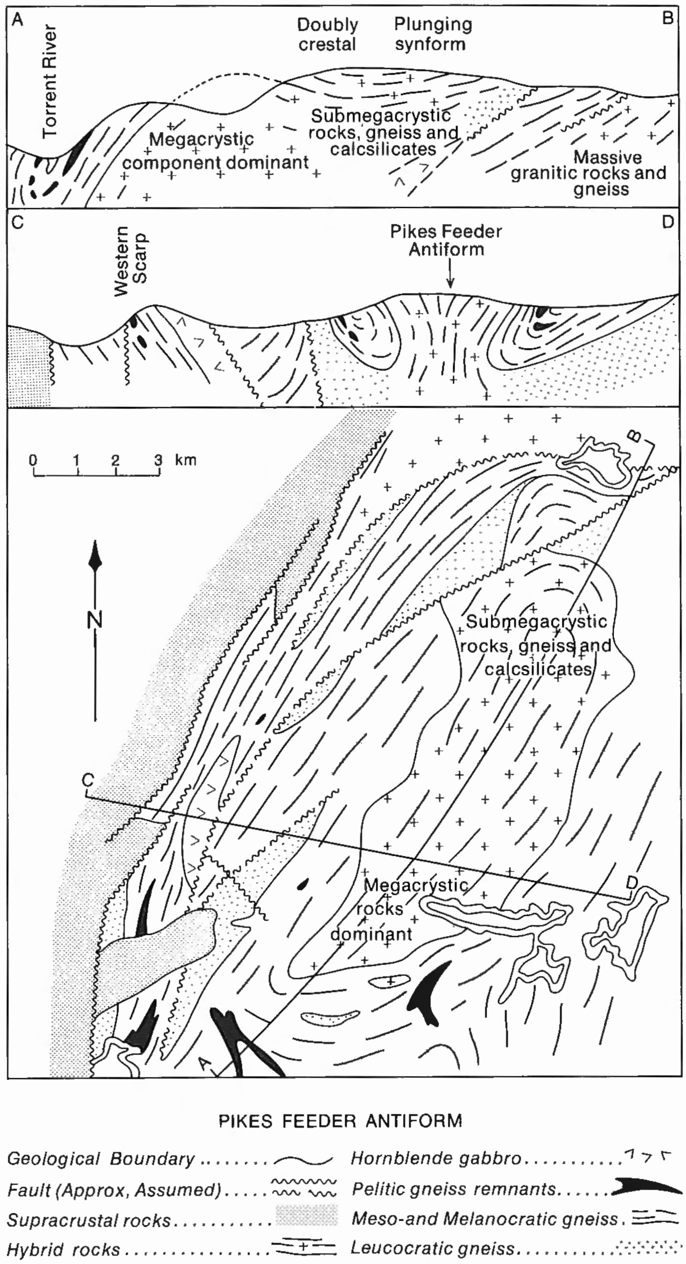


Figure 58. Interpretive structural sections through Pikes Feeder antiform.

satellite body is implied. Because the gneiss surrounding the satellite body is commonly schistose and difficult to differentiate, the sequence as presently known cannot be compared with that in the gneiss elsewhere. Furthermore top determinations are not possible. The relative merits of the second and third hypotheses therefore cannot be judged directly. The third hypothesis is preferred (Fig. 59) because it provides in simple outline a possible explanation for the repetition of leucocratic gneiss bands that occurs north of the satellite pluton. Detailed comparison between model and observations cannot be made however, because the structure is poorly exposed along its western and northern extremities and is complicated by a complex series of faults.

In the region north of the satellite pluton leucocratic gneiss occurs in two small dome-like structures at the crest of a north-trending open antiform and extra bands of leucocratic gneiss occur on the flanks. The dome-like occurrences are considered to represent the lower leucocratic megastratum in normal position whereas the extra bands may be interpreted as representing the lower limb and nose of the proposed overturned fold (Fig. 59, Section A-B). The abrupt disappearance of the western complex leucocratic gneiss band as it is traced northward may reflect decreased amplitude of the westward overturned recumbent fold as it draws abreast of the 'head' of the Hooping Harbour pluton. The elongate zone of steep to vertical dips, which lies west of the satellite pluton, and also terminates northward, may reflect steepening of foliation near the nose of this recumbent fold.

In view of the above discussion it is suggested that the Hooping Harbour pluton was emplaced into the surrounding gneiss from below (as opposed to emplacement from the south as would be the case if it were a true 'tadpole pluton'), and that whereas the north part of the pluton was able to spread laterally, the southern part was pinched along an axis nearly parallel to northeast-trending older folding (suggested by the regional lineation) in other parts of the northern Long Range. This pinching caused the main body of the magma in the southern part of the pluton to flow into the core of a large antiform that developed in the roof of the magma chamber and was overturned to the west. Megacrystic magma was carried along in the core of this fold and penetrated to the crest in the vicinity of the satellite pluton. The early greenschist metamorphic overprint (pre-diabase), which is not found about the other northern plutons, may result from enhanced dissemination of heat brought on by folding and attenuation of the southern part of the Hooping Harbour magma chamber.

The proposed deformation of the Hooping Harbour and satellite plutons implies the development of a large open antiform between the 'tail' and satellite pluton. This antiform, apparently disrupted by faulting, is evident through foliation in the gneiss and continues north to the region between the 'head' and the Cloud River pluton. Between the plutons this antiform tightens and has a locally developed northeast-trending mineral lineation parallel to its axis (Fig.57B-8). The structure is complicated by a second set of more restricted northerly trending lineations and minor fold axes for which relative age relations are not known. A second but more restricted north-trending antiform is evident in the gneiss east of the tail of the Hooping Harbour pluton and south of Hooping Harbour.

Late, large scale, open, northwest- to west-trending folds, indicated by plunge reversals in the northeast-trending lineations, chiefly east of the Hooping Harbour pluton, intersect the northerly trending folds at a high angle. Near the axis of one of these folds, on the north shore of Hooping Harbour, a northwest-trending minor fold with local parallel mineral lineation warps the older northeast-trending lineation. These late northwest-trending folds appear to project across the Hooping Harbour pluton but cannot be followed with any confidence far into the gneiss to the west. They thus postdate emplacement of the Hooping Harbour pluton. They are, furthermore, younger than the upright folding inferred to exist in the East Coast region and thought to have postdated emplacement of the Hooping Harbour pluton. If the northwesterly trending folds of the Northwest Brook region are associated with emplacement of the surrounding plutons as seems likely, then the northwesterly trending folds in the Hooping Harbour and East Coast regions are probably of distinctly later development.

Where antiforms of the northeasterly and northwesterly trending folds intersect along the northern and eastern margins of the Hooping Harbour pluton, extensive areas of leucocratic gneiss are exposed, but on the west side of the pluton the disposition of leucocratic gneiss bodies is more complex. There, two small bodies of leucocratic gneiss are evident along the crest of the northtrending antiform, and prominent leucocratic bands lie on both limbs. To the extent that the western complexity may be due to repetition by

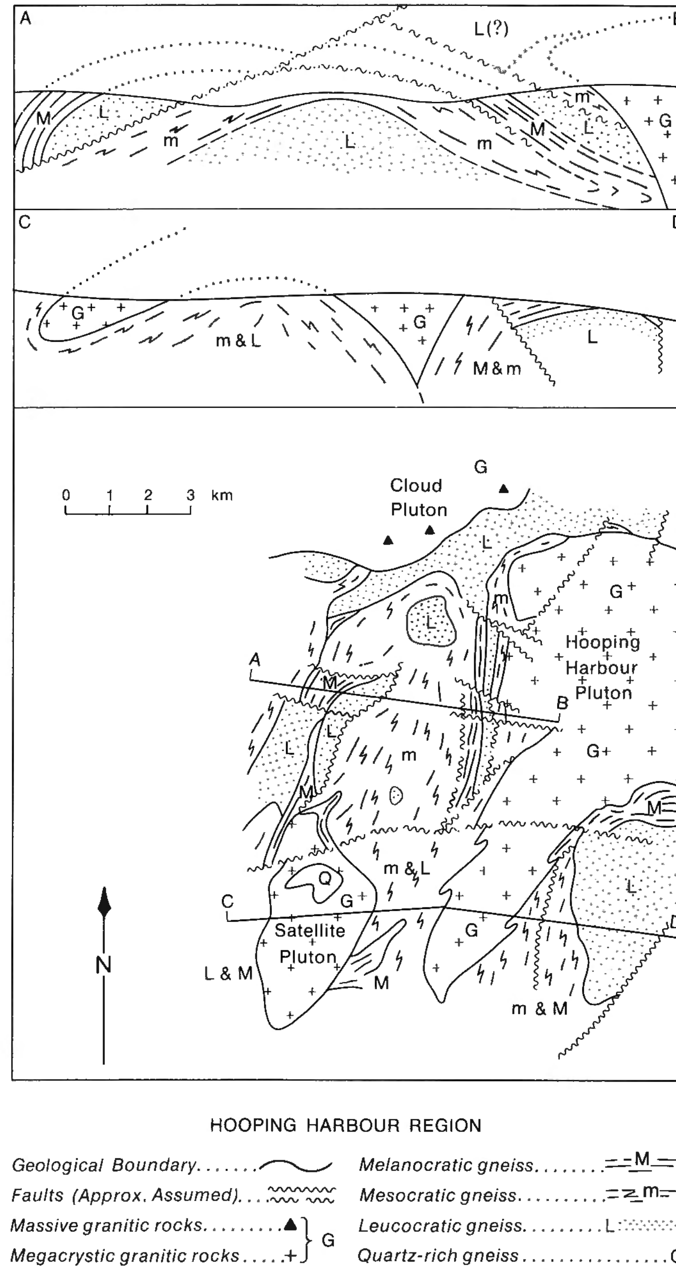


Figure 59. ***Structural sections showing the interpreted relationship between the Hooping Harbour megacrystic granitic pluton and its satellite pluton, and a speculative interpretation of associated folding in the gneiss farther north. The northern section (A-B) provides a possible explanation for the apparent doubling of major leucocratic gneiss bands unique to this region.***

folding discussed in connection with deformation of the 'tail' of the Hooping Harbour pluton, the remaining occurrences support the interpretation that the darker gneiss are lying on top of a megastratum of leucocratic gneiss.

East Coast region

The east coast region extends southward from a fault about 1.6 km north of Wild Cove to the south limit of mapping. On the east it is bounded by the sea and by the Wild Cove fault (Betz, 1939) extending from Wild Cove to southwest of

Sugarloaf hill. On the west it is bordered in succession southward by the Hooping Harbour pluton, the river flowing into the northeast arm of Hooping Harbour, and by a fault extending from the south shore of Hooping Harbour to the southern limit of mapping west of the head of Fourché Harbour.

Between Wild Cove and Hooping Harbour, but particularly near Wild Cove, mineral lineation in the gneiss trends more easterly than the common regional lineation in the Hooping Harbour region. Although the data are limited, it is possible that several nearly upright, gently south-plunging folds are present in this region (Fig. 57B-9). Foliation parallel to this trend is present for some distance into the 'head' of the Hooping Harbour pluton, and both pluton and marginal gneiss show intense greenschist facies alteration. These folds thus may postdate emplacement of the pluton, but they apparently precede intrusion of the diabase dykes. They may therefore be related to deformation presumed to have pinched the 'tail' of the Hooping Harbour pluton. Farther south toward East Arm of Hooping Harbour, foliation in the melanocratic and mesocratic gneiss becomes homoclinal, dipping 30 to 65° southeast. Near Sugarloaf hill leucocratic gneiss is brought to the surface. To the extent that the leucocratic gneiss may be expected to lie beneath the darker gneiss, this suggests either that a northeast-trending, gently plunging, westward overturned synform with an axis west of East Arm is present, or that the leucocratic rocks along the coast have been thrust up against the darker gneiss to the northwest.

South of Hooping Harbour the east coast region is broken into blocks, principally by two steeply dipping faults that are visible as gullies or clefts at the coast, and locally, in the southern part of the region, mylonite has been found along their lineaments. The more westerly of these faults extends from the south shore of Hooping Harbour about 1 km east of the town, southwest to the limit of mapping west of Fourché Harbour. The eastern fault extends from near Green Head at the entrance to Hooping Harbour, southwestward to a confluence with the former fault in the valley about 1.5 km northwest of Willamsport. These two faults are called the west and east branches of the Green Head fault respectively, as a single prominent lineament continues for several kilometres southwest of their confluence and beyond the map area. A third mylonite-bearing fault extends northwest from the first cove east of Williamsport, and a series of steeply dipping but more easterly striking faults, along which no mylonite was seen, probably intersects the Green Head fault from the east.

Foliation in the block between branches of the Green Head fault is commonly gently to moderately dipping and not markedly discordant with that in the gneiss west of the west branch of the fault. Because the block occupies a belt of low relief and is largely tree-covered, the structure in the gneiss is not as well known as that in the rocks on either side. Nevertheless, the rough correspondence of lithologies within the block and those farther west suggests that movement on the west branch is not great. Foliation in the gneiss southeast of the Green Head fault and northeast of the mylonite-bearing fault east of Williamsport, strikes northeast with a moderate southeast dip. Dips are shallower in the vicinity of Green Head and steeper near Fourché Harbour. Crenulations and minor fold axes trend and plunge northeast. Several mesoscopic isoclinal folds with axial planes dipping gently southeast are visible in the cliff face on the north shore of Silver Cove (Fig. 60). One of these is truncated by a southeast-dipping fault along which pegmatite appears to have been intruded. The axis of pegmatite boudins within the gneiss trends roughly 160°, parallel to boudin axes within the Torrent Cove assemblage. Furthermore, the style of these folds is in contrast to that of minor folds associated with the regional mineral lineation where it (the regional mineral lineation) is well developed in other parts of the map area. The data suggest that these folds are younger than the regional mineral lineations in the western part of the map area, and are older than the open northwest-trending folds defined by lineation plunge reversal. They probably developed during emplacement of the Hooping Harbour pluton (at the same time as Torrent Cove assemblage). Poles to foliation and fold axes (including crenulations) from the coastal exposures are shown in stereogram Figure 57B-10. Foliation poles form a partial girdle that is roughly perpendicular to the axis of minor folds, and this together with the folds at Silver Cove, suggests that the gneiss is tightly folded with southeast-dipping axial planes.

To the extent that the major leucocratic gneiss bodies appear to lie beneath darker gneiss in other parts of the map area, the appearance of an extensive belt of leucocratic gneiss northeast of Fourché Harbour against a southwest-plunging structure west of the Green Head fault, suggests that the rocks southeast of the fault have been uplifted. Movement along the Green Head fault thus appears to have had a vertical component with the southeast side moving up.

In the cove immediately east of Williamsport and on the hillside to the northwest, mylonite is exposed with fine lenticular laminae dipping north to northeast. Farther west near Williamsport and on the south shore of Fourché Harbour

Figure 60. Isoclinal folds at Silver Cove south of Hooping Harbour. (GSC 160135).

leucocratic to mesocratic gneiss contains gently plunging, north-trending crenulations and fold axes. Foliation strikes north roughly parallel to linear features and is shallow dipping near Williamsport but steepens to vertical near Squally Point. Structural trends thus appear discontinuous with those farther north and it seems likely that a distinct structural domain exists south of the Green Head fault and southwest of the fault east of Williamsport.

Folding in Labrador and Quebec

Henley Harbour region

The Henley Harbour region extends from the northeast corner of the map area southwest along the coast to a small granitic pluton immediately west of Wreck Cove (Fig. 56). Inland the boundaries are imperfectly known but they may extend northwesterly from this pluton to the north border of the map area.

The Henley Harbour region is characterized by northeasterly trending, gently plunging mineral lineations and minor fold axes, and by gently dipping foliation (35° and less), except in the area about Chateau Bay where dips up to 60° were observed. Poles to foliation and linear features from the gneiss near Henley Harbour are shown in stereogram Figure 57B-11. Poles to foliation tend to lie on a great circle perpendicular to the maximum of linear features. A prominent pole maximum corresponding to southeast dips of about 45° is evident. This suggests that the gneiss is tightly folded and overturned to the northwest about fold axes plunging about 20° southwest. A schistose lineated granitic body extending southwest from York Point appears to be a lenticular sheet-like mass that has been involved in this folding, and may occupy the core of one of these folds.

Farther southwest along the coast the southwestward plunge of lineations decreases and near Woody Cove the lineation plunges northeast. A similar plunge reversal occurs in the gneiss inland to the northwest suggesting the presence of a late open northwest-trending synform. On the west coast of Wreck Bay isoclinal recumbent minor folds with axes plunging northeasterly at 10° are present against a small granodiorite body in which inclusions of gneiss are present. At the western margin of the granodiorite the plutonic rock is intensely lineated parallel to minor fold axes in the adjacent gneiss which plunges 10 to 20° southeastward. A large inclusion of gneiss at the coast in the central part of the granodiorite also shows this latter lineation, although the surrounding granitic rock appears nearly massive. These observations suggest that the granodiorite body intruded into the folded eastern gneiss sequence and that folding west of the granodiorite postdates its emplacement. By this interpretation isoclinal northwestward overturned folding of the gneiss of the Henley Harbour region predates the folding farther west.

Red Bay region

The Red Bay region encompasses all of the Labrador and Quebec area west of the Henley Harbour region. It is characterized by moderately and steeply dipping gneiss (dips mostly greater than 35°) in which minor fold axes and mineral lineations are also commonly steeply plunging. Structural domains apparently exist on a smaller scale than is evident in the Henley Harbour region, as is suggested by the frequent change in trend and plunge of linear features. The central and western parts of the area are dominated by hornblende granite, and mangerite intrusions, which extend beyond the western boundary of the map area. Gneiss near the hornblende granite and mangerite plutons is steeply dipping and though concordant with pluton contacts over much of their extent is in some areas extensively discordant. The eastern limit of extensive steeply dipping foliation in the gneiss lies some 13 to 22 km east of the plutons and some 3 to 13 km east of the eastern limit of granulite facies metamorphism suggested by the appearance of sub-mesoperthitic alkali feldspar (Fig. 56). There is some indication therefore that an aureole of structural complexity characterized by steep dips and plunges exists about the complex of hornblende granite and mangerite plutons.

Argument has been made that the Bradore Bay granodiorite is younger than the other plutons mapped north of the Strait of Belle Isle, and to the extent that this is true, doming apparent in the gneiss about this pluton may be younger than structures related to emplacement of the other plutons. The possibility must therefore be considered that a domain of younger metamorphism and deformation lying to the south of the Red Bay region has barely reached the north coast of the Strait of Belle Isle.

Folding on Belle Isle

The oldest structures observed in the gneiss on Belle Isle comprise northwest-trending isoclinal folds suggested by minor fold noses in which the limbs have been stretched out parallel to foliation. These minor folds appear roughly parallel to mesoscopic folds evident locally in the 150-m bluffs that surround the island. They are also approximately parallel to a sporadically preserved hornblende lineation in the gneiss. In the southwest part of the island the predominant strike of foliation in the gneiss, although variable, is also northwesterly roughly parallel to the northwesterly trending fold noses. In the eastern and northern parts of the island foliation strikes northeast and dips southeast. Somewhat more open, northeast-trending, gently plunging minor folds with southeast-dipping axial planes have locally deformed the earlier, sporadically preserved, hornblende lineation. Furthermore, because the mineral lineation in this part of the island appears to have been steepened and rotated southward, plunges vary from 5° northwest to 20° southeast in the southwest and from 20° to 35° south-southeast in the northern and eastern parts of the island. The data suggest that complex early folding, perhaps dominated by northwest-trending isoclinal folds, was followed in the northern and eastern parts of the island by northeast-trending folding that produced less tight, northwest-overturned, minor folds, and was responsible for the homoclinal southeast dip of foliation in these regions. This later period of deformation also appears to be complex, because although the superimposed deformation in northeast Belle Isle was clearly not as severe as the earlier deformation that is most evident in the southwest, it nevertheless has almost completely reoriented foliation in the gneiss to southeastward dips, and has produced northeast-trending minor folds. In contrast, the dykes of the Long Range swarm on Belle Isle, although they locally have a strong southeast-dipping cleavage, for the most part, maintain their northwest dips and have not been rotated into parallelism with foliation in the enclosing gneiss. It appears therefore that most of the northeasterly trending deformation of the gneiss took place before effusion of the Lighthouse Cove basalt and emplacement of their related dykes, and hence is of Precambrian age. It may be that the main period of northeasterly trending deformation of the gneiss was late Grenvillian, analagous to deformation in the Long Range which affected only part of the eastern megacrystic (Hooping Harbour) pluton.

Gneiss along the northeast coast of the island is overlain by the Bateau Formation with a structurally conformable, steep, southeast-dipping contact exposed at the southern limit of exposure of the Bateau. The northern exposures have been broken into at least three northeasterly trending wedges that appear to have been rotated with respect to one another about northeast-trending axes. Beds in the southern, predominantely conglomeratic, wedge dip 78° southeast; beds in the middle quartzite wedge dip 5° to 15° north; and beds in the northern quartzite wedge dip 35° southeast near the coast but are apparently roughly vertical near the contact. The gneiss below the Bateau Formation is finely fractured and altered over a broad zone north of Bateau Cove but this zone appears to narrow and die out southward. The data indicate that the western contact of the Bateau Formation may be faulted over most of its extent, with the amount of movement perhaps increasing northward in porportion to the extent of alteration in the basement gneiss. By this interpretation, the northern Bateau fault wedges of relatively clean quartzite are thought to have been thrust over and against the gneiss to the west.

The Bateau Formation is intruded by diabase dykes related to the Lighthouse Cove basalt, and these dykes have a variable and locally strongly developed southeast-dipping cleavage. The foliation in the dykes may be younger than the faulting of the Bateau Formation, but a more detailed examination of the dykes in the Bateau fault wedges is necessary to establish this. Northeast-trending, steeply dipping zones of schistosity are also present in the Lighthouse Cove Formation at Barbers Cove, and an arkosic lens at this locality has been folded between flows, forming a northeast-trending minor fold that plunges 5° southwest. This latter deformation apparently preceded the biotite zone metamorphic overprint on Belle Isle as biotite of this overprint is randomly oriented. Still, the two are probable post-Precambrian and closely related.

Gneiss along the southern and western coasts of the island is overlain by supracrustal rocks comprising the Lighthouse Cove, Bradore, Forteau, and White Point formations. These formations, for the most part, occupy two northerly trending grabens separated by horsts of Precambrian gneiss-dyke rock upon which remnants of the lowermost formation, the Lighthouse Cove basalt, are preserved. The basalt on the horsts dips southeast whereas the sediments in the grabens dip predominantly south to southwest. The west graben is truncated by a zone of reverse faulting that follows the western shore of the island as far north as Lark Harbour, and may continue en échelon northeast of Wreck Cove. Normal faulting probably entirely preceded reverse faulting, but it could have occurred during intervals of dilation between pulses of reverse faulting.

The Lighthouse Cove and Bradore formations of southwest Belle Isle are for the most part competent and show little penetrative deformation. Schistosity is however evident in the basalt at Lark Harbour where the underlying contact is nearly vertical. The less competent Forteau shales are little deformed between Round Head and Scotswood Cove except where they occur near the east boundary faults of the grabens. There they are contorted and show well developed superimposed cleavage. Minor folds in the contorted zone of the western graben plunge southward at a steeper angle than bedding to the west. North of Round Head the Forteau Formation is compressed into tight folds with strong eastward dipping to vertical axial plane cleavage (Fig. 39). At Lark Island these folds plunge northward, but from Lark Harbour south to Roundhead Cove they are nearly horizontal.

North- to northeast-trending, steeply dipping or near-vertical faults are clearly concentrated in and along the margins of the supracrustal rocks at the southwest end of Belle Isle (Fig. 61). North of Green Cove, the gneiss along the west coast is more fractured and oxidized than the gneiss northeast of Wreck Cove and along much of the eastern margin of the island. At Lark Harbour a near vertical fault is again evident, separating basalt and minor sediments from the underlying basement. It seems likely therefore that a steeply dipping fault zone is present along the western margin of the island as far north as Lark Harbour. Westward overturned folds in the Forteau Formation, which is most

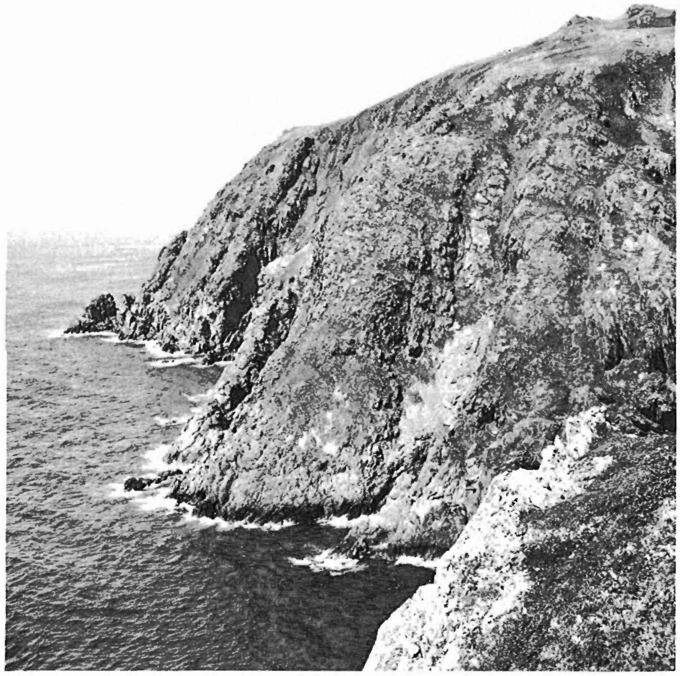


Figure 61. View looking north at flows of the Lighthouse Cove Formation north of White Point Cove, Belle Isle. The flows are down-faulted against gneiss (in the foreground) and are cut by many parallel minor faults near the west margin of the island. The flows dip about 32° southward toward the observer. (GSC 160195).

deformed along the western margin of the island, suggest that the supracrustal rocks have been compressed and overidden from the east. The northeastward structural plunge north of Lark Harbour may indicate that movement along this zone died out north of Lark Harbour. A second broad zone of fracturing and oxidation of the gneiss is evident at Wreck Cove. Along the western margin of this zone, on the west shore of the Cove, a northeast-trending fault clearly defined by carbonate lenses (Fig. 62) dips about 60° east. Traverses across the uplands south of Wreck Cove revealed that more than normal shattering and alteration of both gneiss and dykes are present locally as far southwest as the hills southeast of Green Cove, but similar brecciation and alteration were not observed east of White Point Cove. It is possible therefore that movement along the western fault south of Lark Harbour was accommodated farther north along a parallel fault zone through Wreck Cove.

On the southern coast of Belle Isle the supracrustal rocks have been down-dropped along two southward plunging grabens. The marginal faults along the east boundaries of the grabens are characterized by faulted slivers of the more competent supracrustal rocks and by contortion and superimposed cleavage in the less competent Forteau shales. In contrast, along the west marginal faults the supracrustal rocks, including the Forteau Formation, show little deformation to within a metre of more of the faults. The west marginal fault exposed at Roundhead Cove dips roughly 50° east. Beds of the Bradore Formation in the hanging wall are underformed, but a wedge of brecciated indurated gneiss extends along the footwall. Some late right-lateral movement along the fault is indicated because a sliver of arkose from the hanging wall appears to have peeled off westward against the gneiss bluff that forms the south wall of the cove (Fig. 63). Similar right-lateral movement, evident in offset of the Bradore-Lighthouse Cove contact along faults within the east graben, was reported by Williams and Stevens (1969). Both boundary faults of the western graben are truncated by the western fault zone near White

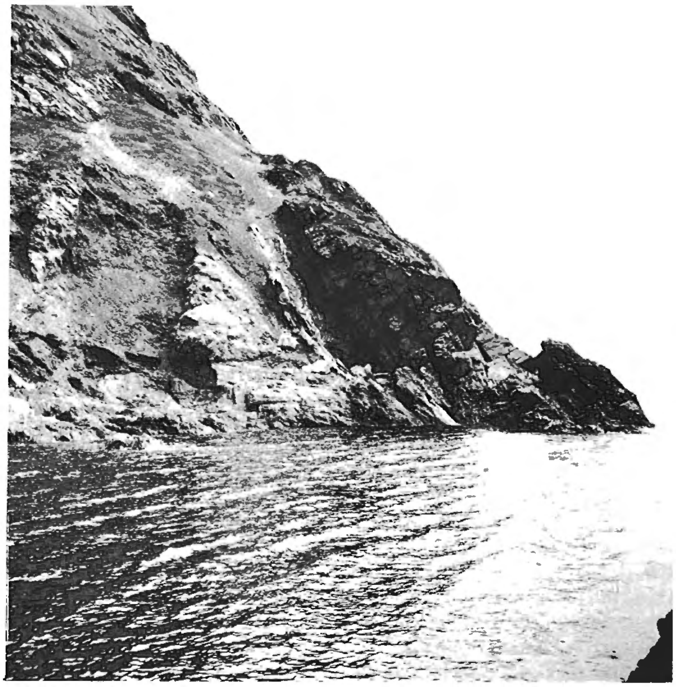


Figure 62. Carbonate lens along the fault at the west shore of Wreck Cove, Belle Isle. (GSC 160171).

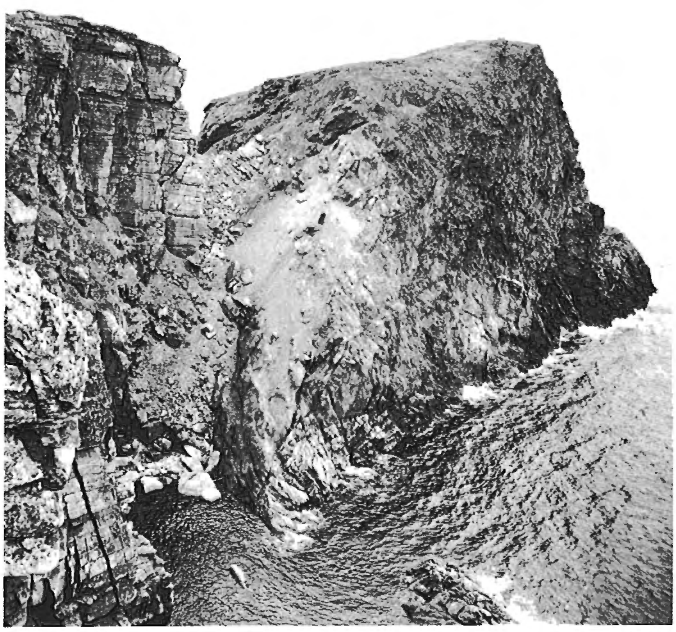


Figure 63. View of Round Head, Belle Isle looking south. A fault dipping roughly 50° east separates the Bradore Formation to the east from Grenville gneiss to the west. A zone of brecciated indurated gneiss occurs along the fault and a sliver of Bradore arkose at the water line has pealed off the hanging wall suggesting some right-lateral movement along the fault. (GSC 160187).

Point, suggesting that reverse faulting there is later than formation of the graben. Furthermore, deformation of the supracrustal rocks in the graben along the east boundary faults, but not along the west boundary faults, may indicate that the neighbouring gneiss was overthrust from the east after formation of the grabens. Thus the grabens are probably earlier than thrusting, but graben formation could have taken place during periods of dilation within a protracted period of intermittant thrusting.

Faults

Summary of faulting

Catazonal deformation associated with emplacement of the anorthosite suite of intrusions undoubtedly produced surfaces of discontinuity which have been folded and recrystallized with their enclosing gneiss so that virtually no evidence of their existence remains. Later deep-seated fault movements that produced penetrative cataclasis and recrystallization through zones up to 300 m thick are suggested by the Torrent Cove assemblage and by the Fourche Point (quartz-eye) schist. Cataclasis evident in both of these units is probably of late Grenville age; in the Torrent Cove assemblage it is thought to be related to diapiric spreading of the Hooping Harbour pluton, and deformation which accompanied or immediately followed its emplacement. Other faults of this general age probably exist and a few have been postulated west of the Hooping Harbour pluton, but they have been folded and direct evidence of their existence has not been found.

Fault movements in the basement complex are suggested by northeast-trending felsitic cataclastic rocks, by mylonite and protomylonite near the east coast of the Long Range, and by east-northeast- to north-trending breccia zones and faults throughout the map area. In part these are probably Hadrynian because faulting of this age likely accompanied uplift to produce the late Hadrynian unconformity developed after the Grenville orogeny and before deposition of the Hadrynian-Cambrian strata. Early faulting of the Bateau Formation, and breccia dykes associated with the Long Range dyke swarm provide further evidence of faulting of this age. Other faults of this group are Paleozoic. Some faults cut fossiliferous rocks of Cambrian to Early Ordovician age. Lower Paleozoic normal faults produced grabens that were locally truncated by reverse faulting possibly during the Ordovician orogeny. The Cabot fault, which may record major transcurrent movement (Wilson, 1962), skirts the east coast of the Great Northern Peninsula and cuts rocks of early Mississippian age.

Faults in the northern Long Range

Two prominent topographic lineaments extend from the south shore of Hooping Harbour southwest to a point northwest of Williamsport where they coalesce and continue as one lineament to the southern border of the map area. Several exposures of mylonite were found on the valley floor south of the convergence of these two lineaments, and both probably represent branches of the same fault. The fault is called the Green Head fault after Green Head, Hooping Harbour, near which its eastern branch passes. Dips measured on the fault planes where they are exposed at Hooping Harbour are 50° northwest for the western branch of the fault and 40° southeast for the eastern branch. The straightness of the eastern branch trace near Hooping Harbour, and its swing to the west north of Williamsport, however, argue against shallow eastward dip measured, and suggest that this fault is for the most part more steeply dipping. No mylonite was found associated with the faults at Hooping Harbour. Furthermore, although the east fault may continue a short distance north of Eastern Arm, Hooping Harbour, and some brecciation is evident in rocks along the east wall of the valley there, the west branch lineament stops at the north shore of Hooping Harbour. It seems likely therefore that movement along the Green Head fault decreases northward and dies out altogether near Hooping Harbour. The appearance of an extensive area of leucocratic gneiss southeast of the eastern fault may indicate that the southeast block has been thrust up against the rocks to the northwest (because leucocratic gneiss is believed to underlie the darker gneiss in this region). If so, then the movement is in the same sense as that observed by Foley (1937) on southeast dipping faults near the coast at Great Harbour Deep. Age relations between the Green Head fault and the diabase dykes are unknown and direct evidence of the age of movement along the fault is thus not available.

Rocks of cataclastic origin, the Fourché Point schist and protomylonite southwest of Sugarloaf hill, provide further direct evidence of fault movement in Grenville gneiss along the east coast. These movements may be of differing ages: that represented by the Fourché Point schist having probably preceded emplacement of the Long Range dyke swarm, and that represented by the protomylonite at Sugarloaf hill probably followed emplacement of the swarm. The Fourché Point schist contains late biotite similar to that in gneiss immediately to the west (K-Ar biotite age 434 ± 18 Ma) thus indicating that schist was uplifted and cooled below the argon blocking temperature in late Ordovician time after emplacement of the Ordovician klippe.

Betz (1939) considered the Wild Cove fault, which skirts the eastern margin of Grenville gneiss between Sugarloaf hill and Chimney Bay (lying mostly within the supracrustal rocks), to be the most important structural lineament in the Canada Bay area. He reckoned that some 1200 m of strata were missing along the fault at Wild Cove. Near Sugarloaf hill the Wild Cove fault apparently bifurcates, the west branch reaching the coast southwest of Sugarloaf hill where a metre or more of Bradore Formation, structurally overlain by protomylonite, intervenes between two blocks of Grenville gneiss, and the east branch reaching the coast south of Sugarloaf hill between the eastern block of gneiss and schist of the Ordovician klippe. On the uplands west of Sugarloaf hill and north of the point where the two branches of the Wild Cove fault are thought to coalesce, there is a thick zone of schistose rocks against the western margin of the gneiss. Within this zone a wedge of Bradore arkose was found in fault contact with, and overlying a wedge of rock similar to the protomylonite at the seashore but severely brecciated. The relations suggest that extensive fault movement has occurred north of the junction of the two branches of the fault after formation of the protomylonite. The western branch may therefore be part of an older fault system.

The Cabot fault zone (Wilson, 1962), which lies just offshore along the east coast of the Great Northern Peninsula, is presumably responsible for the remarkable straightness of this coastline. Betz (1943) considered this fault to be predominantly due to high-angle thrusting from the east. Wilson (1962) on the other hand considered the fault to be predominantly transcurrent with a large but undetermined direction of movement. The latest movement along this fault, as shown by deformation of Carboniferous strata at Cape Rouge and Conche peninsulas, was post-early Pennsylvanian (Baird, 1966).

Chiefly northeast-trending, steeply dipping faults are evident about the remainder of the periphery of the Grenville inlier from Wild Cove to Pikes Feeder Pond. Between Beaver River and Wild Cove, where exposure is best, these faults form northeast-trending lineaments that intersect the contact between the gneiss and supracrustal rocks but die out in either direction away from this contact. Major faults in this group occur near the lower valleys of Northwest Brook and Cloud River with the intervening terrain apparently downdropped. Deformation, which increases toward the faults, is evident in the supracrustal rocks near these faults and the gneiss is locally fractured and stained along them. The dip of the bedding in the lower part of the Bradore Formation increases somewhat irregularly in successive fault blocks from 5° near Boony Lake to 35° at Wild Cove and near Sugarloaf hill. Movement along these faults is therefore likely partly rotational.

A series of northeast-trending steeply dipping faults form the northwest boundary of the highlands, and a lineament projects northeastward beyond the interpreted contact between the gneiss and the Bradore Formation. These faults have resulted in a stratigraphic separation of approximately 300 m or more (Woodard, 1957). Grenville gneiss in the vicinity of these anomaly faults is commonly stained red and appears more altered than the adjacent Paleozoic sediments.

The Sandy Bay fault (Nelson, 1955; Woodard, 1957) may be projected northward along a northeast-trending lineament to the vicinity of Ten Mile Pond where Grenville gneiss is upthrown on its southeast side. Woodard (1957) suggested that, in the area to the south, the fault dips steeply southeast and has produced a stratigraphic separation of at least 420 m. Gneiss near the fault near Ten Mile Pond is locally much fractured and stained.

The age of the northeast-trending peripheral faults is not known, but all have involved the Bradore Formation and are therefore post-Early Cambrian. Some are known to have affected the St. George Formation (Woodard, 1957), and presumably related faults have affected the Humber Arm Formation (Nelson, 1955). The age of faulting may therefore be entirely post-Early Ordovician.

Late, northerly to easterly trending faults, along which fracturing and staining were found locally and quartz veining was observed rarely, are evident within the Grenville inlier. Many additional faults of this type may remain unrecognized because discontinuities due solely to fault movement are more difficult to recognize in the gneiss. Nevertheless it seems doubtful whether any of these faults have recorded major movements because scarps and fault valleys within the inlier have low relief differential in comparison to those about the periphery. Such faults are recognized near the north Torrent River dome, and may extend east-northeast from Torrent River valley across unmapped terrain to the satellite and Hooping Harbour plutons. A single prominent northwest-trending lineament near the north end of Pikes Feeder antiform is the only one of this orientation known to be accompanied by local fracturing and staining.

Faults in Labrador and Quebec

North- to northeast-trending faults are evident at Bradore Bay, near Red Bay and possibly between Henley Harbour and Castle Island. These faults are marked locally by topographic lineaments, breccia zones, and by spectacular staining and weathering of the gneiss. Another set of east- to northeast-trending faults are evident within the Bradore and Forteau formations, and parallel lineaments are present in the gneiss to the north. These latter faults are largely restricted to a roughly semicircular area extending between Pinware River and Baie des Belles Amours.

The Bradore Bay fault, a northerly trending fault exposed at the village of Bradore Bay, is marked by a zone of pale pink crumbly quartz-rich breccia more than 3 m thick containing pebble-like fragments of quartz-rich gneiss as well as sheared slaty fragments. The fault appears to strike approximately parallel to the gneiss and dips moderately east. Rocks in the vicinity of the fault are unusually stained red-purple and locally show colour banding similar to that found in the basal Bradore beds. North of Bradore Bay, movement along the fault appears to be confined to a belt of quartz-rich and pelitic gneiss that has been offset slightly to the west along a fault that forms the northern limit of the supracrustal strata farther east. The Bradore Bay fault is therefore probably older than the easterly to northeasterly trending faults.

The Red Bay fault forms a prominent scarp near Red Bay and its lineament can be followed for a distance of 25 km north of the town. The trace of the lineament reaches the coast west of Red Bay but a possibly related lineament, approximately parallel to the Red Bay fault, starts in the hills just west of Red Bay and is almost parallel to the coastline where it reaches the coast at Carrol Cove. Both of these lineaments locally show spectacular staining in the gneiss along their traces. A lineament roughly parallel to the northern part of the Red Bay fault is present in the valley north of Black Bay. A raised ridge along the floor of this valley contains a breccia zone at least 1 m wide in which quartz fragments up to 8 cm in diameter, and minor feldspar fragments, are present in a fine-grained siliceous matrix. The rock is shattered, with purple stain along fractures and minor epidote in veins. Similar breccia in a zone 60 m or more wide is present near the top of a ridge farther north and may represent a continuation of the same fault. A north-trending lineament just west of Barge Bay forms an unusually straight but shallow valley along which a prominent northeast-trending joint is slightly offset, suggesting a metre or more of left-lateral displacement.

In the Henley Harbour area a prominent lineament strikes northeast from the coast about 2.5 km east of Woody Cove. Although the gneiss dips gently on both sides of the lineament a pronounced change in strike occurs across it, suggesting that the lineament is a fault. The movement on this fault probably predates the Bradore Formation for, although there appears to be some parallel jointing in the Bradore Formation where it lies across the lineament, there is no similar variation in bedding. The gneiss along the east coast of Henley Island is highly shattered and friable, but similar alteration is not evident in the overlying Bradore and Lighthouse Cove formations. This may indicate that there is a fault zone of Hadrynian age passing along the coast and between Henley and Castle islands.

East- to northeast-trending faults north of Blanc Sablon and limited within the semicircular topographic feature between Pinware River and Baie des Belles Amours, are marked by a series of scarps that, in many cases, have their northwest side down-thrown producing a series of southeastward tilted blocks, and fault-damned lakes. Where the gneiss close to these faults was examined it was locally friable, fractured, and red-stained. Movement along these faults is mostly dip slip as suggested by the restricted length of the faults, the production of scarps, and minor offset of steeply dipping structures in the gneiss. A throw in excess of 90 m is evident on the fault limiting the main mass of supracrustal rocks north of Blanc Sablon. The age of faulting is clearly post-Early Cambrian because the Forteau Formation has been involved.

Faults on Belle Isle

Evidence for faulting on Belle Isle has been described in the discussion of the structure of Belle Isle. Three phases of faulting are probably present: the first responsible for thrust emplacement of the northern wedges of the Bateau Formation possibly prior to emplacement of the Long Range dyke swarm, the second responsible for horst and graben faulting at the south end of Belle Isle, and the third responsible for thrusting of the Grenville gneiss upward and against the supracrustal rocks along the western margin of the island. The latter two phases may have occurred separately in the order given, or they may have occurred concurrently during a protracted period of alternating compression and tension. In either case both later phases followed emplacement of the Long Range dyke swarm and the Lower Cambrian strata.

The pattern of faulting expressed on Belle Isle resembles that present on a larger scale in the Long Range where the age relations are not as clear. Early thrusting of the Bateau Formation prior to emplacement of the Long Range dyke swarm may be related to fault movements along the Green Head fault (though the age of the latter with respect to the dykes is not known). Graben faulting with rotation of beds between blocks that is evident at the south end of Belle Isle resembles block faulting northwest of Canada Bay. Late thrusting of the Grenville basement up and against supracrustal rocks has occurred along the western margin of Belle Isle and of the Long Range inlier.

Aeromagnetic interpretation

Aeromagnetic maps at a scale of one inch to four miles (Can. Geol. Surv., Maps 7357G and 7367G) have been used in conjunction with structural interpretation based on the current mapping to interpret aeromagnetic anomaly patterns within the Strait of Belle Isle map area. It is suggested that three geological features have been particularly important in generating these patterns within the Precambrian rocks:

(1) Granulite facies metamorphism has probably been responsible for development of the short wave length, high relief aeromagnetic pattern apparent over gneiss in the granulite terrane.

(2) The granitic rocks tend to produce higher aeromagnetic anomalies than do the surrounding gneiss. Locally, however, plutons or parts of plutons have little aeromagnetic expression and in some such areas there are grounds for believing that the granitic rocks may be thin and are underlain by gneiss at no great depth.

(3) Areas where the gneiss is thicker, either as a result of original deposition of sediments or as a result of synformal folding produce negative aeromagnetic anomalies.

Northern Long Range

Gneiss over most of the western part of the northern Long Range was affected by early high-grade metamorphism that barely reached granulite facies. The aeromagnetic patterns shown by this region are more subdued than those of the granulite terrane and the short wave length, high relief anomalies are not present. Thus it appears that high-grade metamorphism in the northern Long Range was not severe enough to produce the anomaly pattern found in the granulite terrane north of the Strait. Later, lower grade metamorphic overprints do not appear to have produced recognizably distinct anomaly patterns in the gneiss.

The northern and western megacrystic plutons, the diorite plug, and, to a lesser degree, the Cloud River pluton, all show high aeromagnetic anomalies with respect to the surrounding gneiss. The southwestern part of the Leg Pond pluton lacks the characteristic high anomalies, locally is slightly foliated, and passes southwestward into a zone of hybrid rocks. Southeast of this part of the pluton, hills are capped by one or more megacrystic granitic sills. It is possible, therefore, that this corner of the pluton consists of one or more sill-like extensions from the main body to the northeast that has been locally infolded into the gneiss.

The Hooping Harbour and satellite plutons, in contrast to the other megacrystic plutons, are characterized by low anomalies and low aeromagnetic relief. This may be due to greenschist facies metamorphic overprint or it may be due to the possible thinness of the remaining plutonic rock that makes up these bodies.

Pikes Feeder antiform is characterized by a high linear anomaly that increases in magnitude toward the southern part of the core of the antiform where magacrystic granitic rocks are most prominent. This suggests that the antiform has a 'root' of metacrystic granite.

Hybrid rocks that surround the diorite plug and extend south of the map area have an aeromagnetic relief similar to that in the granulite terrane but the amplitude of the anomalies is lower. This may indicate that the maximum metamorphic grade attained during early metamorphism increases southward. Such an interpretation is consistent with reports of fresh hypersthene in the southern part of the Long Range (Bostock and Cumming, 1973).

Large scale folding of the gneiss has been interpreted about Pikes Feeder antiform, about the Hooping Harbour and satellite plutons, and between Leg Pond and Horse Chops plutons. Low aeromagnetic anomalies which flank Pikes Feeder antiform on the south and east follow interpreted synformal axes. Farther north similar low anomalies follow the Cloud River folds which have repeated the gneiss sequence. Structural interpretation of the satellite pluton has suggested that it is rootless and occupies the core of an eastward overturned inverted anticline. The inverted crest of this fold thus lies to the west of the pluton where there is a belt of north-trending low aeromagnetic anomalies. Farther east a second belt of low anomalies follows the 'tail' of the Hooping Harbour pluton and projects northward into the east part of the 'head'. Inward dips about this elongate pluton suggest that it pinches out at depth, and its position therefore may mark a locus of thickening of the gneiss that has moved in underneath to fill the space vacated. The antiform that separates Leg Pond and Horse Chops plutons near Boony Lake is the only large antiformal structure mapped that is accompanied by a large negative aeromagnetic anomaly. The coincidence of extensively exposed quartz-rich gneiss with this anomaly may perhaps indicate that quartz-rich gneiss, which elsewhere is thin, has been tectonically thickened at the core of this antiform. It is possible therefore that many belts of negative aeromagnetic anomalies within the northern Long Range mark sites along which the gneiss has been downfolded and is therefore thicker than elsewhere.

Labrador and Quebec

In Labrador and Quebec the aeromagnetic anomaly pattern of the granulite terrane is characterized by high relief, short wave length anomalies that are common in granulite facies metamorphic terranes elsewhere. The magnitude of anomalies decreases in the eastern amphibolite facies terrane and there is a noticeable decrease in aeromagnetic relief. Large areas of particularly high magnetic anomalies are common within and along the margins of areas underlain by the more melanocratic hornblende-bearing granitic rocks.

Three areas of extreme low aeromagnetic anomalies are present in a belt stretching from Bradore Bay to the northern margin of the map area near Pinware River. The Bradore Bay anomalies occur along and to the east of easterly dipping quartz-rich gneiss with pelitic and some iron sulphide-bearing interbands. The central area of low anomalies occurs along but within the margin of the large hornblende- bearing granitic pluton, where the granitic rocks are leucocratic. The northern and largest area of low anomalies occurs in a region of poor exposure, but it is perhaps significant that most of the known occurrences of sillimanite in the gneiss north of the Strait of Belle Isle are found in the country about this area of low aeromagnetic anomalies. If the central area of low anomaly is considered to be due to similar gneiss overlain by granitic rocks, then it is perhaps possible that these three anomalous areas represent a northeasterly trending, disrupted belt of paragneiss with infolded pelitic gneiss. Although the gross elongation of the northern area of low magnetic anomalies is northeasterly, a finer pattern of aeromagnetic relief strikes west-northwest to northwest across it and is roughly parallel to foliation in the underlying gneiss and granitic rocks. The gross elongation of the anomalous area is thus parallel to the older (northeast) folding evident in the Henley Harbour region, whereas the fine pattern is parallel to northwest-trending structures that are contemporaneous with or postdate emplacement of the hornblende-bearing granitic rocks.

Tectonic history

The evolution of Precambrian rocks of the Strait of Belle Isle area proposed in this memoir is complex and evidence for some of the relationships suggested is less persuasive than for others. This evolution, summarized in Table 11, is discussed in the following paragraphs, with attention given to some alternative possibilities.

The true age of deposition of the protolith of the gneiss within the basement complex is unknown and indeed, whether all the sediment from which the gneiss was derived was deposited during the same Precambrian era is uncertain; however the similarity of lithologies and their occurrence in similar proportions (with the possible exception of minor pelitic bands) throughout most of the map area suggests that a single sedimentary sequence was involved. The gneiss has been intruded by two major sets of plutonic rocks, those of the anorthosite suite, which may have been accompanied by small, now foliated biotite-granite plutons, and the later biotite granite bodies that are mostly massive or megacrystic. Although some estimate can be made of the time of emplacement of these two plutonic suites, the interval that separated emplacement of the last of the anorthosite suite of intrusions from the biotite granite may have been short. During the late Hadrynian Era the gneiss and plutonic rocks were intruded by diabase dykes some of which are feeders to basalt flows that form an early part of the post-Grenville supracrustal rock sequence. Precambrian and early Paleozoic rocks along the eastern margin of the map area were involved in the Ordovician orogeny in Newfoundland, but little evidence of later metamorphism or deformation is known.

Events related to the anorthositic orogenic phase

The anorthosite suite of intrusions, comprising distinctive monzonitic to granitic lithologies in addition to anorthositic rocks (Emslie, 1973) is represented within the map area by mangerite, hornblende granite, and probably by smaller bodies of gabbroic rocks. Traces of granulite facies metamorphism are present throughout the western part of the map area, but this metamorphism is best developed in the granulite terrane where it clearly postdates the metagabbro intrusions but probably predates emplacement of the hornblende granite. High-grade regional metamorphism was therefore a part of the orogenic phase which produced the anorthosite suite in the Strait of Belle Isle area (although such high-grade, regional metamorphism is not a necessary accompaniment to the intrusion of this suite).

The age of emplacement of the anorthosite suite within the Grenville Province is uncertain because of high-grade metamorphism, but the Michikamau intrusion north of the Grenville front has been assigned an age of emplacement of about 1460 Ma (Emslie et al., 1976). A similar date has been suggested as defining the most likely age of emplacement of the anorthosite massifs within the Grenville Province (Wynne-Edwards, 1972). Correlation therefore suggests that the metagabbro bodies of the present map area have a minimum age of about 1460 Ma and that the surrounding gneiss is of pre-Grenville, Helikian or older age.

Early northeast-trending folds in the Henley Harbour region are older than smaller scale, more steeply plunging fold domains that surround the anorthosite suite intrusions in

Table 11. Summary of proposed geological evolution of Precambrian rocks in the Strait of Belle Isle map area

Age	Date (Ma)	Nature of Rock	Deformation	Metamorphism
Helikian and/or Aphebian	>1400	Deposition of protolith of Basement Gneiss complex (including early basic volcanic rocks?), Torrent Cove assemblage and Fourché Point schist.		
Helikian	± 1400	Intrusion of Anorthosite Suite: Metagabbro Mangerite Hornblende granite (foliated granitic rocks and diorite may have been emplaced during this interval).	Regional, northeast-trending isoclinal folds probably accompanied high grade metamorphism.	Amphibolite facies in Henley Harbour region and lower granulite facies elsewhere likely accompanied mangerite intrusions.
	± 1130	Intrusion of megacrystic and massive granitic plutons.	Diapirism, doming, local northwest-trending tight to isoclinal folding(?), development of hybrid rocks. Late northeast-trending tight folds in eastern part of map area. Phyllonitization.	Amphibolite facies in central and west parts of map area, greenschist facies in eastern part.
		Intrusion of diabase dykes of Labrador and Quebec.		
			Early faulting along the eastern margin of the Grenville inlier(?). Large scale, open, northwest-trending folds in the east only.	
Hadrynian	± 900	Deposition of Bateau Formation.	Uplift and erosion. Reverse faulting and rotation of fault blocks(?).	
	605	Intrusion of diabase dykes of the Long Range swarm, and effusion of Lighthouse Cove flows.		
late Early Cambrian	± 540	Deposition of Bradore Formation and Forteau Formation.	Normal faulting(?)	
Middle Ordovician			Development of schistosity-cleavage in dykes and supra-crustal rocks along eastern margin of the map area; emplacement of Ordovician allochthons; reverse faulting(?).	Greenschist facies overprint along the eastern margin of the map area.
Late Ordovician and later	± 434		Uplift and erosion. Normal faulting(?) Reverse faulting(?)	

the Red Bay region. They are similar in structural style to northeast-trending folds that developed in the Long Range during the metamorphic maximum. Both regional northeast-trending folds and the smaller scale fold domains of the Red Bay region may be related to the same orogenic phase, that of emplacement of the anorthosite suite of intrusions. By this hypothesis, regional northeast-trending deformation occurred during the metamorphic maximum after intrusion of the metagabbro and perhaps about the time of emplacement of the hot dry mangerite plutons. Later emplacement of the large hornblende granite plutons occurred under slightly less severe metamorphic conditions, but was perhaps responsible for the aureole of structural complexity that surrounds the anorthosite suite intrusions in the Red Bay region.

Events related to late Grenville granitic plutonism

Subsequent to anorthositic plutonism the extensive, more hydrous plutonism that occurred in the northern Long Range (1130±90 Ma, Pringle et al., 1971) was much more restricted north of the Strait of Belle Isle. Megacrystic and massive granitic plutons, responsible for local amphibolite facies (low pressure, Abukuma facies series) overprints, were diapirically emplaced. During or shortly after emplacement of these plutons, gneiss chiefly along the east margin of the area was refolded along northeasterly trending axes with folds overturned to the northwest, and the rocks involved in this folding were overprinted by greenschist facies metamorphism. Still later, open warping along northwesterly trending axes is evident in the eastern part of the Long Range and in Labrador as well.

Northwesterly trending foliation or folding is evident at several localities in Labrador and Quebec, on Belle Isle, and in the Long Range, but the age of folding at these localities is largely unknown. Northwesterly trending isoclinal folds on southwestern Belle Isle and in the north Torrent River dome (Long Range) are older than their surrounding structures and could be remnants of much older deformation. In this regard, Wynne-Edwards (1969) has shown that regional north to northwesterly folding in the Kempt Lake-Mont Laurier map area in the central part of the Grenville Province predates the predominant northeasterly trends of the Grenville orogeny and may be of Hudsonian age. Northwesterly trending deformation in Labrador and Quebec in part accompanies or postdates emplacement of the hornblende granite. Folds with north to northwesterly trends in the Northwest Brook region are thought to have formed during emplacement of the megacrystic plutons in an area beyond the deformation that affected rocks along the eastern margin of the Long Range.

Post-Grenville uplift occurred in the western part of the northern Long Range and in Labrador and Quebec at roughly 850-950 Ma, as suggested by a range of K-Ar biotite and whole rock age determinations. Uplift of the eastern gneiss may have been slightly later because K-Ar hornblende and muscovite ages from the eastern gneiss are about the same as biotite ages from the western gneiss. That late Precambrian uplift of eastern gneiss did occur is indicated by equal development of the pre-Bradore regolith on both sides of the northern Long Range.

The Long Range dyke swarm was intruded perpendicular to the present regional Bouguer gravity gradient as determined by Weaver (1967). As interpreted by Weaver in his section GH, this gradient marks the transition between a thick section of Precambrian gneiss and granitic rocks on the northwest, and a thinner section of these rocks probably overlying heavier more basic intrusive rocks to the southeast. A change in the direction of the regional gradient between the northern Long Range and Belle Isle is paralleled by a similar change in mean trend of dykes in the Long Range swarm (Fig. 64). This suggests that the crustal conditions responsible for the gravity gradient may be as old as the dykes, and hence that a zone of crustal weakness marked by thinning of the Grenville craton was already present in the area intruded by the dyke swarm at the beginning of the Paleozoic and perhaps in Hadrynian time. This zone thus provides a likely locus for intrusion of late Hadrynian dykes such as those of the Long Range swarm. Emplacement of the Long Range dyke swarm and deposition of the Bateau and Bradore clastic sequences reflect a period of change in crustal stress from relative compression to relative tension. Normal faulting, both on Belle Isle and in the northern Long Range, further indicates perhaps only slightly later extension. These events likely accompanied thinning of the crust that began in Hadrynian time and eventually lead to the opening of the proto-Atlantic in late Hadrynian or early Cambrian time (Dewey and Bird, 1971).

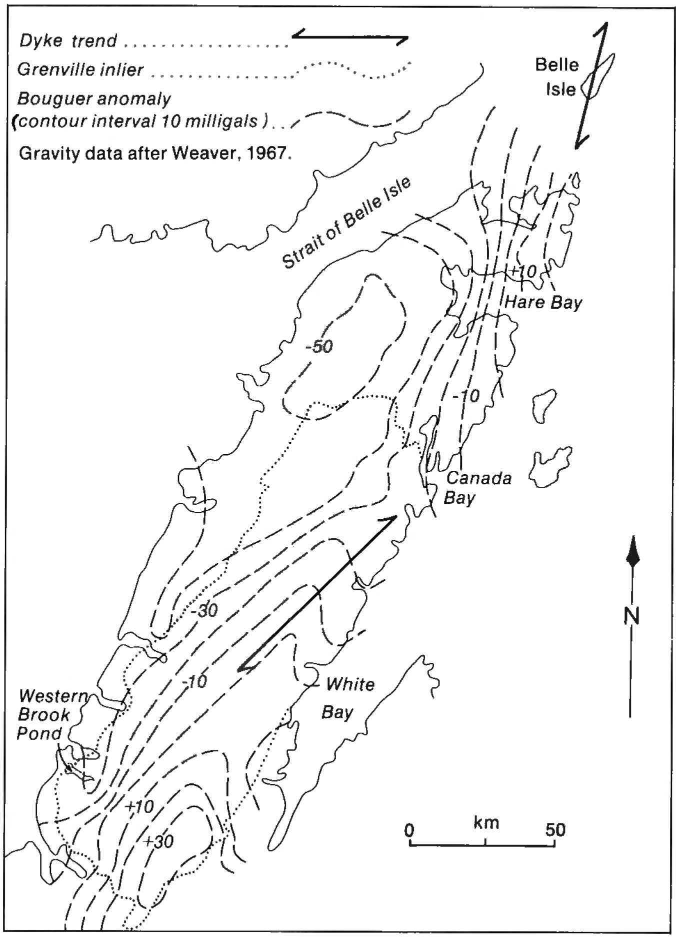


Figure 64. Map showing the average trends of dykes of the Long Range swarm and of Bouguer gravity anomalies, northwest Newfoundland.

Ordovician orogeny

Opening of the proto-Atlantic involved generation of oceanic crust and mantle, and development of an Atlantic-type continental margin (Stevens, 1970) upon which the later Hadrynian, Cambrian, and possibly early Ordovician sediments were deposited. Subsequent closing of the proto-Atlantic during the Ordovician orogeny in Newfoundland involved late Middle Ordovician westward transport (obduction) of the Ordovician klippen (Rodgers and Neale, 1963) including ophiolite complexes onto the continental margin where they persist in the allochthonous masses about Bonne and Hare bays. Poole (1967) and Stevens

(1970) have suggested that these allochthonous masses are small remnants of a much more extensive allochthonous sheet because significant amounts of ultramafic debris are wide spread in Lower Ordovician and younger clastic sediments in western Newfoundland. Final emplacement of this sheet by gravity sliding from the east would indicate that the Long Range gneiss complex was depressed relative to the root zone of the allochthonous sheet during Middle Ordovician time.

The second greenschist facies overprint in the northern Long Range (and probably that on Belle Isle as well) occurred between late Lower Cambrian and late Ordovician, and therefore presumably represents the culmination of metamorphism in this area during the Ordovician orogeny. The later Devonian (Acadian) orogeny, prominent in central Newfoundland, is thought to have had little metamorphic expression within the Precambrian highlands because the gneiss near the east coast yields K-Ar biotite ages of Cambrian to Ordovician. This Ordovician overprint affects the gneiss to the southeast to a limit that extends southwestward diagonally across the Precambrian inlier, following the zone of intrusion of the Long Range dyke swarm from the southern end of the Hare Bay allochthon toward the Bonne Bay allochthon at least as far as the hills west of Great Harbour Deep. That greenschist facies metamorphism was superimposed upon the rocks farther south is implied by observations of earlier workers (Fritts, 1953; Pontin, 1967) who recorded that the gneiss is more altered in the southern part of the Grenville inlier than in the north. It has already been suggested that the Long Range dyke swarm lies along the transition zone between a thick Precambrian gneiss section to the northwest and a thinner section of similar rocks to the southeast (Weaver, 1967). The further coincidence of the northwestern limit of late greenschist facies metamorphism with thinning of the crust may indicate that gneiss to the southeast was downwarped relative to gneiss to the northwest and thus may provide a rough indication of the extent reached by the allocthonous sheet during the Ordovician orogeny.

ECONOMIC GEOLOGY

No deposits of commercial value have developed within the Precambrian rocks of the map area. Minor occurrences of ore minerals and possible leads, that may be worth further attention, are reviewed in the following paragraphs.

Copper

A minor occurrence of chalcopyrite was reported by Alexander Murray (Chapter 2 in Murray and Howley, 1881, p. 7) in leucocratic gneiss. He states,

"At the extremity of the Cape, on the east side of the entrance to Hooping Harbour, copper pyrites were observed in a small mass of quartz, constituting a nest in the white gneiss of that part ..." No other occurrence of this type was found during the current work.

Specimens of massive chalcopyrite and ilmenite, said to have been collected from the drift in the vicinity of Red Bay, were shown to the writer by local people. As the metagabbro at Red Bay is a likely source for this mineralization, and is one of the largest basic intrusions within the map area, it may merit further attention. Although small ultramafic zones have been found within other metagabbro bodies no copper minerals were recognized.

Small amounts of malachite stain were found along a sheared diabase dyke at the falls in the creek valley north of Williamsport. No copper minerals were observed in association with other similar dykes and since most of the dykes are less than 60 m thick it seems unlikely that they would contain important economic mineralization.

Sulphides in pelitic gneiss

Small amounts of iron sulphide, chiefly pyrrhotite, are widely associated with pelitic gneiss within the map area. This suggests that conditions favourable to sulphide deposition occurred in the sedimentary environment responsible for the aluminous sediments that gave rise to the gneiss. Insofar as some of the amphibolite within the map area may be of volcanic origin it is possible that contemporary volcanism may have contributed ore metal ions to the seas from which the aluminous sediments were deposited. A more detailed delineation of the pelitic gneiss coupled with analyses for ore minerals may be worthwhile if further prospecting of the basement gneiss complex is envisaged. Interpretation of the aeromagnetic data may provide assistance in delineating infolded remnants in which these gneiss occur.

Fluorite

In the northern Long Range fluorite was found locally in the Precambrian rocks (gneiss and massive granite) along the margins of the Grenville inlier as an accessory mineral and more rarely as small crystals in vugs. Fritts (1953) reported the occurrence of fluorite in a pegmatite with radioactive minerals near Leg Pond. Woodard (1957) observed fluorite in limestone of the Table Head Formation on St. John's Island off the west shore of the Great Northern Peninsula as follows:

"Minor vein fillings and replacement masses of fluorite occur in Well Cove strata in several localities. These are normally associated with minor shear zones, and the fluorite occurs most commonly as thin coatings on the shear walls."

In view of the restriction of fluorite occurrences to the margins of the Grenville inlier, and to the vicinity of faults within the Paleozoic rocks, it is possible that these small amounts of fluorite in the Precambrian rocks have been introduced during late faulting.

Accessory fluorite is present in two samples of leucocratic hornblende granite collected about 25 km north-northwest of Pinware, north of the Strait of Belle Isle. Small amounts of secondary muscovite, observed in the granitic rocks at the southern occurrence, are possibly indicative of a weak greisenization. The sample collected was examined spectrochemically; tin and tungsten were found to be less than 0.005 and 0.01 per cent of the rock respectively. Fluorite forms in excess of 1 per cent of some beds in calc-silicate rocks about 11 km northwest of Wreck Bay.

Phosphate

Collophanite (aphanitic apatite) chips were observed in the basal glauconitic beds of the Bradore Formation at two places along the western margin of the Grenville inlier, one in the creek valley south of Leg Pond and the second in the Bradore outlier on the highlands farther south. Traces of possibly similar material were found in a specimen from near Boony Lake. The chips are largely free of inclusions and locally form up to several per cent of the beds in which they occur. Although they do not appear in sufficient abundance to be of value where found, even for local use, they suggest that phosphate-bearing deposits may exist in this vicinity, perhaps along the gneiss-Bradore unconformity. All of the known collophanite occurrences within the map area are in arkosic beds containing glauconite and it is therefore possible that these two minerals are, in this case, related in origin.

PART 2

LOWER PALEOZOIC AUTOCHTHONOUS STRATA OF THE STRAIT OF BELLE ISLE AREA

L.M. Cumming

INTRODUCTION

Location

Lower Paleozoic platformal rocks of shallow marine origin border Precambrian terrane along the north and west margins of the Great Northern Peninsula of Newfoundland from the vicinity of Norris Point in the south to Cooks Harbour in the north, a distance of some 280 km. In a considerable portion of this area, Paleozoic strata form a distinctive low-lying terrain that is commonly referred to the West Newfoundland Coastal Lowlands. This memoir describes the Paleozoic rocks which form the extreme northern part of the lowland, from the vicinity of Port au Choix on the south to Cape Norman on the north, and also those Paleozoic remnants preserved in Labrador along the north shore of the Strait of Belle Isle and on Belle Isle itself. The area in general falls within the confines of 50°30'N and 52°00'N, and 55°05'W, and 57°30'W, and includes parts of map sheets 2L, M and 12 I, P of the National Topographic System.

Recent offshore surveys have established that the seafloor beneath a significant portion of the immediately adjacent northern Gulf of St. Lawrence and Strait of Belle Isle comprises Paleozoic terrane which is essentially an extension of the Paleozoic sequences of southern Labrador and western Newfoundland. However, the offshore Paleozoic rocks are not described in any detail in this memoir, other than in the context of appraisal as a host rock to a proposed tunnel-crossing for the transmission of electrical power beneath the narrowest part of the Strait of Belle Isle from Labrador to the Island of Newfoundland.

The Strait of Belle Isle area is rich in human history. Viking explorers landed there in about 1000 A.D. and a National Historic site has been established at L'Anse-aux-Meadows to commemorate this early settlement. In the 16th century the area became a base for fishing vessels operating out of European ports, during which time the area was visited by such eminent explorers as Jacques Cartier in 1534, and later James Cook who charted part of the coastline in 1765. The Island of Newfoundland played an important role in the pioneering of early trans-Atlantic flights. In 1928, Greenly Island became the terminus of the first successful east to west aircraft flight across the Atlantic Ocean by the German seaplane "Bremen". At present, some 14 communities dot the west and north coast of Newfoundland's Great Northern Peninsula; some of the larger centres are Flower's Cove, St. Anthony, Main Brook, Roddickton, and Englee. The principal livelihood in these centres is the fishing industry, although it is conceivable that if exploration for lead and zinc deposits prove successful in future, that the mining industry could be an additional source of income for the area. Along the north shore of the Strait of Belle Isle, the communities of Lourdes du Blanc Sablon, Forteau, and L'Anse-au-Loup are located near Paleozoic terrane and livelihood in these communities is also dependant on the fishing industry.

The communities along the west and north coasts of the Great Northern Peninsula are linked to the Trans-Canada Highway by a gravel-surfaced road which terminates at St. Anthony, the home of the famous Grenfell Mission, at the northern end of the peninsula. In southern Labrador there is only 80 km of gravel-surfaced coastal road which terminates to the east at Red Bay.

A ferry service operating a summer schedule on a daily basis between St. Barbe and Blanc-Sablon is the principal transportation link between southern Labrador and the Island of Newfoundland.

Previous work

The Paleozoic and older platformal sedimentary rocks of the Strait of Belle Isle area have been examined on and off for more than a century. The contributions of some of the early visitors to the area were mainly in the form of general observations of the geological terrane and fossil collecting. Some of the more significant recordings include the work of Bayfield (1845) who first described the 7 to 9 m thickness of columnar basalts resting on conglomerate at Chateau Bay. He was also the first to collect and record fossils in the area (Bayfield, 1845, p. 457), some of which he regarded as corals, and which were later described as archeocyathids by Billings (1865), and others. This extinct phylum of animals occurs as a common reef-building organism of the Forteau Formation. In 1872, T.C. Weston (Weston, 1895) examined sections on the north side of the Strait of Belle Isle at Fox Cove and L'Anse-au-Loup and collected fossils from the upper part of the Lower Cambrian sequences exposed at those localities.

The first systematic approach to geological mapping of the area was by James Richardson of the Geological Survey of Canada in 1860 (unpublished field notes, Public Archives, Ottawa). Using Cook's admiralty charts as base maps, Richardson conducted a reconnaissance survey of the western Newfoundland coast from Strait of Belle Isle to Bonne Bay, the results of which were incorporated in Logan's Geology of Canada (1863, p. 288 - 291, and p. 865 - 880). During the course of this survey the Cambrian and Ordovician succession was divided into 17 lithostratigraphical units which were described in units from A to Q in ascending order of succession. Richardson's units A to N inclusive embraced Lower and Middle Cambrian and early Middle Ordovician rocks which he correlated with the Potsdam Group and Calciferous Formation of western and northern New York state, correlations which are now known to be only partly correct. His units O to Q which he considered equivalent to the Quebec Group of the Gaspé region are now known not be in stratigraphic sequence with underlying units A to N, but instead to be allochthonous strata. The general distribution of Richardson's rock units are illustrated on Logan's (1863) Geological Map of Canada, which was recently reprinted (Zaslow, 1975, map 7).

Perhaps the most comprehensive program for its time to be carried out in Western Newfoundland was initiated in 1910 by Schuchert and Twenhofel who carried out a two-month survey of the area. The work was continued by Schuchert and Dunbar during a two-month summer field season in 1918, followed two years later in 1920 with additional field studies by Dunbar. The results of these extensive investigations were published by Schuchert and Dunbar (1934) in Memoir 1 of the Geological Society of America.

Subsequent to the above studies, extensive fossil collections were made in the Strait of Belle Isle, the dating of which has shed considerable new light on the age and regional correlation of the lower Paleozoic rocks of that area. Some of the more significant contributions made in a biostratigraphical context are by Barnes and Tuke (1970), Cooper (1956), Cumming (1967c), Erdtmann (1971a), Fåhraeus (1970), Fong (1967), Kindle and Whittington (1965), Okulitch (1946), Walcott (1912), and Whittington (1965).

In terms of general geology and local and regional mapping of the platformal strata of Great Northern Peninsula and Belle Isle, major contributions have been made by Murray (Chap. 2 in Murray and Howley, 1881), Cooper (1937), Betz (1939), Baird (1956), Clifford and Baird (1962), Tuke (1966), Gillis (1966), Williams (1967) and Williams and Stevens (1969).

The Ordovician strata of western Newfoundland have recently become recognized as an important potential source of zinc deposits. Detailed analyses of some of the potential zinc-bearing carbonate rocks were recently completed by Collins and Smith (1972, 1973, 1975).

Offshore geological mapping of Paleozoic terrane based on shallow, high-resolution, seismic reflection surveys was carried out in the northern part of the Gulf of St. Lawrence by Shearer (1973). More recent mapping of the same region based on reflection seismic and shallow core hole drilling was conducted by Haworth and Sanford (1976).

In 1973, a study was begun to look into the feasibility of constructing a tunnel beneath the Strait of Belle Isle for the transmission of electrical power from generating stations in Labrador to the Island of Newfoundland. To appraise the geology beneath the Strait of Belle Isle at its narrowest point, extensive geological geophysical investigations were carried out by the Nova Scotia Research Foundation (1973). Shallow marine seismic surveys and diamond drill holes in Labrador and insular Newfoundland have provided considerable new information on the lower Paleozoic rocks along a narrow north-south corridor in this offshore area.

Present work

Most of this section of this memoir is based on field work carried out during the summer field seasons of 1969, 1970, and 1971 (Cumming, 1970, 1971, 1972).

Field work in 1969 was limited to a two-week period during which Cambrian and Ordovician sections were examined between Port au Port Peninsula and Port au Choix. The 1970 and 1971 seasons consisted of regional field mapping of terrain within the Strait of Belle Isle area accessible by highway, secondary roads, and trails. In 1971, the relatively inaccessible hinterland between the coast and Precambrian highlands was completed in a six-week period using a Bell G2 helicopter.

In 1974 and 1975 diamond drill cores from holes completed in Labrador and insular Newfoundland were made available to the writer through the courtesy of Teshmont Consultants Ltd., of St. John's, Newfoundland and Gull Island Power Company Ltd., Montreal. Stratigraphic data from these wells provided considerable new information on the Paleozoic succession in this particular part of the map area.

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Access to early fossil collections, field notebooks, and other information of C. Schuchert and C.O. Dunbar was made possible by Professor Dunbar during the writer's visit to the Peabody Museum, Yale University in 1972. Valuable assistance in the dating of the Ordovician succession was rendered by Dr. Rousseau Flower, New Mexico Bureau of Mines and Mineral Resources, Socorro, New Mexico, who visited the writer in the field in 1971 and made extensive cephalopod collections. Much helpful geological information was also provided by B.B. Young of Cominco Ltd., D. Bartlett of New Jersey Zinc Co. Ltd., and C.J. Cajka of Mineral and Metal Division of the Department of Energy, Mines and Resources, Ottawa. In addition, helpful information was made available by J.H. McKillop, then Director of Newfoundland Mineral Resources Division, and R.D. Hughes, Department of Geology, Memorial University.

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STRATIGRAPHY

Regional geological setting

The Lower Paleozoic and some Hadrynian rocks of the Strait of Belle Isle area form the extreme northeastern part of the Anticosti Basin, the latter an elongated structural depression that underlies most of the northern Gulf of St. Lawrence, Anticosti Island, and extends into the west coast area of Newfoundland from the vicinity of Port au Port Peninsula in the south to Cape Norman in the north (see Haworth and Sanford, 1976). In the western and central parts of the basin, Ordovician and Silurian rocks, more than 4500 m thick, dip southward from the Canadian Shield in Quebec and eventually disappear beneath allochthonous rocks of the orogen immediately offshore from Gaspé Peninsula, and in central Gulf of St. Lawrence. In the northeastern part of the basin, quite different stratigraphic and structural conditions prevail. From Port au Port Peninsula to Port au Choix, Cambrian, Ordovician, Silurian, and Devonian rocks in the northern Gulf of St. Lawrence dip southeast and are truncated by uplifted Precambrian rocks of the Great Northern Highlands[1]. Thus, a considerable part of northern Anticosti Basin lies within a half-graben containing a substantial thickness (up to 4500 m) of Paleozoic sedimentary rocks, with only the lower part of the sequence preserved onshore as remnants along the uplifted margins of the Great Northern Highlands.

In the Strait of Belle Isle area (Port au Choix to Cape Norman and on Belle Isle), upper Hadrynian, Cambrian, Upper and lower Middle Ordovician rocks up to 2100 m thick have a regional southeast dip of 1 to 2°, except where locally interrupted by predominantly northeast-trending fault

[1]Physiographic region on GSC Map 1399A, "Physiographic Map of Eastern Canada and Adjacent Areas".

systems that developed either as a result of uplift of the Precambrian Great Northern Highlands, or compression of the platform during emplacement of allochthonous rocks from the Appalachian Geosyncline to the southeast or both.

The progressive development of upper Hadrynian and lower Paleozoic sedimentary sequences within the map area can be interpreted, to a large extent, in terms of plate tectonic models. In the late Hadrynian, rifting was taking place within the Strait of Belle Isle area, while farther east new oceanic crust was produced as the original continental plate separated and moved apart to form the proto-Atlantic Ocean. During this process, the continental shelf formed a part of what is now the northern Gulf of St. Lawrence, Strait of Belle Isle and the adjacent Newfoundland coastal lowlands, with the shelf-slope boundary presumably lying immediately to the east of the Strait of Belle Isle map area. During the late Hadrynian, Cambrian, and Early Ordovician, carbonate facies prevailed on the shelf, with coarse clastic wedges derived from the Canadian Shield being deposited from time to time, the two facies giving place to siltstone, shale, and turbidites on the slope and rise areas to the east. Near the end of Early Ordovician, the continental shelf became uplifted and eroded, and principally the St. George Group carbonate developed a highly porous and karsted surface topography.

In early Ordovician, the proto-Atlantic ocean began to close. Slices of oceanic crust and slope-rise strata became stacked from east to west, and the assembled slices were pushed west toward the shelf. In early Middle Ordovician, following deposition of the Table Head limestone upon the eroded St. George carbonate, the shelf subsided and deep water shale and greywacke (derived from the moving allochthon in the east) of the Goose Tickle Formation were deposited. The assembled allochthon slid by gravity into the shale basin.

The physiography, surface drainage, and vegetation of the land surface in the Strait of Belle Isle area are phenomena which appear to be related to the underlying bedrock terrane; certain gross aspects of which are readily observable even on small-scale satellite imagery as illustrated in Figure 65. The most obvious feature, located immediately south of the Strait of Belle Isle, is a poorly drained land surface area characterized by numerous lakes and bogs, typical of the Lower Ordovician St. George terrane. A contrasting area, with only a few lakes and ponds extending from Cape Norman to Canada Bay, is underlain by early Middle Ordovician limestone of the Table Head Formation, a well-drained terrane which supports a more vigorous forest growth. A similar vegetated area outlines the Table Head limestone in a north-south elongated window through allochthonous rocks immediately south of Hare Bay. Of note also, are the outlines of the relatively thinly drift-covered and well-drained Precambrian terrane of the Great Northern Highlands. The Hare Bay Allochthon terrane is also reasonably well defined, particularly the outline of the ultramafic rocks which form the White Hills north of the entrance to Hare Bay.

The upper Hadrynian and lower Paleozoic platformal strata of the Strait of Belle Isle area consist of a variety of sedimentary rock types including conglomerate, volcanic rocks, sandstone, siltstone, shale, limestone, and dolomite that are herein classified into nine principal rock-stratigraphic divisions (Table 12). These units, with a composite thickness of about 2100 m, are representative of deposition in a wide variety of conditions ranging from subtidal to supratidal environments. The youngest of the rock units are in turn equivalent to units deposited elsewhere on the St. Lawrence Platform, and a rough correlation with lower Paleozoic rocks of Anticosti and Mingan islands is illustrated in Table 13.

General description

Labrador Group

Definition. The term Labrador series was proposed by Schuchert and Dunbar (1934) for the eroded remnants of Lower Cambrian rocks that form a narrow, intermittent fringe of outcrop along the southern Labrador coast from Bradore Bay to a point near Chateau Bay, a distance of some 110 km. Two principal rock units were recognized and defined within this series and consist of sandstone succeeded by limestone and shale; these were named the Bradore and Forteau formations, respectively. A more complete sequence of the Labrador series was recognized on the Newfoundland side of the Strait by Schuchert and Dunbar (1934) and consists of the Bradore and Forteau formations, the latter succeeded by sandstone which they named Hawke Bay Formation.

In more recent years, the series designation has been dropped in favour of the more modern term Labrador Group by Nelson (1955) and others, but the original stratigraphic boundaries of the unit have been more or less followed by most workers in the area. On the basis of investigations by Williams and Stevens (1969), Strong and Williams (1972) and this writer, it is apparent that Labrador Group is no longer an acceptable term as applied in its original context. Coarse clastics and volcanic rocks named Bateau and Lighthouse Cove formations by Williams and Stevens (1969) on Belle Isle, have an identifiable facies equivalent within the lower Bradore (Labrador Group) in the Chateau Bay area of southeastern Labrador and in rocks at Canada Bay previously referred to Cloud Mountain Formation by Williams (in Poole et al., 1970) and Clifford (1965). Consequently, it is herein proposed that the Labrador Group be redefined to include the Bateau and Lighthouse Cove formations wherever these units are recognizable in the Strait of Belle Isle area, in addition to the Bradore, Forteau, and Hawke Bay formations which have been traditionally included within its definition.

Distribution and thickness. The Labrador Group is widely distributed in the Strait of Belle Isle area. It forms the bedrock surface at two separate localities on the Labrador coast (see Map 1495A, in pocket): (i) a belt 40 km long straddling the Quebec and Labrador boundary , and (ii) a belt 25 km long in southeastern Labrador between Wreck Cove and Chateau Bay. Rocks of the Labrador Group also form the bedrock surface in a band 6 to 29 km wide around the uplifted margins of the Great Northern Highlands extending from the Highlands of St. John to Canada Bay. The Labrador Group as herein redefined is also exposed around the uplifted Precambrian core of Belle Isle, on the east, south, and west coasts of the island and adjacent islets, where at one place or another each of the formations from Bateau to Forteau equivalent are locally exposed. The distribution of the Labrador Group in the adjacent offshore areas of the northern Gulf of St. Lawrence and Strait of Belle Isle has recently been established by Haworth and Sanford (1976). In the northern Gulf, its northeast-trending erosional edge parallels the shoreline some 24 km offshore, and comes ashore in Labrador to include the remnants at Bradore and Forteau bays. East of Capstan Island its edge hugs the coastline a short distance offshore as far as the northeastern exit of the Strait where it swings northward onto the Labrador Shelf (Sanford et al., 1979).

From its present outcrop distribution, the Labrador Group dips to the southeast beneath the St. George Group and Table Head Formation, and is presumably present in the subsurface beneath the Paleozoic autochthonous and allochthonous rocks throughout the Strait of Belle Isle area.

The Labrador Group progressively thickens in a southeast direction, from 140 m on the Labrador coast to 390 m on the south side of the Strait of Belle Isle (drill hole near Flower's Cove), and to 735 m (composite) on Belle Isle with only parts of the overall sequence exposed at the latter locality. An estimated total thickness on Belle Isle is in the order of 990 m.

Lithology. The Labrador Group contains a variety of rock types consisting of conglomerate, quartzite, volcanics, sandstone, shale, limestone, and dolomite, a summary description of which follows under the headings of the individual formations.

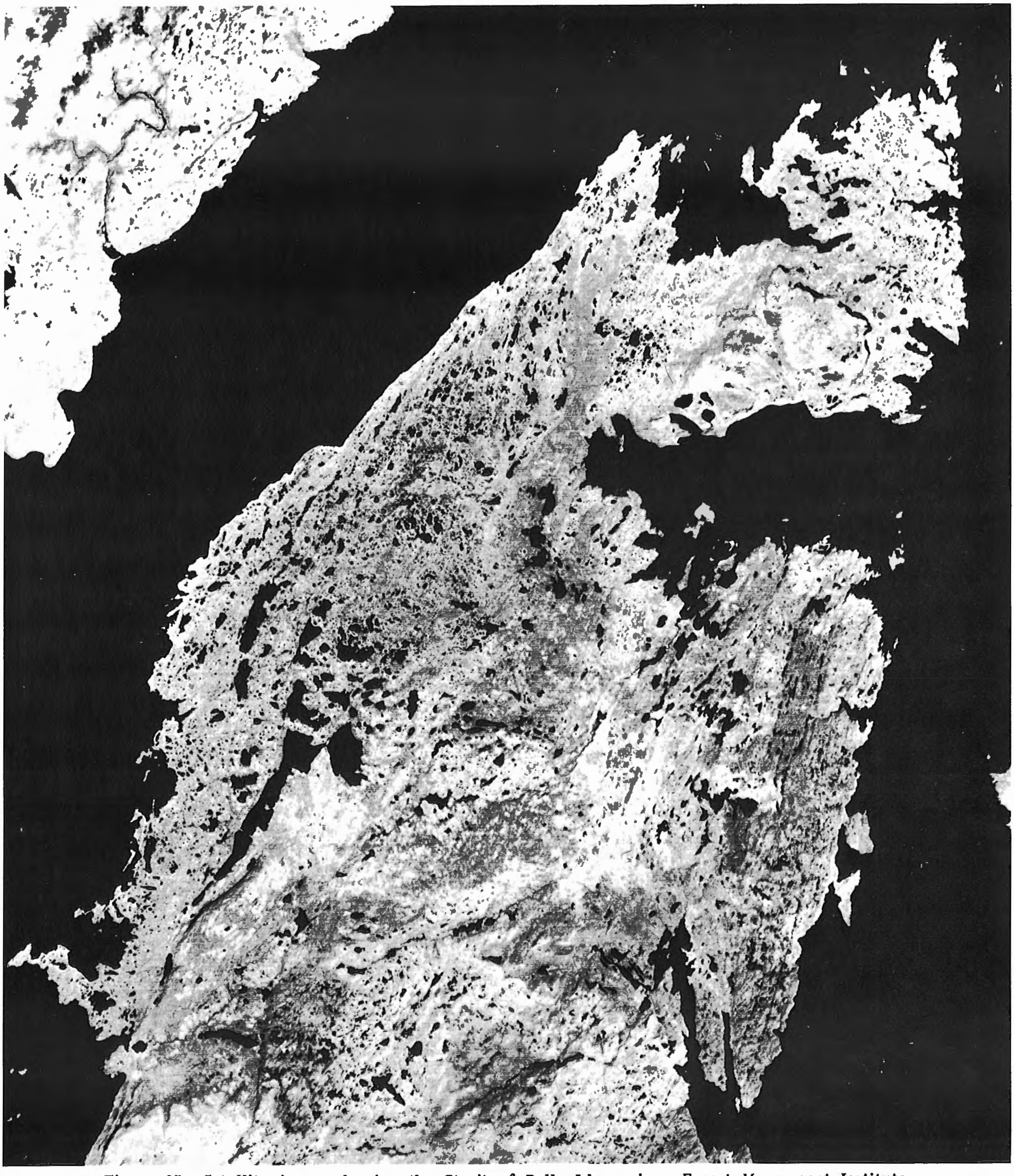


Figure 65. Satellite image showing the Strait of Belle Isle region. Forest Managment Institute, (Landsat image E-1573-14082).

The origin of this suite of rocks is considered to be closely related to the processes of rifting, distension, and crustal separation of the North American and European plates in Hadrynian and early Paleozoic time. The continuous depositional sequence from Bateau through Lighthouse Cove to Bradore formations successively would appear to be of mainly continental origin and of a lithological character (conglomerate, volcanics, redbed clastics) in keeping with the type of sedimentation associated with the initial processes of continental breakup and separation.

The younger rock units of the Labrador Group (Forteau and Hawke Bay formations) are of marine origin and the product of deposition on a broad shallow continental shelf, the latter trending in a northeast direction across what is now the northeast Gulf of St. Lawrence, southeastern Labrador and the Great Northern Peninsula. During this period the proto-Atlantic ocean to the east was in the spreading stage with carbonate, coarse clastics, and minor shale forming on the shelf, and presumably giving place to shale, siltstone, and turbidites in the shelf slope and rise environment east of the Great Northern Peninsula.

The lower boundary of the Labrador Group is everywhere in sharp and unconformable contact with crystalline basement rocks of Helikian and earlier age. On Belle Isle and in the Chateau Bay area (too small to show on Map 1495A) conglomerate of the Bateau Formation lies directly on the basement. Locally on Belle Isle volcanic rocks of the Lighthouse Cove Formation rest on basement rocks. Elsewhere, redbed clastics of the Bradore Formation are the oldest rocks of the Labrador Group and rest directly on the basement.

The upper boundary of the Labrador Group with the overlying Eddies Cove Formation is apparently abrupt and probably disconformable. This contact has nowhere been found at the surface, but in the general southeast-dipping sequence of strata exposed east of Flower's Cove the grey and pink orthoquartzitic sandstone of the Hawke Bay was observed to be overlain by grey, finely crystalline dolomite and shale of the Eddies Cove Formation. The upper contact of the Labrador Group with the Eddies Cove Formation occurs at 72.9 m in the Yankee Point core (see Appendix).

Age and correlation. The lower part of the Labrador Group (Bateau, Lighthouse Cove and Bradore formations) is unfossiliferous and has thus not been dated except for the tentative ages of Hadrynian to Early Cambrian, a designation based solely on stratigraphic position beneath known Lower Cambrian sequences. In contrast, the succeeding beds of the Labrador Group contain normal marine faunas including trilobites, archaeocyathids, and brachiopods which confirm an Early Cambrian age for both the Forteau and Hawke Bay formations.

The age and stratigraphic relationships of each formation comprising the Labrador Group are discussed in greater detail in the individual descriptions of formations to follow.

Bateau Formation

Definition. The name Bateau Formation was proposed by Williams and Stevens (1969) for the plutonic boulder conglomerate, quartzite, siltstone, slate, and minor volcanic rocks that lie directly on Precambrian gneiss along the east shore of Belle Isle from Greenham Bight to Eagle Cove. The type section is at Bateau Cove in the northeastern part of the island.

Stratigraphically succeeding strata (exposed elsewhere on the island) are volcanic rocks of the Lighthouse Cove Formation, the latter being represented by similar volcanics in such widely separated localities as southeastern Labrador and the Canada Bay area. The Lighthouse Cove volcanics were apparently erupted onto a highly irregular terrain. In some localities they lie directly upon Precambrian gneiss whereas in others they lie directly on relatively flat-lying redbed clastics. Whether or not these underlying clastics, which occur at Chateau Bay and Cloud Hills, are in part equivalent and have some degree of stratigraphic continuity with the type section of the Bateau is presently unknown. According to Williams and Stevens (1969) however, the lithologies represented in the Bateau Formation are similar in some respects to basal clastics elsewhere, and it may thus be reasonable to conclude that the thin unit of redbeds that underlie the volcanics at Table Head in southeastern Labrador and in the vicinity of Cloud Hills near Canada Bay (see Clifford, 1965) are a part of the same depositional cycle as the Bateau Formation of Belle Isle and in some measure stratigraphically equivalent. These Bateau-like clastics at Cloud Hills are too thin to show on Map 1495A (in pocket). This problem can only be resolved by deep test-drilling, but in the meantime, the pre-volcanic clastic sequences whether they be on Belle Isle, in southeastern Labrador, or at Canada Bay, are considered to be related and thus included in the present description of the Bateau Formation. Stukas and Reynolds (1974) from the ^{40}Ar-^{39}Ar ratio calculated an age of 605±10 Ma for the Lighthouse Cove feeder dykes. This date indicates a Hadrynian age for the underlying Bateau Formation.

Distribution and thickness. The regional continuity of the Bateau Formation is unknown, but on the basis of the considerable degree of stratigraphic overlap of immediately succeeding volcanic rocks of the Lighthouse Cove Formation on Belle Isle, the distribution of the Bateau Formation must on a regional scale also be considered to be sporadic. A highly irregular basement surface must be assumed to have prevailed on Belle Isle and elsewhere throughout the Strait of Belle Isle area during the late Hadrynian and Early Cambrian, with the deposition of coarse clastics being confined mainly to depressions on the Precambrian surface. Thus, the distribution of Hadrynian(?) - Lower Cambrian basal clastics was probably fairly uniform on the stable platform (basal Bradore?), in contrast to sporadic and locally very thick and highly variable facies being deposited east of a line extending from the Chateau Bay area of southeastern Labrador to Cloud Hills, Newfoundland, from whence their local distribution was probably controlled by a series of horst and graben structures formed as a result of initial rifting and foundering of the eastern continental margin of North America.

At the type section on Belle Isle, the Bateau has a composite thickness of about 244 m. Its thickness elsewhere is unknown, except for the occurrences of possible genetically related strata up to 7.5 m thick in southeastern Labrador near Table Head and in the Cloud Hills area near Canada Bay, where depressions on the Precambrian surface contain up to 6 m of strata.

Lithology. Because of the difficult logistics of getting to and from Belle Isle, the writer did not examine the lower Paleozoic and older sequences in that locality. For the purpose of this report the writer has essentially paraphrased Williams and Stevens (1969) lithological descriptions of the succession of formations on Belle Isle (Bateau, Lighthouse Cove, Bradore, Forteau and White Point) and attempted to fit them into a regional context with the remainder of the Strait of Belle Isle area.

The lithological composition of the Bateau Formation on Belle Isle is highly variable from place to place, but it can, in general, be divided into two lithic types: a lower conglomerate unit up to 90 m thick which is exposed for a distance of about 1 km north of Bateau Cove, and on two promontories immediately north and south of Greenham Bight; and an upper quartzite unit up to 152 m thick beginning about 1 km north of Bateau Cove and extending northward to Eagle Cove near the extreme northern end of the island.

At the type section (Bateau Cove), the lower unit, about 90 m thick, consists of pale purplish conglomerate with interbeds of purplish sandstone. The conglomerate beds contain a framework of well rounded pebbles or boulders of gneiss, granite, chert, and quartzite that vary from 10 cm to more than 0.5 m in diameter. The interbeds of sandstone, which range from 30 cm to 2 m thick, are arkosic and crossbedded. On the promontory immediately north of Greenham Bight, only about 15 m of the lower unit is exposed and there they consist of slaty quartzite in beds 30 to 120 cm thick with interbeds of crossbedded quartzite. South of Greenham Bight, the lower unit is about 74 m thick and considerably more variable in lithological composition. The section exposed at this locality in ascending order of succession is as follows: coarse conglomerate containing boulders up to 1 m in diameter (3 m); interbedded dark grey slate, pyritic siltstone, and orthoquartzitic sandstone (9 m); grey siltstone and slate with cleavage parallel to bedding and displaying distinctive kink bands (31 m); grey siltstone and slate with thin sandy interbeds (15 m); green conglomerate, composed mainly of discoidal quartz fragments from 2.5 to 20 cm in diameter and locally siltstone and sandstone boulders up to 0.6 m in diameter (1.5 m); green epidote-rich volcanic rocks (15 m). This lower unit of the Bateau Formation is cut by diabase dykes both at Greenham Bight and Bateau Cove and these dykes presumably constituted the feeder outlets for the Lighthouse Cove flood basalts which stratigraphically succeed the Bateau Formation.

The upper unit of the Bateau is exposed along the extreme northeastern shore of Belle Isle from a point beginning about 1 km north of Bateau Cove and extending to Eagle Cove. It consists of hard white quartzite in beds 30 to 60 cm thick, with an overall thickness of some 152 m. The quartzite locally displays graded bedding, and although principally composed of a quartz framework, the lower parts of some of the beds contain pebbles of granite, chert, and feldspar. As in the lower unit of the Bateau, the quartzite is cut here and there by diabase dykes.

The possibly related occurrences of clastics which underlie the volcanic flows (Lighthouse Cove equivalents) in southeastern Labrador and at Cloud Hills near Canada Bay are lithologically similar and, in a general way, similar to the much thicker lower member of the Bateau Formation as exposed on Belle Isle. However, because of the highly irregular Precambrian basement terrane that prevailed during the Hadrynian - Early Cambrian and that the basal clastics probably were local pockets of terrestrial sediments, the regional distribution of these beds is expected to be spotty and irregular. Because the basal clastics foreshadowed deposition of the overlying volcanics it would seem reasonable to imply a broad regional correlation of these beds as well as the volcanic rocks which succeed them, both of which for many years were considered a part of the Bradore and Cloud Mountains formations of southeastern Labrador and Canada Bay respectively. The recent acceptance of the term Lighthouse Cove (as proposed by Strong and Williams, 1972) for the sheet volcanics on Belle Isle and occurrences along the northeastern margin of the Precambrian Great Northern Highlands and in Labrador require, for the sake of consistency, to be excluded from the Bradore and Cloud Mountain formations respectively; and this was essentially the approach taken for the construction of the regional geological Map 1495A (in pocket). Similarly, the basal clastics beneath the Lighthouse Cove, wherever these are present, are thus also excluded from the Bradore and Cloud Mountain formations and are briefly described here along with the Bateau Formation.

Where these beds are present in southeastern Labrador (Table Head) they consist of red arkose and sandstone. Elsewhere in that area (i.e. Henley Harbour) the basal beds are normally absent due to overlap of succeeding volcanic rocks of the Lighthouse Cove, the latter lying directly upon Precambrian gneiss except for an intervening small lens of red conglomerate 1.5 m thick.

In the Cloud Hills area to the southeast similar redbeds containing a thin basal conglomerate, succeeded by arkose and sandstone, occur here and there beneath the Lighthouse Cove volcanics and outcrop locally along the margins of the latter formation. As in Labrador, the beds are discontinuous and are commonly overlapped by the succeeding volcanic flows, in which case the latter lie directly on Precambrian basement rocks.

The east-dipping, unconformable contact of the Bateau with the Precambrian on Belle Isle is well exposed at Bateau Cove and south of Greenham Bight. Elswhere on the island the Bateau is either in fault contact with the Precambrian or separated from the latter by diabase intrusions. The upper contact of the formation with the succeeding Lighthouse Cove Formation is nowhere exposed on the island.

At Table Head in southeastern Labrador and at isolated localities along the northeast margin of the Great Northern Highlands the contact of basal clastics with Precambrian gneiss and overlying volcanic rocks of the Lighthouse Cove is well exposed.

Age and correlation. No fossils have yet been found in the Bateau Formation and its age is therefore uncertain. It can be tentatively dated as Hadrynian (?) and/or Early Cambrian on the basis of its stratigraphic position at the base of what appears to have been a continuous depositional sequence gradually rising upward through a succession of units and into rocks of known late Early Cambrian age (Forteau Formation). Radiometric dating of the Long Range dyke swarm supports a late Hadrynian age of the Bateau Formation. (See description of Lighthouse Cove Formation below).

Regional stratigraphic relationships of the Bateau are also unknown, although possible correlation with similar facies in widely separated parts of the map area at Table Head, Labrador and Cloud Hills, near Canada Bay, are herein implied. The Bateau may also be equivalent to the quartzite of uncertain age on the White Islands which lie southeast of Cape Bauld, a correlation proposed by Williams and Stevens (1969) and discussed in the accompanying section by Williams and Smyth.

Lighthouse Cove Formation

Definition. The name Lighthouse Cove Formation was proposed by Williams and Stevens (1969) for the volcanic rocks which overlie Precambrian gneiss and are succeeded by arkosic sandstone of the Bradore Formation along the southeast, south, and west shores of Belle Isle. The type section is at Lighthouse Cove at the extreme southern end of the island.

Although Williams and Stevens (1969) correlated the basalts on Belle Isle with similar flows in southeastern Labrador and at Cloud Hills, Strong and Williams (1972) were the first to propose that the lavas in all three areas be included in the Lighthouse Cove Formation. Prior to this proposal, the volcanic rocks at Table Head, Henley Harbour, and on St. Peter Islands had been included in the Bradore Formation (Christie, 1951) and at Cloud Hills near Canada Bay included in the Cloud Mountains Formation (Clifford, 1965; Williams in Poole et al., 1970). The definition of the Lighthouse Cove as proposed by Strong and Williams (1972) is accepted by the writer in the present memoir.

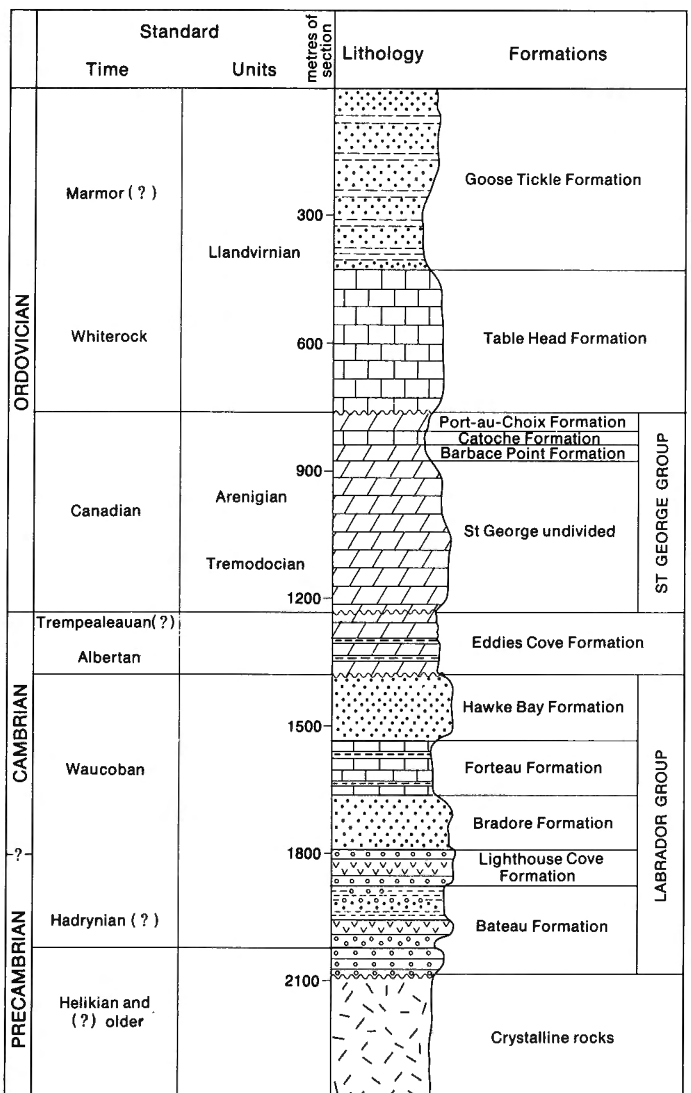
Distribution and thickness. On Belle Isle, the Lighthouse Cove Formation is exposed as isolated patches at Barbers, Lighthouse, and Blandfords coves and in areas north of Round

Head and White Point respectively, all on the southeast, south, and southwest parts of the island, in addition to a small isolated area at Lark Harbour on the west-central part of the island. In southeastern Labrador small isolated remnants of Lighthouse Cove Formation are present at Table Head, Henley Harbour, and on St. Peter Islands. Near Cloud Hills, west of Canada Bay, the formation represents the bedrock surface in a belt averaging about 1.5 km wide along the northeastern margin of the Precambrian Great Northern Highlands extending for a distance of about 24 km immediately west of Canada Bay.

The Lighthouse Cove Formation reaches its known maximum thickness of 310 m at Barbers Cove on Belle Isle, with thinner sections present north of White Point (120 m) and at the type section at Lighthouse Cove (90 m). In southeastern Labrador, the formation is 3.6 m thick at Table Head and upwards to 27 m at Henley Harbour. At Cloud Hills the formation is at least 15 m thick with its top (as in southeast Labrador) nowhere exposed.

The subsurface distribution of the Lighthouse Cove Formation is unknown, but its western depositional edge would probably in general parallel a north-south line drawn from the vicinity of Henley Harbour in southeastern Labrador to the head of the Precambrian Great Northern Highlands.

Table 12. Composite stratigraphic column of Lower Paleozoic authochthonous rocks of the Strait of Belle Isle area

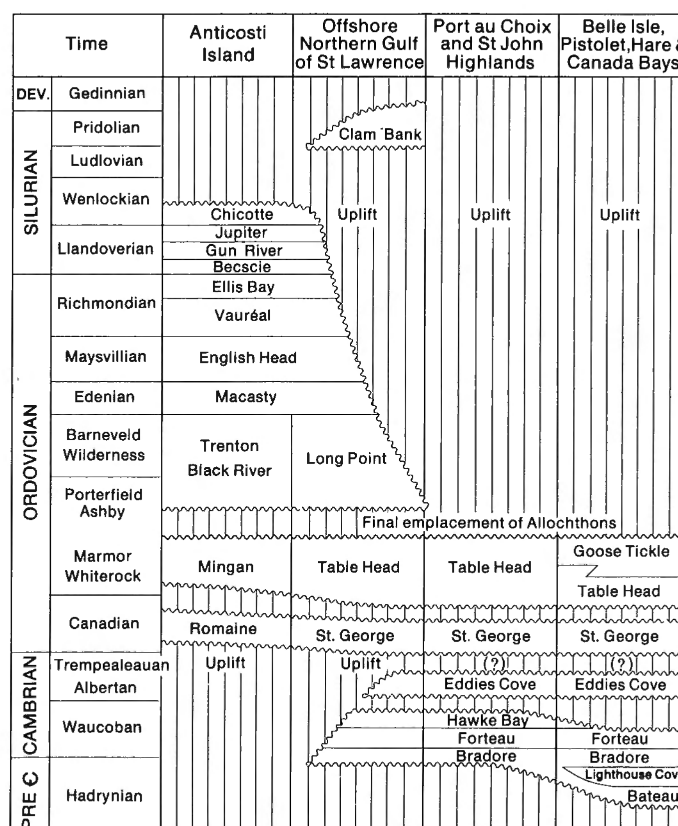


Theoretically, the volcanic flows could occur in subsurface over a substantial portion of the Strait of Belle Isle area to the east of this line, although it is not known whether these occur as a blanket or sporadically because of a highly irregular (rifted) Precambrian basement surface.

Lithology. On Belle Isle, Williams and Stevens (1969, p. 1151 - 1152) described the Lighthouse Cove Formation as black to dark green and purple to reddish brown basalts and pyroclastic rocks. Dark green flows are most abundant and are present in all of the outcrop areas. These rocks are fine to medium grained and locally amygdaloidal with calcite, quartz, chlorite, and epidote amygdules. Some of the flows, particularly at Barbers Cove, display poorly developed columnar jointing and are about 10 metres thick. Green and purple agglomerate occur interlayered with the flows at Barbers Cove and near the lighthouse at the south of the island. The agglomerate consists of fragments 5 to 15 cm in diameter of green or purple basalt with calcite amygdules. All are set in an altered green matrix. Layering within the volcanic rocks is well displayed at Barbers Cove. Several distinctive purple or green flows and agglomerate horizons can be distinguished, and the sequence also includes a 0.3 to 0.6 m bed of arkosic sandstone typical of the overlying Bradore Formation.

At Cloud Hills and in southeastern Labrador, volcanic rocks of the Lighthouse Cove Formation are lithologically similar to rocks at the type section on Belle Isle and are thus thought to have originated from a similar source. In this regard, Strong and Williams (1972) have established that there is no detectable chemical difference between the Lighthouse Cove lavas and the analyzed dykes from Belle Isle and Great Northern Peninsula of Newfoundland, and consequently any of these intrusive sources could have acted as feeders to the lavas. As on Belle Isle, the basalt at Cloud Hills and in Labrador locally display varying degrees of

Table 13. Regional correlation in Anticosti Basin



columnar jointing, the more prominent being at St. Peter Islands in southeastern Labrador where columns up to 24 m high rise vertically above the shoreline of some of the islands.

At the type locality at Lighthouse Cove on Belle Isle both the lower and upper contacts of the formation are well exposed. There, the volcanic rocks are in sharp planar contact with pink and grey basement gneiss, and are in turn conformably overlain by arkosic sandstone of the Bradore Formation. At Barbers cove, the formation is steeply dipping and the underlying basement gneiss is cut by numerous diabase dykes which nearly meet and join the volcanic flows of the Lighthouse Cove Formation. In Labrador and at Cloud Hills the volcanic rocks of the Lighthouse Cove Formation locally overlie conglomerate and sandstone (described in the Bateau Formation in this memoir) but overlap these strata here and there to lie directly on Precambrian gneiss. Although redbed clastics of the Bradore Formation would normally be expected to succeed the Lighthouse Cove volcanics in both of these localities, their contact is nowhere exposed either in Labrador or at Cloud Hills.

Age and correlation. The Lighthouse Cove volcanics form a part of what would appear to be a continuous depositional sequence (Labrador Group) which could conceivably have begun in the late Hadrynian and continued into the Early Cambrian. The oldest faunally dated rocks of this sequence is the Forteau Formation which contains a marine fauna confirming a late Early Cambrian age. Dykes of the Long Range dyke swarm appear to be feeders to the flows of the Lighthouse Cove Formation. Several radiometric whole rock K-Ar ages of the dykes on great Northern Peninsula have yielded ages of 805±35 Ma (Pringle et al., 1971); 751±100 Ma (Wanless et al., 1968); and 334±100 Ma (Clifford, 1965). The last age reflects a Paleozoic metamorphic overprint. The first two have been shown to reflect excess argon by Stukas and Reynolds (1974) who, with the ^{40}Ar-^{39}Ar method, concluded that the primary age of the dykes is 605±10 Ma, i.e. latest Hadrynian. If this last age is accepted, then the Bateau and Lighthouse Cove formations and maybe the lowest part of the Bradore Formation are Hadrynian.

Bradore Formation

Definition. The name Bradore Formation was proposed by Schuchert and Dunbar (1934) for the reddish arkose, sandstone, and minor pebble conglomerate typically exposed in cliffs east of Bradore Bay on the Labrador side of the Strait of Belle Isle. The same beds were previously referred to Division A of James Richardson as described in Logan (1863, p. 865). The formation as described by Schuchert and Dunbar includes all of the strata which lie between the gneissic rocks of the Precambrian below and fossiliferous limestone and shale of the Forteau Formation above. This definition is followed in the present memoir for the western part of the Strait of Belle Isle map area. In the eastern half of the area (east of a north-south line from Chateau Bay to the head of Great Northern Highlands) the base of the Bradore is defined at its contact with the underlying volcanic rocks of the Lighthouse Cove Formation, and this is essentially the same definition proposed by Betz (1939) for the Cloud Mountains Formation in Canada Bay region. The term Cloud Mountains as applied by Clifford (1965) and Williams (in Poole et al., 1970), which included the Lighthouse Cove and older clastics, is therefore no longer applicable and should be discontinued.

Distribution and thickness. Remnants of the Bradore Formation have a wide distribution along the north shore of the Strait of Belle Isle where they outcrop in two principal belts: (i) in a broad region some 40 km long straddling, in part, the Quebec and Labrador boundary and extending inland for a distance of up to 19 km between Bradore Bay and St. Modeste; and (ii) in a narrow belt about 24 km long bordering the southeastern Labrador coast between Wreck Cove and Henley Harbour (see Map 1495A).

On Belle Isle, the Bradore is present locally along the south shore of the island between Scotswood Cove and Blandfords Cove, with other more widely separated occurrences near Round Head and White Point on the southwest shore.

In insular Newfoundland, the Bradore forms an almost continuous outcrop belt that varies from 1 km or less to 6 km wide along the extreme northern margin of the Precambrian

Figure 66. Massive arkosic sandstone of the Bradore Formation, on northeast side of L'Anse au Loup Little Pond, Labrador. (GSC 23-1-71 LMC).

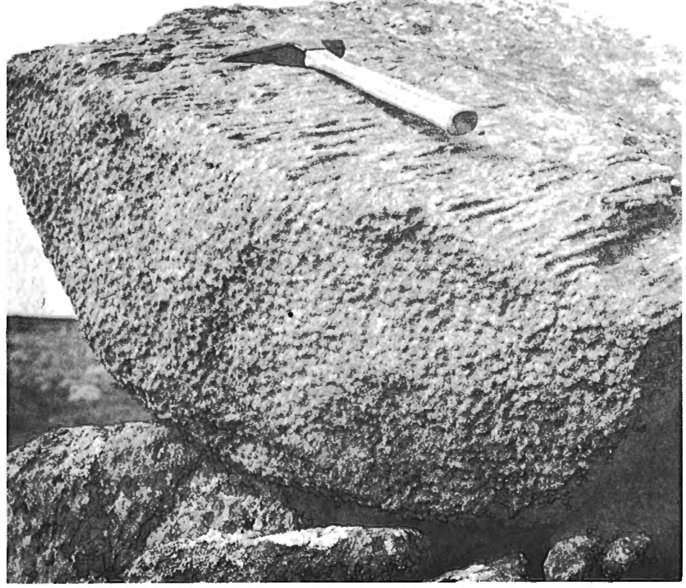


Figure 67. Block of Bradore Formation red arkosic sandstone near L'Anse-au-Loup Little Pond, Labrador showing the great number of tubes which are at right angles to the bedding surfaces. On the top of a bedding surface they occur as small knobs 0.6 cm in diameter. These inorganic structures were formerly assigned to the trace-fossil Scolithos linearis *and the rock received the popular name 'pipe-rock'. (GSC 23-10-71 LMC).*

Great Northern Highlands, extending from headwaters of Salmon River to Canada Bay. Farther south, the formation is locally exposed as faulted remnants along the western margin of the Precambrian Great Northern Highlands adjacent to the Highlands of St. John. Immediately to the west along the lower steep slopes of the uplifted Highlands of St. John, the Bradore Formation is intermittently exposed for a distance of about 24 km between Eddies Cove West and Bartletts Harbour. The Bradore sandstone also forms a narrow outcrop belt about 14 km long bordering the southeast margin of a tilted fault block which lies at the extreme southwest end of Ten Mile Lake.

Recent mapping of the northeast Gulf of St. Lawrence and southern approaches to the Strait of Belle Isle by Haworth and Sanford (1976) established the offshore distribution of the Bradore for a distance of about 320 km bordering Quebec and Labrador. Although the formation obviously continues through the Strait and onto the Labrador Shelf it has not been differentiated from other Cambrian strata in the latter regions (see Sanford et al., 1979).

The subsurface distribution of the Bradore is extensive. It everywhere dips beneath younger strata of the Forteau and presumably underlies the entire lowland area of the Great Northern Peninsula in addition to adjacent offshore regions of the northern Strait of Belle Isle and southern Labrador Shelf.

Where complete sections of Bradore have been measured in the Bradore - Forteau Bay areas of Labrador, the formation has an average thickness of about 65 m. This increases to 120 m where the sandstone lies in an anomalous local depression on the Precambrian surface in an area bordering St. Modeste. Across the Strait in Newfoundland, a drill hole completed near Yankee Point penetrated 113 m of Bradore between the depths of 350 m and 464 m where gneissic rocks of the Precambrian were encountered. Up to 90 or 120 m of Bradore strata are locally exposed on Belle Isle as recorded by Williams and Stevens (1969) and similar Bradore strata at Canada Bay have a known maximum thickness of 175 m.

Lithology. The Bradore Formation is typically a red, massive weathering, sandstone and conglomerate that is lithologically consistent over a considerable part of the Strait of Belle Isle area. At the type locality in Labrador the beds consist of medium- to coarse-grained, interbedded arkosic sandstone and orthoquartzitic sandstone containing sporadic intervals of quartz pebble conglomerate and minor shaly sandstone. The Bradore is coarser and more arkosic at the base where the distinctive reddish colouration of the formation is most dominant. This progressively gives way upward to a finer sandstone which is light red to pink in colour containing occasional light grey to greenish grey interbeds, the latter forming a distinctive band 4.5 to 7.5 m thick at the top of the formation. The sandstone and arkosic sandstone occur in beds 10 cm to 1 m or more thick (Fig. 66); the thicker massive beds account for the near-vertical cliffs typically exposed on the Labrador side of the Strait of Belle Isle and in the Highlands of St.John.

Commonly associated with the upper more massive beds of the Bradore along the Labrador coast are local preponderences of parallel, tube-like structures that intersect the sandstones perpendicular to the bedding. These structures were described as *Scolithus linearis* by Schuchert and Dunbar (1934), but may not be worm tubes as originally suspected. In fact, such structures (Fig. 67) may be inorganic in origin and have formed by escape of natural gas bubbles rising through the unconsolidated sand, a concept proposed by Hoegbom (1915). Other characteristic sedimentary structures occurring within the Bradore include the nested or rhythmically precipitated ferruginous colour bands (Liesegang rings) which are well displayed on the north shore of Greenly Island where they vary anywhere from 30 cm to 3 m in diameter (Fig. 68).

An almost identical lithological sequence consisting of orthoquartzitic sandstone, arkosic sandstone and pebble conglomerate is exposed along the western margin of the Highlands of St. John and locally in adjacent margins of the Precambrian Great Northern Highlands in Newfoundland, although nowhere in this area is the formation exposed in its entirety.

On Belle Isle, the Bradore consists of grey to pink, purple, brown and red arkosic sandstone and siltstone containing sporadic interbeds of pebble conglomerate. The framework largely comprises quartz and feldspar with minor associations of metamorphic rock fragments and opaque iron ore grains. Although purplish coloured sandstone occurs sporadically throughout the Bradore, in this region the upper part of the formation is marked by a consistent purplish zone 1.5 to 9 m thick. Locally, near the southern tip of the island, this purplish zone is overlain by 0.6 m of coarse, brown-weathering sandstone containing fragments of iron ore (specularite) up to 15 cm in diameter, in addition to discoidal fragments of purple amygdaloidal lava in cobbles up to 30 cm long.

Bordering the Great Northern Highlands on the north, the Bradore consists of a similar lithological sequence of fine- to -medium and locally coarse-grained sandstone and arkosic sandstone which are red, purplish to brown in colour containing interbeds of dark red to black quartz pebble conglomerate. These beds constitute the Cloud Mountains Formation of Betz (1939), a term which has been discontinued in this memoir.

In Labrador and adjacent parts of insular Newfoundland, the Bradore rests with abrupt and unconformable contact on gneissic rocks of Precambrian age (Fig. 69). East of a southerly trending line extending from the vicinity of Chateau Bay in southeastern Labrador to the northern end of the Great Northern Highlands, the formation rests with abrupt but probably conformable contact on volcanic rocks of the Lighthouse Cove Formation. The upper contact of the Bradore is also reasonably abrupt but gradational with the Forteau Formation, this coinciding for the most part with the change from red or grey-green sandstone or both (Bradore) to red and grey fossiliferous limestone of the Forteau Formation.

The Bradore sandstone displays a variety of sedimentary structures including large crossbedding and megaripple marks. On Belle Isle, in Labrador, and elsewhere the orientation of these structures indicates a dominant southeast flow of paleocurrents from the Canadian Shield. Because of the general lack of organic materials associated with the Bradore and the red oxidized nature of the beds, the formation may be, in large part, of continental origin, facies of which presumably developed in a fluvial-deltaic environment on the borders of an ocean basin some distance to the east.

Age and correlation. The Bradore Formation is unfossiliferous and its age is therefore not precisely known. However, because of its conformable stratigraphic position beneath the Forteau Formation, the latter known to be of late Early Cambrian age, it must also be assumed to be for the most part Early Cambrian. The lower Bradore beds, however, may be Hadrynian since radiometric analyses suggest that the conformably underlying Lighthouse Cove volcanics are of latest Hadrynian age.

Forteau Formation

Definition. The name Forteau Formation was proposed by Schuchert and Dunbar (1934) for the fossiliferous limestone and shale of Early Cambrian age in Labrador and Newfoundland which conformably overlie the Bradore Formation and in turn are succeeded by the Hawke Bay Formation. The beds were originally defined as unit B by James Richardson as described in Logan (1863, p. 865).

Figure 68. Colour banding of Leisegang rings, caused by ferruginous staining, forms a concentric pattern in flat-lying arkosic sandstone beds of the Bradore Formation on the north coast of Greenly Island. (GSC 5-1-69 LMC).

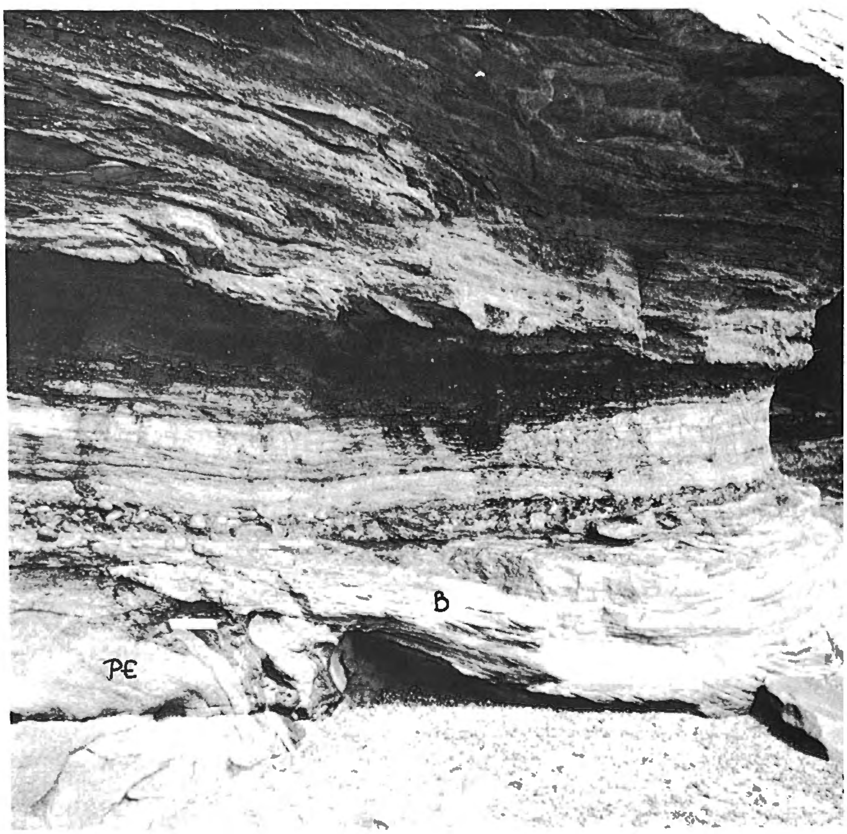


Figure 69. Basal beds (B) of Bradore Formation (arkose and conglomerate) deposited upon the peneplained surface of Precambrian granitic gneiss (PЄ); south coast of Labrador, northeast shore of Baie-Blanc-Sablon, Quebec, at 57°07'30"W. Note the coarse conglomerate layer near the base of the Bradore Formation. (GSC 201248-E).

The type section is in a stream valley north of L'Anse-au-Clair, Labrador. There the top of the formation has been eroded; consequently a more ideal type section would have been the cliffs along the western margin of the Highlands of St. John in insular Newfoundland, where a complete Forteau sequence including its upper and lower contacts is well exposed.

In the southeastern part of the map area near Canada Bay, the Forteau Formation has been enlarged to embrace pinkish coloured limestone previously defined as Dèvils Cove Formation by Betz (1939), a term which, in regional context, appears to be no longer valid and should be discarded. Thus, in the Canada Bay area as well as elsewhere in the Strait of Belle Isle, the term Forteau as used by the writer applies to all of the fossiliferous limestone, shale, and argillite which lie between the coarse clastic facies of the Bradore below and the sandstone of the Hawke Bay above.

Distribution and thickness. The Forteau Formation has wide distribution in the Strait of Belle Isle area. In Labrador, it occurs on Ile au Bois, and caps the hills between Bradore Bay and West St. Modeste. The most northerly occurrence is an outlier some 19 km north of the village of Forteau.

In insular Newfoundland, the Forteau forms an irregular outcrop belt 1 to 16 km wide around the margins of the Precambrian Great Northern Highlands extending from Hawke Bay in the extreme southwest, to Canada Bay in the southeast. On Belle Isle, the Forteau outcrops intermittently from Scotswood Cove in the southern part of the island west to Round Head and White Point, and extends as far north as Lark Harbour where it forms several small islands off the western shore.

In the offshore, the formation has been mapped by Haworth and Sanford (1976) in a belt up to 13 km wide extending for 250 km along the Quebec and Labrador coast beneath the northern Gulf of St. Lawrence and Strait of Belle Isle. In fact, the bedrock surface beneath most of the Strait comprises Forteau limestone and shale, a terrane less resistant to erosion than the sandstone of either the underlying Bradore or overlying Hawke Bay. This presumably accounts, in part, for the erosional development of the strait in its present position between what is now Labrador and the island of Newfoundland.

From Labrador, the Forteau dips gently southeast beneath sandstone of the Hawke Bay Formation, and is presumably present everywhere in subsurface beneath Paleozoic rocks.

The Forteau has a maximum thickness of 56 m in Labrador where only part of the formation is preserved. It is 116 m thick in the Highlands of St. John by direct measurement, and 119 m thick in a diamond drill hole completed near Yankee Point on the Strait of Belle Isle, where the formation was encountered between the depths of 225 and 344 m. On Belle Isle, the top of the Forteau is not exposed and thus, thicknesses of 31 m at Blandfords Cove and Round Head, and possibly 46 m at White Point, as recorded by Williams and Stevens (1969), represent only part of the formation. Beds of the White Point Formation of Williams and Stevens (1969) which overlie the Forteau Formation near White Point are too thin to show as a separate unit on Map 1495A and have been included with the Forteau. At Canada Bay, the Forteau is upwards to 224 m thick indicating a progressive thickening of the formation in an eastward direction.

Lithology. In Labrador, the Forteau is composed of fossiliferous limestone containing archaeocyathids, red and green variegated shale, and sandy and oolitic limestone. In the lower part of the formation the limestone consists of numerous irregularly distributed archaeocyathid reefs in lenses up to 9 m thick, which in turn grade laterally to thin-bedded limestone with interbedded red and grey shale. Higher in the formation, the limestone is more regularly bedded, medium to coarsely crystalline and oolitic in character, and contains only minor archaeocyathid development, confined mainly to the uppermost beds of the formation. More rarely, the limestone is nodular and impure and contains thin bands of grey shale throughout.

On Belle Isle, the Forteau consists largely of grey shale and argillite with interbeds of calcareous shale with a metre or more of buff and brownish weathering dolomitic limestone uniformly present at the base of the formation.

The Forteau Formation in the Highlands of St. John contains a lithological sequence typical of most of the exposures in the western Newfoundland area. There it consists predominantly of grey and bluish grey limestone and shaly limestone, even to nodular bedded, containing thick dark grey shale interbeds near the base of the formation and blue-grey shale interbeds at the top. A characteristic of the formation in this region is the reddish colour of the basal limestone (1 m) where it rests on the Bradore Formation, a feature apparently common to the basal Forteau as far east as Canada Bay. Another common characteristic of the Forteau everywhere is the presence of oncolites (button algae) which invariably occur in the upper part of the formation, but may also be present here and there throughout the sequence.

In the Canada Bay area, the Forteau as redefined consists of fossiliferous pink limestone at its base (about 14 m thick) with interbedded greenish shaly beds, succeeded by fossiliferous greenish grey and dark grey shale, bluish black limestone, shaly limestone, and shale, all about 210 m thick.

In general the lower contact of the basal limestone of the Forteau is conformable with the sandstone and arkose of the underlying Bradore Formation. The upper contact of the Forteau is also conformable with the sandstone of the overlying Hawke Bay, or in the absence of the latter in disconformable contact with dolomite and argillite of the Eddies Cove Formation.

The Forteau Formation appears to represent the first certain marine transgression of the eastern Canada craton. The sea was shallow and thus favoured development of resistant archaeocyathid reefs, a facies which prevailed as low barrier reefs across what is now southeastern Labrador and adjacent western insular Newfoundland during the initial transgressive phase of sedimentation, as well as later in 'Forteau time' when the sea gradually began to regress. Along the inner shelf area (southern Labrador) normal marine carbonate deposition prevailed in association with the reefal facies. Farther out on the shelf in western insular Newfoundland, substantial terrigenous muds were deposited with the carbonate, the former progressively increasing eastward towards Canada Bay area and presumably eventually giving way to fine clastics at or beyond the edge of the continental shelf somewhere to the east of the present map area.

The abrupt termination of coarse clastic deposition of the Bradore and general lack of such coarse detritus in the succeeding Forteau can only have resulted from the submergence of a broad region of the southern Canadian Shield, with the Strait of Belle Isle map area lying seaward of the strandline and of the reach of river systems flowing east into the 'Forteau' sea.

Age and correlation. The Forteau Formation is richly fossiliferous, the predominant fossils being trilobites, brachiopods, gastropods and archaeocyathids of the Pacific (North American) faunal realm (Okulitch, 1946; Fong 1967; Balsam, 1973). The presence of *Olenellus logani* (Walcott), *Olenellus thompsoni* (Hall), *Bonnia senecta* (Billings), and *Bonnia parvula* (Billings) provides a direct correlation with the North American Bonnia - Olenellus zone denoting a late Early Cambrian age for the Forteau Formation (Fritz, 1973).

The redbeds, previously referred to Devils Cove (Betz, 1939) and here included as a basal unit of the Forteau, contain a predominant tentaculid fauna consisting of *Helcionella rugosa* (Hall), *Hyolithes princeps* (Hall), and *Hyolithes impar* (Ford) in addition to the trilobite *Callavia broeggeri* (Walcott). The latter fauna is more typical of the Lower Cambrian Atlantic faunal realm of southeastern Newfoundland and its correlation with the Forteau of western insular Newfoundland and Labrador therefore is purely lithological. It should be pointed out however that redbeds are a common constituent of the basal metre of the Forteau Formation in western insular Newfoundland, and their correlation with the basal redbeds in the Canada Bay area, as herein proposed, seems reasonable.

In the Siberian Platform of USSR, rocks of Early Cambrian age have been divided into three stages on the basis of archaeocyathid assemblages by Rozanov and Debrenne (1974). They considered the Forteau containing such species as *Archaeocyathus* and *Cambrocyathus* as correlative to the uppermost late Early Cambrian Alankian or Lenian stage, a correlation not unanimously agreed upon by all North American biostratigraphers.

Hawke Bay Formation (Map unit €HB)

Definition. The name Hawke Bay Formation was proposed by Schuchert and Dunbar (1934) for the 'quartzites' that overlie limestone and shale of the Forteau Formation, a unit which they considered in turn to be overlain by dolomite of the St. George series (Group). The Hawke Bay strata are essentially the same clastic beds defined as unit C by James Richardson and described in Logan (1863), p. 865), but also locally included a metre or more of the overlying dolomite exposed on the shores of Hawke Bay which constitute the lower part of Richardson's unit D. A type section of the Hawke Bay Formation was not designated by Schuchert and Dunbar, but as the most complete sequence (105 m, with base defined) occurs in the face of South Summit of the Highlands of St. John about 10 km north of Hawke Bay, that locality is here recommended as the type section for the formation.

Although the lower boundary of the 'Hawke Bay quartzites' was clearly defined by Schuchert and Dunbar (1934), the upper contact with the overlying St. George was found to be a fault. They included in their Hawke Bay upwards of 21 m of succeeding interbedded limestone and shale along the north and south shores of Hawke Bay. This interbedded limestone and shale sequence is herein regarded as the basal member of a carbonate sequence probably in excess of 66 m thick and containing a middle Middle Cambrian fauna. These beds presumably rest disconformably on the Lower Cambrian Hawke Bay sandstone, although no such erosional break was readily apparent in any of the exposures examined by the writer. On the other hand, Middle Cambrian carbonate of the Eddies Cove Formation near Canada Bay rests with disconformable contact on Lower Cambrian Forteau strata and thus an erosional discontinuity of wide regional extent exists between the Lower and Middle Cambrian sequences throughout the Strait of Belle Isle map area. On the basis of the above described time-stratigraphic relationships, the Hawke Bay as used in the present memoir applies only to orthoquartzitic sandstone that occurs in the face of South Summit of the Highlands of St. John.

Near Canada Bay, the Hawke Bay Formation apparently has no stratigraphic equivalents (perhaps due to removal by erosion) and gradually pinches out southeasterly.

Distribution and thickness. The Hawke Bay sandstone forms the bedrock surface in widely separated localities west and north of the Precambrian Great Northern Highlands in insular Newfoundland. The more prominent outcrops occur around the margins of the summits of the Highlands of St. John, and massive and resistant sandstone beds cap the Highlands. South of the Highlands, Hawke Bay sandstone also occurs in a tilted fault block where it forms an outcrop belt some 16 km long, across Hawke Bay. Offset to the south and east is another belt of strata which occurs in a northeast-trending, down-faulted block 10 km long, which extends the length of Western Brook Pond and is truncated on the north by an east-trending fault passing on the north side of Western Brook Pond. Immediately to the east of both the north and south summits of the Highlands of St. John, the sandstone is locally present but is not differentiated from the underlying strata of the Forteau Formation (Fig. 70).

To the north of the Precambrian Great Northern Highlands the Hawke Bay forms an outcrop belt up to 6 km wide beginning at Round Lake on the west and extending east for a distance of about 25 km (Fig. 71), where the formation eventually wedges out beneath carbonate and sandstone of the overlying Eddies Cove Formation.

In the offshore, the undifferentiated Hawke Bay and Eddies Cove formations form the bedrock surface in a belt up to 19 km wide beneath the northern Gulf of St. Lawrence from whence they extend through the southern part of the Strait of Belle Isle (Haworth and Sanford, 1976).

From the northern Gulf of St. Lawrence, Strait of Belle Isle and north border of the Great Northern Highlands, the Hawke Bay dips gently to the southeast beneath younger carbonate of the Eddies Cove Formation and is presumably present in the subsurface throughout the western part of the map area. The absence of the Hawke Bay at the surface in the area immediately west of Canada Bay suggests that the beds in that region have been removed by erosion during an early Middle Cambrian hiatus. The Hawke Bay in subsurface north of Canada Bay may also have been removed.

The Hawke Bay Formation is 105 m thick at South Summit in the Highlands of St. John, but the upper beds have been removed by erosion and its true thickness in this locality is not known. The most complete sequence was penetrated by a core hole near Yankee Point on the south shore of the Strait of Belle Isle where the formation has a total thickness of 153 m.

Lithology. The Hawke Bay is composed of pink and white, fine- to medium-grained orthoquartzitic sandstone. It contains varying amounts of interbedded dark grey to black, and locally green or red shale and sandy shale throughout and minor quartz pebble conglomerate. Although the shale interbeds are but a minor constituent at South Summit, they become more numerous to the north and constitute fairly thick zones (up to 3 m) in the Yankee Point core hole.

The Hawke Bay appears to rest with conformable contact on the Forteau Formation, a boundary which is well exposed in the Highlands of St. John. Its upper contact relations with the succeeding Eddies Cove Formation, however, is not well established. The contact is apparently nowhere exposed but, on the basis of the trilobite faunas contained within the Hawke Bay and Eddies Cove formations, a disconformable contact between the two must be assumed. The contact between the Hawke Bay sandstone and Eddies Cove dolomite in the core hole at Yankee Point is abrupt and probably disconformable although there is no detailed physical evidence to support or negate the presence of a disconformity in that area. On the other hand, the Eddies Cove Formation cuts out the Hawke Bay near Canada Bay.

The Hawke Bay sandstone is characterized by a variety of primary sedimentary structures such as crossbedding, ripple and desiccation marks, all indicative of a shallow subaqueous environment of deposition. The formation also contains fragments of *Olenellus* sp. and *Paterina* sp. as well as other organisms suggestive that marine conditions prevailed at least during a part of 'Hawke Bay time'. Grabau (1936, p. 533) interpreted the Hawke Bay quartzite as a regressive phase of the Lower Cambrian sea, a concept

Figure 70. Resistant quartzite of the basal part of the Hawke Bay Formation overlies less resistant limestone and shale of the Forteau Formation at Torrent River (50°40'08"N; 56°57'54"W). (GSC 7-6-70 LMC).

Figure 71. Resistant quartzitic sandstone of the Hawke Bay Formation at 51°05'47"N; 56°23'51"W in the Salmon River area. Note the massive beds up to 2 m thick and the prominent pattern of jointing; view at N25°E. (GSC 5-5-70 LMC).

supported in more recent years by more detailed biostratigraphic studies which lends credence to a regression of the sea at the close of the Hawke Bay depositional cycle. It is therefore highly probable that the clastic detritus transported by river systems originating on the Canadian Shield was deposited as strandline deposits as the Cambrian Sea regressed to the east away from the Precambrian craton.

Age and correlation. The Hawke Bay Formation contains a sparse marine fauna including the worm burrow *Bergae uria* sp. (Arai and McGugan, 1968), in addition to fragments of *Olenellus*, *Paterina*, and *Hyolithes* (Schuchert and Dunbar, 1934, p. 24). The presence of *Olenellus* confirms an Early Cambrian age for the Hawke Bay, the latter constituting the youngest formation of the Labrador Group in the Strait of Belle Isle area.

The distribution of rocks equivalent to the Hawke Bay has not yet been established beyond the limits of the present map area. However, orthoquartzitic sandstone similar to Hawke Bay formation caps Gros Morne in the Bonne Bay region some 130 km to the southwest, and is well exposed in roadcuts near Dicks Cove at the head of East Arm of Bonne Bay, Robinsons Cove, and on an unnamed point north of Mill Brook, all in the area of the Gros Morne National Park.

Eddies Cove Formation (Map unit €EC)

Definition. The name Eddies Cove Formation is herein proposed for the dolomite and minor sandy dolomite and shale which overlie the Hawke Bay sandstone in western insular Newfoundland and are in turn succeeded by dolomite of the St. George Group. The type section of the formation is at Eddies Cove on the south shore of the Strait of Belle Isle where Kindle and Whittington (1965) collected the trilobite *Elrathia* and thus established a Middle Cambrian age for the strata exposed in that area. So far as is known, nowhere in western Newfoundland are the lower or upper contacts exposed, but on the basis of lithological composition and diagnostic Middle Cambrian faunas contained therein the formation appears to be in disconformable contact with both the Hawke Bay below and the St. George above. In the eastern part of the map area bordering Canada Bay, rocks equivalent to the Eddies Cove consist of limestone and minor sandstone originally defined as Cloud Rapids Formation overlain by the Treytown Pond Formation by Betz (1939), terms, which for the purpose of this memoir, are discarded in favour of the all encompassing term Eddies Cove Formation. Thus in the Canada Bay area, the Eddies Cove Formation rests directly with regional unconformable contact on the Forteau and, although the upper contact with the St. George is nowhere exposed, the contact is assumed to be a disconformity, largely on the apparent biostratigraphic gap between the two units.

Rocks herein defined as Eddies Cove Formation are not well exposed, particularly in the western part of the map area and although the rocks were clearly defined as unit D by James Richardson (in Logan, 1863, p. 865), they were not well defined by Schuchert and Dunbar (1934), and inclusion of their lower part in the Hawke Bay and upper part in the St. George must be considered an oversight.

Distribution and thickness. The Eddies Cove Formation forms the bedrock surface in widely separated parts of the Strait of Belle Isle area. In the southwestern part of the area they are preserved in tilted, downfaulted, blocks along the southern margins of the Highlands of St. John in the vicinity of Hawke Bay. To the north and east of the Great Northern Highlands the formation forms a discontinuous outcrop belt more than 10 km wide, beginning at Round Lake on the west and extending to the head of Canada Bay where the belt narrows to 3 km in width. From the head of Canada Bay south to White Bay, the formation appears to be absent at the surface because of faulting.

Along the southeastern shore of Strait of Belle Isle, the Eddies Cove forms an extensive northeast-trending outcrop belt up to 6.5 km wide, and some 32 km long extending from Brig Bay to Eddies Cove. From this area, the formation extends into the immediate offshore beneath the Strait of Belle Isle where it is continuous both to the southwest beneath the northern Gulf of St. Lawrence and to the northeast beneath the entrance to the mouth of the Strait of Belle Isle. Its presence in the offshore was confirmed by 3 m of grey finely crystalline dolomite in a diamond drill core obtained from the sea floor some 27 km off Port au Choix (Haworth and Sanford, 1976).

Rocks described under the name of White Point Formation on Belle Isle by Williams and Stevens (1969) are somewhat lithologically similar to the redefined Clouds Rapids strata (Betz, 1939) of the Canada Bay area and are probably equivalent to the Eddies Cove Formation of Newfoundland, although no fossils have yet been found on Belle Isle to confirm or dispute such a proposed correlation. The rocks outcrop at only a single locality on Belle Isle, namely the promontory of White Point, where they are confined to a belt too small to show on the enclosed Map 1495A.

From the presently defined outcrop belts bordering and extending beneath the Strait of Belle Isle and northern Gulf of St. Lawrence, and bordering the Precambrian Great Northern Highlands, Eddies Cove strata dip east and northeast beneath the St. George and are presumably present in the subsurface beneath younger strata throughout the map area.

The thickness of the Eddies Cove in western Newfoundland is unknown, but is at least 67 m in the core hole completed at Yankee Point, a stratigraphic test hole which began in the formation. The thickness of the formation in the vicinity of Canada Bay is also uncertain because nowhere in that area is the entire sequence exposed, although 270 m is probably a reasonable estimate. Williams and Stevens (1969) estimated a thickness of about 30 m for their White Bay Formation (Eddies Cove) on Belle Isle and this probably represents only a fraction of the unit thought to be present in the immediately adjacent offshore area to the west.

Lithology. The Eddies Cove Formation varies considerably in lithological composition from place to place within the Strait of Belle Isle area, and this appears to be due to contrasting depositional environments that prevailed across the platform during its deposition. In the western part of the map area the formation consists of dolomite with interbeds of shale and shaly dolomite, and at the base, minor interbeds of sandstone and sandy dolomite. The thin- to thick-bedded, laminated dolomite is light grey, fine grained, argillaceous, and commonly contains algal and stromatolitic mounds (Fig. 72), the latter occurring in fair abundance in the coastal exposures between Brig Bay and Eddies Cove (Fig. 73). A significant part of the formation is the thinly laminated, interbedded shale and shaly·dolomite which are dark grey and locally contain white fibrous gypsum. Thin orthoquartzitic sandstone and sandy dolomite interbeds are locally present at the base of the formation, particularly in the southwestern part of the map area along the north and south shores of Hawke Bay.

Near Canada Bay, the Eddies Cove Formation consists predominantly of fine- to medium-grained, grey and bluish black limestone and dolomite containing minor nodules of black chert. In the lower 90 m of the formation the limestone contains interbeds of grey quartzitic and brown sandstone in addition to a thin band of pebble conglomerate commonly present at the base of the sequence. Probable equivalent and apparently similar beds on Belle Isle are hard siliceous sandstone, siltstone and chert with interbeds of grey fragmental limestone.

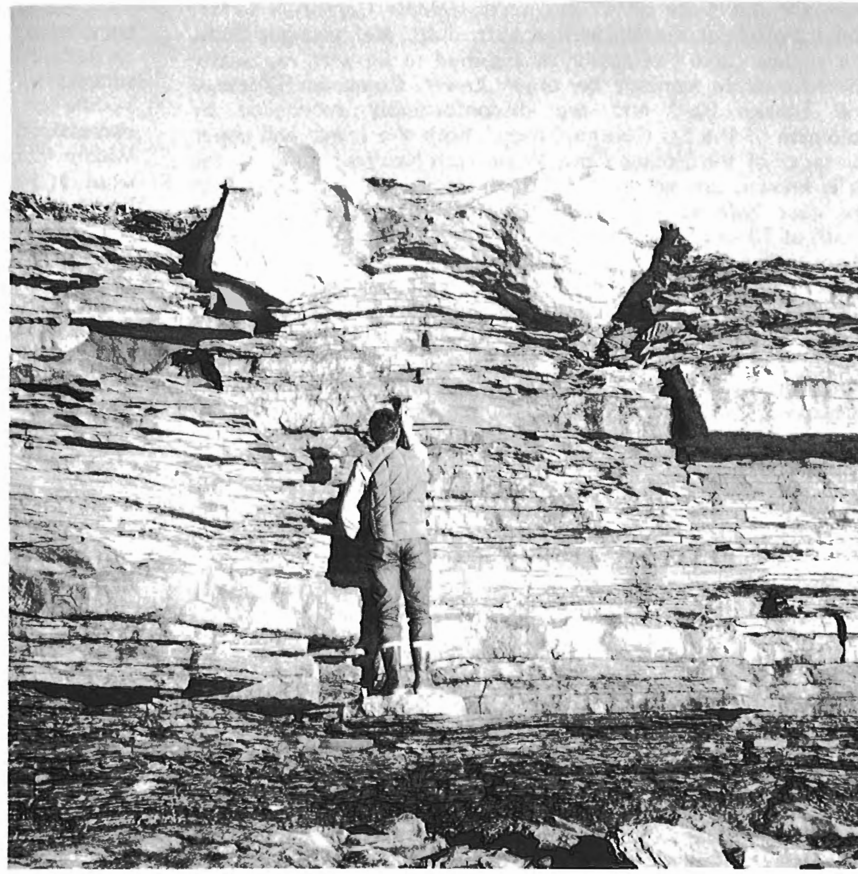


Figure 72. Carbonate mounds in flat-lying, argillaceous dolomites of the Eddies Cove Formation, exposed in the low sea cliffs north of Deadman's Cove. (GSC 4-1-71 LMC).

Figure 73. Intermound desiccation cracks and mud chips at the margin of stromatolite mounds in the Eddies Cove Formation at Flower's Island. Note the uniform mottled texture of the dolomite mounds. (GSC 6-9-71 LMC).

On the basis of its diagnostic Middle Cambrian fauna, and lithological dissimilarities with older and younger beds, the Eddies Cove Formation is assumed to lie with regionally unconformable contact on older Lower Cambrian (Forteau and Hawke Bay) and are disconformably succeeded by dolomite of the St. George Group. Both the lower and upper contacts of the Eddies Cove in western Newfoundland, so far as is known, are nowhere exposed. However, the contact in the core hole at Yankee Point which was intersected at a depth of 73 m is abrupt and clearly defined by the descending change from grey, fine- and medium-grained dolomite to grey, medium-grained, orthoquartzitic sandstone of the underlying Hawke Bay. Easterly towards Canada Bay, the contact relationships of the Eddies Cove with the underlying Forteau Formation can be observed directly. There, the contact as exposed on Cloud River is marked by a slight irregularity where the Forteau is overlain by a thin band of conglomerate which occurs at the base of the Eddies Cove Formation. In the same general region, the contact between the Eddies Cove and overlying St. George is obscured beneath glacial deposits. Paleontological evidence however suggests a substantial hiatus probably existed between the two units.

On Belle Isle, White Point beds tentatively considered as Eddies Cove are faulted against the Forteau to the east so that its contact relationships with older and younger formations are not exposed.

In the western part of the Strait of Belle Isle area, the relatively unfossiliferous Eddies Cove dolomite contains intraformational conglomerate and minor gypsum, and is algal and stromatolitic. These lithological characteristics are indicative of a tidal flat depositional environment, although the local presence of trilobites also suggests that certain of the beds were of open marine origin. The above lithologic character of the Eddies Cove changes laterally to fossiliferous limestone of subtidal origin in the Canada Bay area, a facies relationship which would indicate the gradual progression to deeper marine conditions approaching the margins of the continental shelf to the east.

Age and correlation. The Eddies Cove Formation is sparsely fossiliferous in western Newfoundland but the discovery of *Ehmania* sp. and *Solenoplurella* sp. by Whittington and Kindle (1966) in the type section at Eddies Cove confirms a Middle Cambrian age for these strata. In contrast, fossils occur in considerable profusion in equivalent beds near Canada Bay including the trilobites *Ehmania* sp., *Kootenia* sp., *Solenopleura* sp., and *Wimanella* sp. and the brachiopods *Micromitra* sp. and *Paterina* sp., which also confirm a Middle Cambrian age for the strata (Howell, 1943) described as Cloud Rapids and Treytown Pond by Betz (1939) both of which are herein included in the Eddies Cove Formation.

Although autochthonous Middle Cambrian strata undoubtedly have wide distribution along the west coast of Newfoundland to the south of the Strait of Belle Isle, the only well known occurrence is on Port au Port Peninsula (North, 1971) where the strata consist of 257 m of sandstone, siltsone, shale, and dolomite of the March Point Formation (Lochman, 1938).

St. George Group

Definition. The name St. George series was proposed by Schuchert and Dunbar (1934) for the 624 m of Lower Ordovician carbonate and minor clastics on Port au Port Peninsula that gradationally succeed the Upper Cambrian Petit Jardin Formation and are in turn disconformably succeeded by early Middle Ordovician limestone of the Table Head. The type section as described by Schuchert and Dunbar (1934) is the succession of strata exposed in the coastal area between March Point and The Gravels on the Port au Port Peninsula in southwestern Newfoundland.

As Schuchert and Dunbar traced the St. George series north along the Newfoundland's west coast and into the Strait of Belle Isle area, its definition was expanded to embrace a succession of sparsely fossiliferous strata now known to be of Middle (and possibly Late) Cambrian age. This discrepancy in correlation only recently came to light with the discovery of Middle Cambrian trilobites (Whittington and Kindle, 1966) in what is herein called the Eddies Cove Formation, and the lower boundary of the St. George as applied to the present report has been adjusted upward accordingly.

Prior to the investigations of Schuchert and Dunbar, fairly comprehensive studies of the coastal region between Cape Norman and Hawke Bay were carried out by James Richardson (in Logan, 1863, p. 289). Richardson identified and described five lithostratigraphical divisions (units E to I inclusive in ascending order of succession) which correspond to present definition of the St. George. These are essentially the same lithological divisions applied to the St. George series by Schuchert and Dunbar, except that the latter authors also included Richardson's unit D (Eddies Cove Formation of the present memoir) within their definition.

Further descriptions of the St. George by Nelson (1955) and Woodard (1957) in the Port au Choix area resulted in the introduction of the modern term St. George Group. The first attempt at formal subdivision of the St. George was by Kluyver (1975) who proposed the following three-fold division in ascending order of succession for the Port au Choix area: (i) Barbace Point Formation, (ii) Catoche Formation, and (iii) Port au Choix Formation. All three units fall within the upper half of the St. George Group and correspond to units G to I inclusive of Richardson (in Logan, 1863, p. 289). The distribution of the units defined by Kluyver (1975) as shown on enclosed Map 1495A, is confined to the relatively small area lying within a 16 km radius of Port au Choix.

In addition to the Port au Choix area, the detailed investigations of a northern concession area of Cominco have resulted in a four-fold subdivision of the St. George Group in the extreme northern part of the west Newfoundland lowlands, extending from Cape Norman to the Salmon River. These units (siliceous dolomite; tan, porous dolomite; limestone; and pseudobreccia) are shown on Map 1495A.

It must be emphasized that these units may not be in order of stratigraphic succession. In fact, some of the respective facies are known to indiscriminately interdigitate both vertically and laterally within the overall St. George sequence. Also, as the amount of dolomitization of the original carbonate beds is in many cases a function of primary porosity development, it is difficult to correlate any particular facies zone. Thus, no correlation is yet established between the formal units of Kluyver (1975) and the informal units as proposed by C.J. Cakja of Cominco. In fact, the obvious conclusion that may be drawn from Map 1495A is that little has been accomplished to date in the regional stratigraphic breakdown of the St. George, a fact attributed to the relatively poor exposures within the Strait of Belle Isle area and the reconnaissance nature of the surveys conducted to date. Further subdivision is probably possible but not until additional stratigraphic and paleontological studies can be undertaken to provide the basic framework for further detailed surface geological mapping of the St. George terrane.

Lower Ordovician carbonate rocks of the Canada Bay area were named Chimney Arm Formation by Betz (1939), a sequence which would appear to occupy an identical stratigraphic position as the St. George Group in the western part of the map area. Moreover, it is composed of a succession of dolomite and limestone that is lithologically and biostratigraphically similar to the St. George and the latter term is herein applied to the Canada Bay area to include the beds previously defined as Chimney Arm.

Distribution and thickness. The St. George Group has wide distribution in the west Newfoundland lowlands of the Great Northern Peninsula. It represents the bedrock surface in a continuous belt extending about 160 km from Hawke Bay to Cape Norman. In the southwestern part of the map area it underlies a narrow belt along the coast bordering the Highlands of St. John, from whence it gradually widens to the northeast into a broad outcrop belt some 64 km long and 40 km wide in the extreme northwestern part of the map area. From Second Salmon Pond to near Long Island, the St. George is downfaulted and the outcrop belt offset 13 km to the south. From Beaver Brook to southern Canada Bay, the St. George outcrop belt pinches from 10 km to zero against faults.

In the northern Gulf of St. Lawrence, the St. George Group extends southwesterly from St. Barbe to merge with the Romaine Formation (Haworth and Sanford, 1976). Two core holes, one about 40 km off Pointe Riche, and another 24 km off Blue Cove, confirmed the presence of the St. George.

The St. George is also undoubtedly present beneath younger Table Head and Goose Tickle formations, and the Hare Bay Allochthon in the northeastern part of the map area. There are no complete core hole sections through the St. George Group and most of the good partially exposed surface sections are interrupted by faults and thus only an estimate can be made of its total thickness in the Strait of Belle Isle area. The units (E to I inclusive) described by Richardson (in Logan, 1863, p. 289) south of the Strait of Belle Isle that fit within the present definition of the St. George have an estimated total thickness of 433 m, whereas the thickness of equivalent rocks in the Canada Bay area described under the term of Chimney Arm by Betz (1939) is in the order of some 460 to 550 m. Thicknesses of individual formations of the St. George Group in the Port au Choix area as given by Kluyver (1975) are as follows: Barbace Point 21 m; Catoche 105 m; and Port au Choix 60 m. A narrow belt of undivided St. George and Table Head occurs between 5 to 17 km southwest of Main Brook. This is an area characterized by several northeast-trending normal faults; insufficient information is available concerning the age of the carbonate rocks in this belt and therefore they are not subdivided.

Lithology. Although the St. George Group has wide distribution in the western half of the Strait of Belle Isle area, only locally are its strata sufficiently well exposed to piece together a complete stratigraphic sequence. The best exposures are on Pointe Riche Penninsula, near St. Barbe and in the coastal zone directly southeast of Cape Norman. Between these three areas probably more than 430 m of strata can be examined. These consist principally of dolomite that varies from light yellowish tan to brown and light blue-grey to dark grey. Individual beds, anywhere from a few centimetres to more than 3 m thick, are often intensely dolomitized and medium to coarsely crystalline (Fig. 74). The dolomite locally gives way to light grey, finely crystalline to micritic limestone, the latter more common in the upper part of the formation. Fossils are sparse throughout the St. George, and appear to vary considerably in number and variety throughout the section. On the other hand, stromatolites are particularly common in certain zones within the group. These usually occur in colonies with individual mounds averaging about 1 m in diameter with a vertical relief of up to 60 cm.

Near Canada Bay, the St. George Group (Chimney Arm of Betz, 1939) consists of blue, grey, and black dolomite and limestone in beds of a few centimetres to 60 cm thick. As in the western part of the map area, the beds are frequently massive and generally vary from brown, yellowish tan, white to light grey, Other similar characteristics are the presence of stromatolitic mounds, general paucity of normal marine

Figure 74. Dolomite beds of the St. George Group on the east shore of Ten Mile Lake; the St. George beds in the Strait of Belle Isle area normally have a gentle dip of 2 to 7° southeast. Here bedding is nearly vertical because of a nearby major fault which trends along Ten Mile Lake. (GSC 201827).

faunas, and occurrences of primary sedimentary structures including desiccation markings and intraformational conglomerate, which point to a common shallow marine environment of deposition.

The upper strata of the St. George Group contain the best exposed sequences, and where mapped in detail, the subunits have been added to Map 1495A. In the extreme southwestern part of the map area, near Pointe Riche Peninsula, the three units defined by Kluyver (1975) are as follows:

1. Barbace Pointe Formation. This is a resistant weathering unit comprising thinly laminated, dense, slightly silty to sandy greyish brown dolomite which contains oolitic zones and algal mounds, the latter confined mainly to the upper part of the formation. Collapse breccias, pseudobreccia, and white sparry dolomite occur throughout. A principal characteristic of the uppermost 6 m of the formation is stromatolite mounds (Fig. 75), some of which range in diameter from 1 to 3 m. The upper boundary of the Barbace Point with the softer limestone of the Catoche is sharp but apparently conformable. The base of the formation with older undivided St. George strata is nowhere exposed. Relatively flat-lying Barbace Point strata are exposed along the northwest shore near Port Saunders, where they attain a thickness of 45 m. Farther north, rocks exposed along the north shore of Port au Choix Peninsula dip southwest about 3° where they are gently warped to form a broad anticlinal - synclinal fold pair of low amplitude.

2. Catoche Formation. Overlying the Barbace Point dolomite is a succession of silty limestone, siltsone, and intraformational limestone-pebble conglomerate (23 m) overlain by algal dolomitic limestone (84 m), all of which are included in the Catoche Formation. The highly irregular anastomosing algal beds are characterized by alternating wavy bands of usually darker coloured dolomite and lighter coloured limestone, which range from 2 mm to 1 cm thick. The darker dolomite bands are consistently of algal character. The Catoche beds are well exposed along the outer shore of Port au Choix Peninsula. The formation is richly fossiliferous, especially along the coast just west of Bustard Cove, at Keppel Point, and at Hawke Flat on the south shore of Hawkes Harbour.

3. Port au Choix Formation. Conformably overlying the Catoche is the Port au Choix Formation which consists of tan, sucrosic and vuggy, light to dark greyish brown mottled dolomite 37 m thick. The dolomite is distinctly petroliferous, with thin bituminous coatings and droplets concentrated in vugs and along joint planes. Secondary white sparry dolomite in patches of pseudobreccia occurs widely throughout. Solution-collapse breccias, confined to the upper 9 m of the formation are also common. The upper contact of the Port au Choix with the Table Head is a distinct low-angle regional unconformity, the upper surface of the Port au Choix having a local topographic relief of as much as 1 m (Cumming, 1968). So far as is known, the lower boundary of the St. George Group with the Middle and Upper(?) Cambrian Eddies Cove Formation is nowhere exposed in the Strait of Belle Isle area, and their contact relationships are therefore unknown.

In the northwestern part of the map area, four rock units are delineated on enclosed Map 1495A along the eastern margin of the St. George outcrop area in a belt averaging 5 km in width and extending for some 60 km southeast from Cape Norman. These informal units are defined in probable ascending order as follows:

1. Siliceous dolomite unit is composed of a fine-grained, dense, grey to light brown laminated dolomite in beds 1 m or

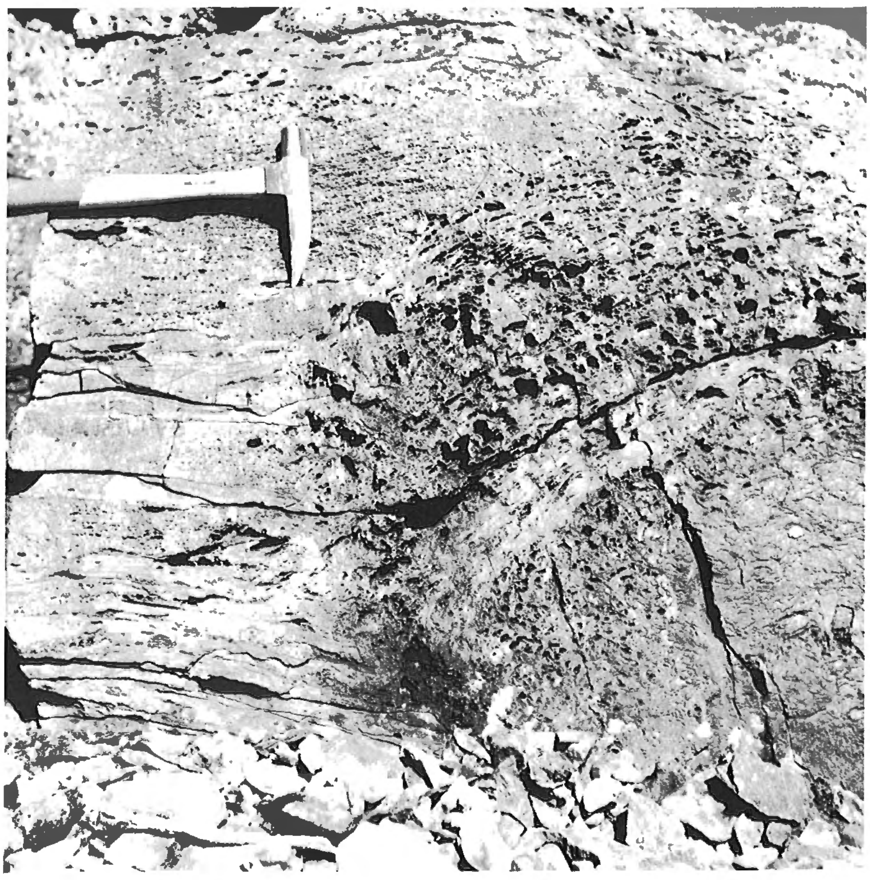


Figure 75. Porous beds at the upper part of a dolomitized stromatolite mound in the Barbace Point Formation (St. George Group) on the north coast of Port au Choix Peninsula near Barbace Point. (GSC 18-11-66) LMC).

less thick. The dolomite contains grey to brown chert blebs and stringers up to 0.6 m long. The chert is of secondary origin as evidenced by the preferential silicification of oolitic zones.

2. Tan, porous dolomite unit consists of homogeneous, sucrosic, tan to dark brown dolomite in massive beds 30 cm to 2.5 m thick. Primary textures have been obliterated for the most part by intense secondary dolomitization, although traces of original algal mottling were observed in some beds.

3. Grey upper limestone unit consists of fine- to medium-grained, light to dark brown pelletoidal limestone in massive beds 60 cm to 3 m thick. The beds are highly stylolitic and bioturbated, and contain abundant invertebrate fossils (more than 5 per cent of the rock by volume) including fragments of cephalopods, gastropods, and brachiopods.

4. Pseudobreccia unit (map unit $1O_{s1}$) comprises lenticular bodies or patches of brownish dolomite averaging 1 cm in diameter surrounded by coarse, pure white sparry dolomite, the latter constituting more than 30 per cent of the rock by volume. The pseudobreccias represent a diagenetic facies of the upper St. George limestone that was formed by replacement of micrite and coarse calcite by dolomite (Collins and Smith, 1973).

The St. George Group contains a variety of sedimentary structures (desiccation marks, intraformational conglomerate, etc.) and lithological characteristics (laminated bedding, abundance of stromatolites (Fig. 76), sparse marine faunas, etc.) that point to a shallow marine environment of deposition. Most of the sequence was presumably the product of tidal flat or shelf lagoon environments, although certain of the strata contain normal marine faunas and other lithological characteristics that indicate deposition in a deeper water (subtidal) medium. Some of the thick, massive-bedded units are of high energy marine origin and were presumably deposited on a shelf edge along the margins of a sea open to the east. The preponderance of this facies here and there throughout the St. George interbedded with low energy carbonate strata of both open marine and tidal flat origin indicates that the shoreline oscillated back and forth across the platform during Early Ordovician time.

A major hiatus followed the retreat of the shallow sea in which the St. George carbonate was deposited, with the result that the carbonate was exposed to subaerial erosion for about 5 million years as postulated by Collins and Smith (1972). Vertical uplift of the St. George carbonate platform resulted in extensive selective leaching and karstification prior to the resubmergence of the platform and deposition of the overlying Table Head Formation. Few sedimentalogical studies of the St. George carbonate have been made to date, therefore, further investigations are needed to more accurately reconstruct depositional environments and thus to more clearly evaluate the potential of the carbonate rocks for base metal deposits.

Age and correlation. The St. George Group contains a variety of marine megafossil organisms including trilobites, graptolites, cephalopods, gastropods, brachiopods, and stromatolites. Microfossils (conodonts) have also been recovered from parts of the St. George and these have been extremely useful in the dating of certain of the strata. Because of the intertidal origin of much of the carbonate sequence, however, fossils are often sparsely present and those that occur in abundance are confined to strata that appear to have been deposited during the brief, normal marine incursions of the Early Ordovician sea. In addition, intense secondary dolomitization common to certain units of the St. George (i.e. shelf edge deposits) tends to obliterate the original organic content of the carbonate and to hamper

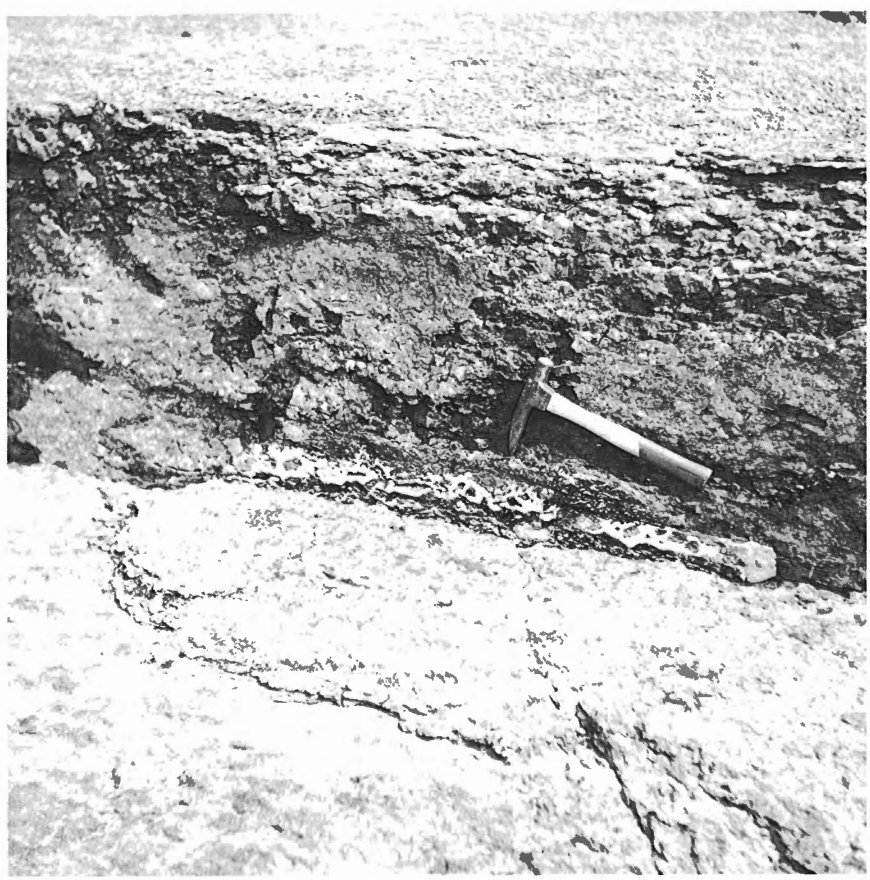


Figure 76. Close-up view of a selectively dolomitized stromatolite mound in flat-lying tan dolomite of the St. George Group at Cape Norman. White sparry dolomite and a pseudobreccia texture are developed due to the greater original porosity of the algal mound. (GSC 123944).

both megafossil and microfossil recovery. Consequently, the St. George is not yet biostratigraphically zoned, and is dated only in general terms. In the western and northern part of the West Newfoundland lowlands, the more richly fossiliferous zones occur in the upper strata of the St. George Group. Rocks of the Catoche Formation, west of Bustards Cove yielded the trilobites *Petigurus nero* and *Bathyurellus marginiatus*, and graptolites *Clonograptus* sp. cf. *C. flexilis* (Hall), an association of which indicates an Early Ordovician (Arenigian) age (Cumming, 1967c).

Cephalopods recovered in the upper part of the St. George near Cape Norman include the *Cassinoceras wortheni* fauna which Rousseau Flower (pers. comm., 1972) considered to be of late Early Ordovician (Cassinian) age. Conodonts recovered from rocks exposed in the same general vicinity (between Boat Cove and Cape Norman) and approximately the same stratigraphic position were identified by Barnes and Tuke (1970) as *Acontiodus staufferi* Furnish, *Distacodus rhombicus* Lindström, *Drepanodus homocurvatus* Lindström, *D. pandus* (Branson and Mehl), *D. simplex* Branson and Mehl, *D. toomeyi* Ethington and Clark, *Oistodus parallelus* Pander, *Oneotodus variabilis* Lindström, *Scolopodus cornutiformis* Branson and Mehl, *S. gracilis* Ethington and Clark, *S. quadraplicatus* Branson and Mehl and *Ulrichodina prima* Furnish. This assemblage also confirms an Early Ordovician (Arenigian) age for the above beds (Barnes and Tuke, 1970). Jones (1971) established a regional correlation of the Newfoundland conodonts with the *Drepanodus gracilis - Scolopodus sexplicatus* Assemblage Zone of northwestern Australia, an association which indirectly implies a late Early Ordovician (lower Arenigian) age for the upper St. George strata in the Cape Norman area.

Relatively sparse fossil collections from the Chimney Arm Formation of the Canada Bay area described by Betz (1939) included the gastropod *Ophileta* (common also to the west coast of Newfoundland) and the brachiopod *Tritoechia*, among other forms, both of which are typical of the Early Ordovician Canadian series of northeastern United States.

On the basis of present information, the lower biostratigraphic limit of the St. George is imperfectly known. However, the sequence presumably spans most of Early Ordovician time and is considered equivalent in part to the Romaine Formation of the Mingan Islands area of Quebec, the Beauharnois and Oxford formations of the Quebec Basin and Ottawa Embayment respectively, and the Beekmantown Group of New York State.

Table Head Formation

Definition. The name Table Head series was proposed by Schuchert and Dunbar (1934) for the Middle Ordovician limestone and shale of western Newfoundland that succeed the Lower Ordovician St. George, and are in turn succeeded by unfossiliferous shale and sandstone which they tentatively correlated with the upper Middle Ordovician Long Point Formation on Port au Port Peninsula. This correlation is now known to be inaccurate. The type section is the sea cliffs which extend north of Table Point on the shore of the Gulf of St. Lawrence some 32 km to the south of the present map area. The Table Head series was divided by Schuchert and Dunbar into three units which they referred to lower Table Head, middle Table Head, and upper Table Head in ascending order. Following more recent investigations of western Newfoundland by Nelson (1955), Whittington and Kindle (1963) and others, the modern term Table Head Formation became adopted and the subdivisions proposed by Schuchert and Dunbar have either been dropped or used informally.

The strata which are now defined as Table Head Formation were recognized and described in the early investigations of western Newfoundland by James Richardson (in Logan, 1863, p. 290) who recognized Divisions K and L (lower Table Head), M (middle Table Head), and N (upper Table Head).

In the western part of the map area only the lower part of the Table Head Formation is preserved, whereas to the east of a line extending from Cooks Harbour to Beaver Brook the complete sequence is exposed. In the latter area, the formation occupies a stratigraphic position between carbonate of the St. George below and shale of the Goose Tickle above. In the Canada Bay area, Betz (1939) referred to equivalents and lithologically similar limestone as the Bide Arm Formation, a term which is here dropped in favour of the term Table Head.

Distribution and thickness. The Table Head has limited distribution in the western part of the Strait of Belle Isle area. It is confined to a narrow belt along the southern part of Pointe Riche Peninsula and nearby St. John, Twin and James islands in the Gulf of St. Lawrence. From here, the formation extends offshore to the southwest beneath the northern Gulf of St. Lawrence (Haworth and Sanford, 1976) where farther west it eventually merges with the Mingan Formation, on Mingan Islands north of Anticosti Island. In the northeastern part of the map area, the Table Head forms an outcrop belt varying from less than 1 km to 16 km in width extending from Pistolet Bay to Canada Bay. It dips east beneath the Goose Tickle Formation and undoubtedly underlies the Goose Tickle and the Hare Bay Allochthon.

North of Canada Bay, in the vicinity of Freshwater Creek, the Table Head is locally exposed as a window within the allochthon. The formation is also exposed in the southern part of Hare Bay at Big Spring Inlet, Brent Islands, and other smaller islands in Shoal Arm and Southern Arm.

In western Newfoundland, the Table Head reaches a thickness of about 335 m at the type section (Cumming, 1965b). Only a fraction of this thickness is preserved near Pointe Riche where some 60 m are exposed along the south shore of the peninsula. At Canada Bay, Betz (1939) records an estimated thickness of 458 m for these limestones although no continuous sections are exposed in that area to provide an accurate measurement of their total thickness.

Lithology. The Table Head Formation is a relatively uniform lithological unit consisting of finely crystalline, dark brownish grey fossiliferous limestone, locally brecciated limestone, and minor dolomitic limestone. The limestone weathers brownish grey. Some beds are algal and commonly weather to form massive cliff sections (Fig. 77). Thin, grey to black carbonaceous shale interbeds and partings are common throughout the formation and these gradually increase in number upward toward the top of the formation. A common characteristic of the Table Head on weathering is to produce a sharp hackly gravel which affords an excellent aggregate for road construction.

Some Table Head strata strongly resemble certain limestone interbeds of the underlying St. George Group. In the absence of fossils, it may be necessary to rely on the predominant shale partings and interbeds in the Table Head and the more likely presence of primary sedimentary structures of shallow marine origin in the St. George to differentiate between the two in some of the more isolated outcrop localities. An additional characteristic of the Table Head, particularly in the region west of Hare Bay and Main Brook, is the presence of numerous sink holes and other extensive evidence of subterranean or underground drainage within that formation.

The Table Head Formation overlies the St. George Group disconformably. In western Newfoundland this relationship is particularly well exposed on Port au Choix Peninsula (Fig. 77), and to the northeast the same contact is also visible on Schooner and Burnt islands, and near Boisee

Figure 77. Disconformity between the St. George Group and the Table Head formation at Port au Choix Peninsula. Note the slight open folding of the St. George strata which occurred prior to the deposition of the cliff-forming Table Head limestone and also note the joint pattern in the St. George dolomite shown underwater. (GSC 123916).

Islands. The Table Head is in turn overlain by black shale and greywacke of the Goose Tickle Formation in the extreme northeastern part of the area, but there is some uncertainty as to the contact relationships between the two. In some localities the contact is sufficiently abrupt to assume an erosional disconformity, whereas in others there is evidence to confirm a gradational relationship. The writer favours a continuous depositional sequence between the Table Head and Goose Tickle. The Goose Tickle shale and greywacke are probably equivalent of the upper Table Head shale and overlying shale and sandstone of the type section at Table Point.

The Table Head Formation is abundantly fossiliferous, finely crystalline, and in general displays characteristcs typical of subtidal open marine environments. This is in sharp contrast to the underlying Cambrian and Lower Ordovician carbonate-clastic sequences which are in large part of tidal-flat origin, and the sudden more widespread continental submergence is undoubtedly a direct consequence of depression of the continental margin just before emplacement of the Hare Bay Allochthon in early Middle Ordovician. The faunas of the Table Head (designated as the Toquima-Table Head realm by Ross and Ingham, 1970) are diverse and typical of those found elsewhere in carbonate along the outer shelf of Ordovician paleocontinental margins.

Age and correlation. The lower Table Head strata exposed in the Pointe Riche area contain a highly diverse fauna including trilobites, gastropods, cephalopods, brachiopods, pelecypods, bryozoans, and conodonts. The presence of *Hormotoma*, *Rafinesquina*, *Galbagnostus*, *Ampyxoides*, *Stegnopsis*, *Peraspis*, *Ectenonotus*, and *Cydonocephalus* confirms an early Middle and possible late Early Ordovician age (Whittington and Kindle, 1963; Whittington, 1965). This, in North American terms, corresponds largely to the Whiterock stage and possibly extending both downward and upward into the Canadian and Marmor stages respectively (Copper, 1956, p. 7,8; Bergstrom, et al., 1974, fig. 7).

Abundant conodonts occur in lower Table Head carbonaceous limestone at Little Spring Inlet, Hare Bay (Fåhraeus, 1970). The occurrence of *Amorphognathus variabilis* Seegeeva and other species of the same genus suggest a correlation with the lower part of the Llanvirn (Fåhraeus, 1970, p. 2068).

Goose Tickle Formation

Definition. The name Goose Tickle slate was proposed by Cooper (1937) for the early Middle Ordovician black shale (slate), siltstone, and greywacke of northeastern Strait of Belle Isle area that overlie Table Head limestone and are in turn tectonically overlain by the Hare Bay Allochthon. The type section is located on the shores of Goose Tickle at the western part of Hare Bay. At Canada Bay, equivalent strata were named Englee Formation by Betz (1939), but that term has been gradually dropped from use in favour of the more modern term Goose Tickle Formation as applied by Tuke (1968).

The Goose Tickle is the youngest of the autochthonous rocks in the Strait of Belle Isle area, and is structurally overlain by rocks of the Hare Bay Allochthon.

Distribution and thickness. The Goose Tickle Formation occupies an outcrop belt 6 to 13 km wide extending from Pistolet Bay to Hare Bay, gradually narrowing to less than 0.5 km along the southwestern margin of th Hare Bay Allochthon between Hare Bay and Canada Bay. The formation is also present on a few small islands in Pistolet and Sacred bays, as well as the south shore of Howe Harbour and Northwest Arm. Between Hare Bay and Canada Bay it is again present in a concentric halo-shaped outcrop belt around Table Head strata, where the two formations are exposed in an elongated window through the Hare Bay Allochthon. In subsurface, the formation undoubtedly underlies all of the onshore part of the Allochthon.

Because of the folded and faulted nature of much of the Goose Tickle terrane, there are no accurate measurements of the thickness of the formation, although Cooper (1937) suggested 1500 m for the type section. On the basis of direct measurements however, about 450 m are exposed at Pistolet Bay (Tuke, 1968), about 240 m at Big Spring Inlet (Smyth, 1973), and 300 m at the type section (Stevens, 1970).

Lithology. In the Hare Bay and Pistolet Bay areas, the Goose Tickle consists mainly of dark grey to black shale, siltstone, and greywacke with minor interbeds of limestone breccia near the base and conglomerate near the top of the formation. The lower strata of the sequence consists of alternating light and dark grey argillite, containing thin limestone breccia interbeds here and there approximately 18 m from the base. This is succeeded by a thick succession (100 m or more) of medium-grained greywacke, siltstone, and shale. The greywacke beds 12 to 45 cm thick are interbedded with the shale throughout most of the sequence, but gradually decrease in number towards the top of the formation where they give place to shale and siltstone (Tuke, 1966). These beds are interbedded locally with shale chip conglomerate, the latter containing poorly sorted, rounded, clasts of black and green shale and angular blocks of limestone. The youngest beds of the Goose Tickle are often obscured beneath weathered detritus of the Hare Bay Allochthon and are thus not well exposed (Tuke, 1968).

In the Canada Bay area, dark grey slate and brown siltstone form the predominant lithology of the Goose Tickle with greywacke confined mainly to the upper part of the sequence and forming only a minor constituent of the overall formation.

The lower contact of the Goose Tickle with the Table Head is generally abrupt but probably conformable, although Tuke (1968, p. 506) reported an erosional discontinuity. The upper contact of the Goose Tickle with the Hare Bay Allochthon is of tectonic origin resulting from the overriding of gravity slides moving westerly across a soft-sediment sea floor (Goose Tickle shale and greywacke). In most places, according to Williams and Smyth (this memoir), the contact is marked by 15 m or more of black shaly mélange, but at Canada Bay the units are bounded by a sharp thrust plane, and the mélange is absent.

Primary sedimentary structures including load casts, sole markings, and paleocurrent features, indicate westward transport of Goose Tickle detritus, derived from the moving Hare Bay Allochthon to the east (Stevens, 1970). The sediments were deposited in a shallow, but rapidly subsiding trough which developed upon a former carbonate shelf. What is now the northeastern part of the Strait of Belle Isle area presumably formed the western edge of the marginal trough.

Age and correlation. Normal marine faunas such as those occurring in the underlying Table Head Formation are sparse in the Goose Tickle, although a few poorly preserved brachiopods and cephalopods have been reported from outcrops on the western shore of Coles Pond, at Little Spring Inlet, and elsewhere. On the other hand, graptolites are locally abundant in the shale beds, and fairly extensive collections of these fossils have been made at Pistolet Bay (Tuke, 1968) and on the western shore of Hare Bay (Erdtmann, 1971a,b). These collections include the following forms: *Climacograptus antiquus* var. *lineatus* Elles and Wood; *C. antiquus* var. *bursifer* Lapworth; *Dicellograptus* sp.; *Dictyonema* sp; *Didymograptus* sp. cf. *D. nitidus* Lapworth; *D.* sp. cf. *D. hirundo* Lapworth; *D.* sp. cf. *D. spinosus; D.* sp. cf. *D. nicholsoni*, Lapworth; *Glossograptus cileatus* Eamons; *G. echinatus* Eamons; *Janograptus* sp.; *Phyllograptus anna*; *P. illiciformis* Hall; and *Tetragraptus quadribrachiatus* Hall.

These assemblages, corresponding to zones 9 and 10 of Berry (1960), indicates an early Middle Ordovician Llanvirnian age for the formation.

The Goose Tickle, now confined to the eastern part of the Strait of Belle Isle area, undoubtedly at one time extended much farther to the west across the platform where it intertongued with upper Table Head shale of the type section, eventually succeeding the latter to spread across much of what is now western Newfoundland and adjacent northeastern Gulf of St. Lawrence.

Pleistocene deposits

The Strait of Belle Isle area was subjected to four principal episodes of glaciation and deglaciation during the Pleistocene period. These are summarized as follows (after Grant, 1969a, 1970):

1. The southeastward advance of Laurentide ice across the Great Northern Peninsula, as evidenced by unidirectional grooving, roches moutonnées and Labrador erratics on the uplands near Roddickton and St. Anthony, and a shelly drift, derived from marine sediments in Strait of Belle Isle, that was spread widely over the area.

2. Subsequent retreat of ice, influenced mainly by a calving bay migrating northeastward along Strait of Belle Isle, proceeding concentrically inland from the present coast to an ice divide near Hare Bay, as indicated by striations and 'De Geer' moraines.

3. Readvance of ice across the lowlands and into the sea from the north flank of the Precambrian plateau of the Great Northern Highlands, fluting and drumlinizing the earlier marine deposits. A lobe of ice deployed westward into the basin of Ten Mile Lake where it deposited a massive, multiple-ridged end moraine. At the northern and eastern margins of the ice, on the other hand, the advancing ice probably abutted against the still wasting older ice-cap of the lowlands. Crossing striae near Main Brook suggest wastage again took place differentially as the sea calved the western flank.

4. Final retreat of the ice was up and onto the Precambrian plateau of the Great Northern Highlands where the ice-cap flowed actively westward as well as eastward over the height of land, leaving abundant striated roches moutonnées.

Topographically, the lowlands of the Strait of Belle Isle area vary from present sea level to about 120 m above sea level. The lowlands are bounded on the east (Canada Bay to Pistolet Bay) by uplands with elevations of 240 to 300 m; on the south by an eastward tilted fault block comprising the Great Northern Highlands rising to 600 - 750 m; and on the west by the Strait of Belle Isle, the other side of which is the dissected peneplain of southern Quebec and Labrador rising to 450 m.

The upland, peneplain and highland surfaces are mostly rugged, boulder strewn and denuded of a glacial drift cover. Consequently, most of the glacial deposits and land forms are confined to the lowland terrane. Despite the paucity of surficial cover and modification of deposits by interglacial marine invasion, a remarkable array of glacial constructional features are preserved, and these are summarized in Figure 78.

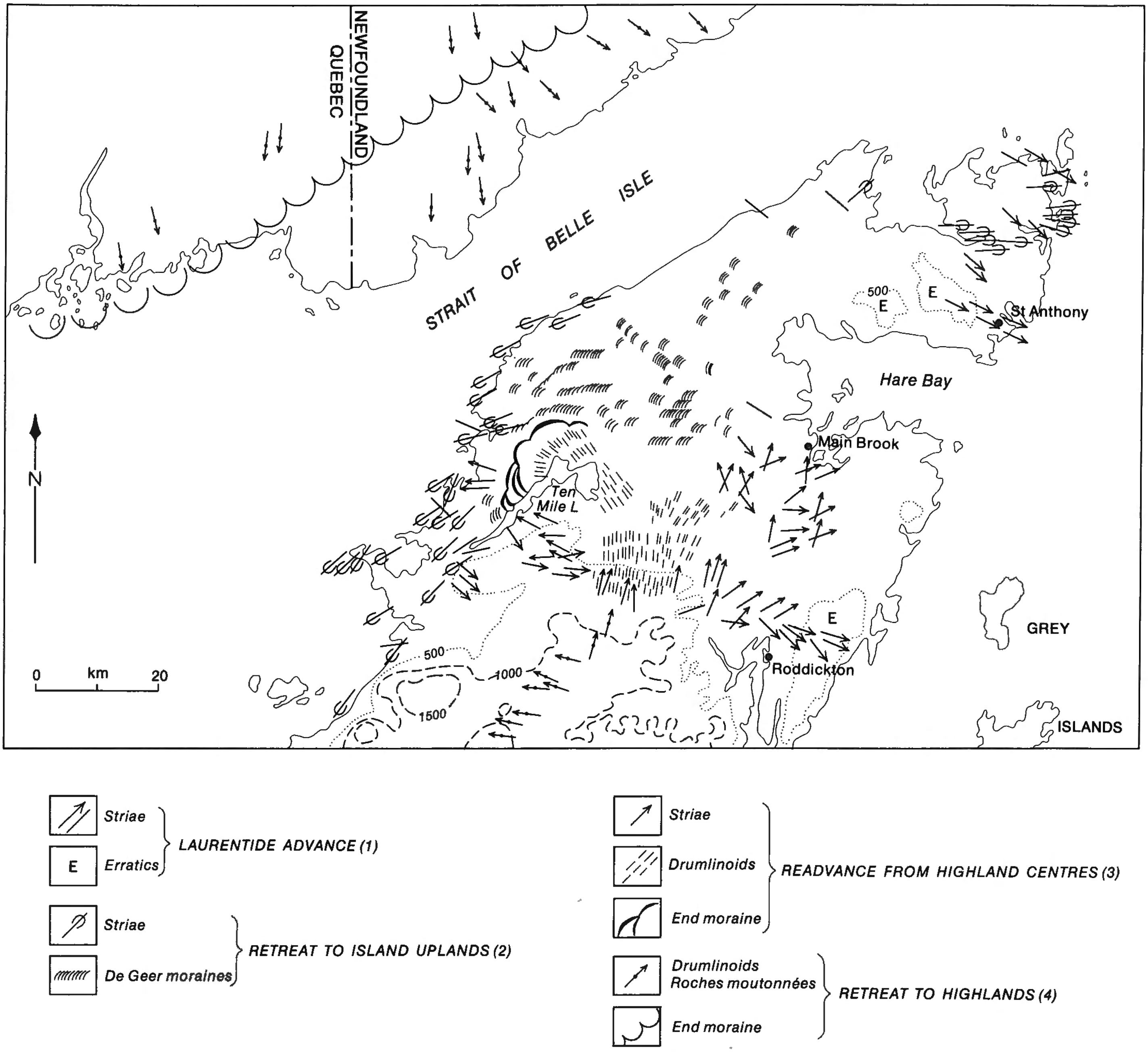


Figure 78. Glacial features of the Strait of Belle Isle area (after Grant, 1970, p. 173)

Glacial striations are well developed, especially on the carbonate rocks, and are found where a till cover has been recently removed, as by wave action or road construction. Most exposures offer evidence of the direction of ice flow, usually by the presence of miniature crag-and-tail on algal, oolitic, and coralline limestone, and pebbly sandstone, as well as by miniature plucked surfaces, stoss-and-lee, and bevelled faces.

Largest and most significant of the constructional features is an end moraine system near Ten Mile Lake. This system is 32 km long with several kettled ridges and a relief of 15 to 45 m. Large areas of numerous, parallel, minor moraines of the 'De Geer' type, representing annual deposits during ice retreat, are best seen in lakes as bouldery spits, islands, and shoals up to 0.8 km long, 150 m apart, 45 m wide, and less than 8 m high. An area of low attenuated drumlinoid ridges occurs 'up-glacier' from the end moraine. Raised strandlines are not common and are usually recognizable below 15 m, except in Quebec and Labrador where they are present up to 150 m.

Postglacial changes of relative sea level vary throughout the area. In Quebec and Labrador the marine limit is shown by excellent trimlines at the lower limit of unmodified drift and perched erratics; massive boulder beaches occur only slightly lower. A prominent ridge in the morainal belt is trimmed off at 140 m above sea level, but beyond this ice-marginal position the sea has washed nearly to 150 m, while younger moraine ridges in behind are unmodified at 123 m, thus clearly demonstrating the variation of the marine limit with the age of deglaciation.

In Newfoundland, on the other hand, no such trimlines are apparent. Shell-bearing beach sediment occurs up to 60 m on both coasts and good De Geer moraines (generally conceded to indicate ice retreat in standing water) occur up to 90 m in the interior. A prominent beach is incised in bedrock at about 120 m above sea level on the White Hills west of St. Anthony. While the area exhibits many subhorizontal structure planes that also outcrop as benches, the horizontal extent and other corroborative evidence suggests that this feature is probably a marine erosion surface. Finally, the interlobate moraine extending west from South Summit of Highlands of St. John is wave-modified below 130 m (Cumming and Grant, 1974). This variation in the marine limit is partly due to the difficulty of recognizing the criteria in the wooded terrain, but it is mainly the result of the deglacial pattern. There is no evidence that the sea flooded in after the retreat of the Ten Mile Lake ice lobe, and it seems that sea level had dropped 15 to 30 m prior to the advance. Moreover, the time interval extrapolated from the change of level and the probable rate of uplift (ca. 7 to 15 m per century) agrees with the evidence of nearly synchronous retreating lowland ice and re-advancing highland ice. At present, sea level on the west coast of Newfoundland is considered to be relatively stationary.

In the adjacent offshore parts of the northern Gulf of St. Lawrence and Strait of Belle Isle, the sea floor is characterized by a succession of bedrock controlled cuestas, the latter bordered by prominent escarpments facing northwest (Haworth and Sanford, 1976). The strata forming the crests of the escarpments are generally exposed on the sea floor, whereas the elongated depressions between are filled with marine sediments or glacial tills or both. For the most part, surficial sediment cover of glacial origin in this part of the offshore is relatively thin and patchy, a fact largely attributable to the fast moving tides and currents and the numerous icebergs that pass from east to west through the Strait. Both phenomena have undoubtedly been important factors in the process of scouring of sediments from the relatively shallow sea floor of the Strait of Belle Isle and their transport to a somewhat deeper and more quiescent environment of deposition in Esquiman Channel farther south.

STRUCTURAL GEOLOGY

Lower Paleozoic platformal rocks contained within the northern part of the Anticosti Basin in the Strait of Belle Isle area were subjected to varying degrees of deformation during three principal tectonic episodes resulting from (i) emplacement of Hare Bay Allochthon in an early phase of the Taconic orogeny; (ii) uplift of the Precambrian basement of the Great Northern Highlands sometime between Middle Ordovician and Middle Devonian, followed by folding and thrusting of the eastern margin of the basin; and (iii) gentle upwarping of the southeastern margin of the Canadian Shield (Grenville Orogen) in late Paleozoic and Mesozoic time.

Presently preserved remnants of lower Paleozoic platformal strata in the Strait of Belle Isle area form part of the much larger structural depression in the Precambrian basement which occupies the entire northern Gulf of St. Lawrence and adjacent onshore areas of western Newfoundland and is referred to as the Anticosti Basin (see Wade et al., 1977). The northwestern limit of the basin extending from the Sept-Îles area in Quebec to Chateau Bay in southeastern Labrador is a reasonably well defined boundary between relatively flat-lying Paleozoic rocks and the highly deformed metamorphic terrane of the Precambrian Grenville Province. Much of the southern and southeastern margins of the basin, on the other hand, are structurally complex and ill-defined. This is particularly true of the northeastern part of the Strait of Belle Isle area between Canada Bay and Pistolet Bay where the eastern margin of the sequence is overlain by the Hare Bay Allochthon, all of which are in turn structurally overprinted by later Paleozoic deformational effects. Elsewhere in the Strait of Belle Isle area, Paleozoic platformal rocks are sharply bounded on the east by Precambrian rocks of the Great Northern Highlands which rise abruptly along a fault-line scarp to lie in juxtaposition to the coastal lowlands and fault-elevated Highlands of St. John.

The platformal rocks of the Strait of Belle Isle area (Fig. 79) dip gently southeast at an average of two degrees, but this regional inclination is locally interrupted by reversals of dip, due to faulting or folding or both. In the western part of the area normal faults are common and the strata may be gently to steeply inclined near major faults, as for example along the north shore of Ten Mile Lake where the beds are vertical (Fig. 74). On the other hand, large fault block systems have elevated above adjacent lowland terrain (i.e. Highlands of St. John) but still retain their relatively flat-lying attitude.

Deformation of platformal strata progressively increases eastwards across the area (Fig. 79) where they (Goose Tickle Formation) eventually display isoclinal folds and cleavage near the western margin of the Hare Bay Allochthon. The most intensely deformed strata occurs in the southeast corner of the map area where the St. George Group and the Table Head Formation were tightly folded probably during the Acadian orogeny. Near Englee, the rocks are recrystallized and flow-folded, but these characteristics are confined to a limited area and dips in the order of 20° are the normal structural style several kilometres to the west.

All lower Paleozoic platformal strata of the Strait of Belle Isle area contain prominent joint systems that are widely spaced in more massive beds (Fig. 80), and more

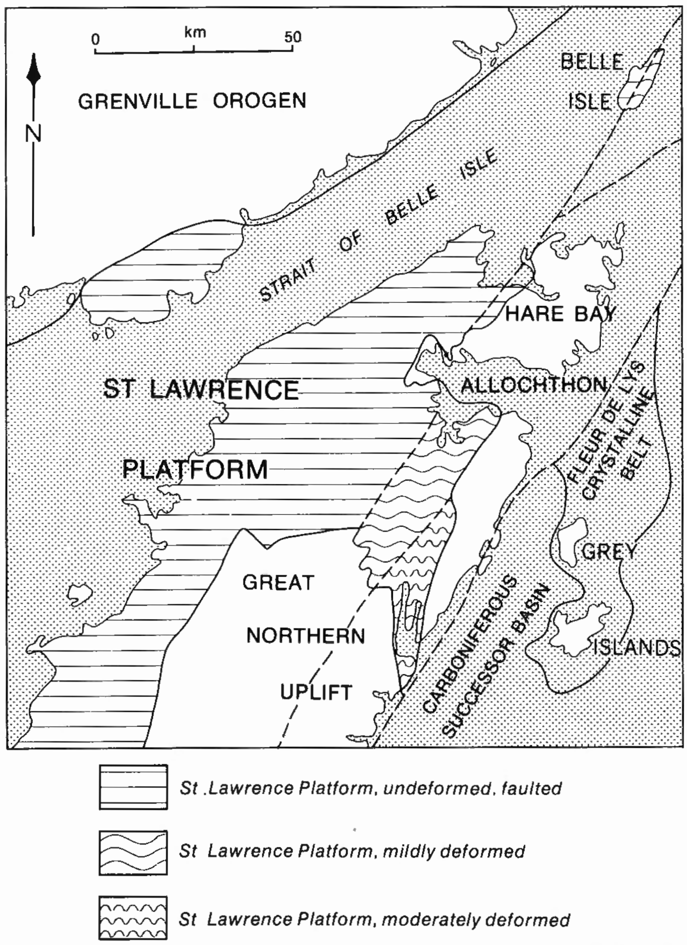


Figure 79. Major tectonic elements of the Strait of Belle Isle area.

Figure 80. Joint system in flat-lying, mud-cracked St. George dolomite on the west side of Pistolet Bay. (GSC 201500-O).

closely spaced in thinner beds. The predominant joint sets trend northeast and northwest as illustrated by rosette diagrams (Fig. 81), and this pattern is consistent throughout the area. The two joint sets belong to one principal system as indicated by the maintenance of a constant angle of 90° to 100° between the sets.

Joint planes, which are within 1 m of minor northeast-trending faults show slickensides that are well developed on the joint planes. More than 1 m from a fault, the joint planes show no evidence of small scale movement. It is therefore inferred that such regional joint systems predated the faulting and that the direction of the fault plane was controlled by a pre-existing joint system. These joint systems appear to have formed as a result of the release of residual stress following regional uplift of this part of the Anticosti Basin.

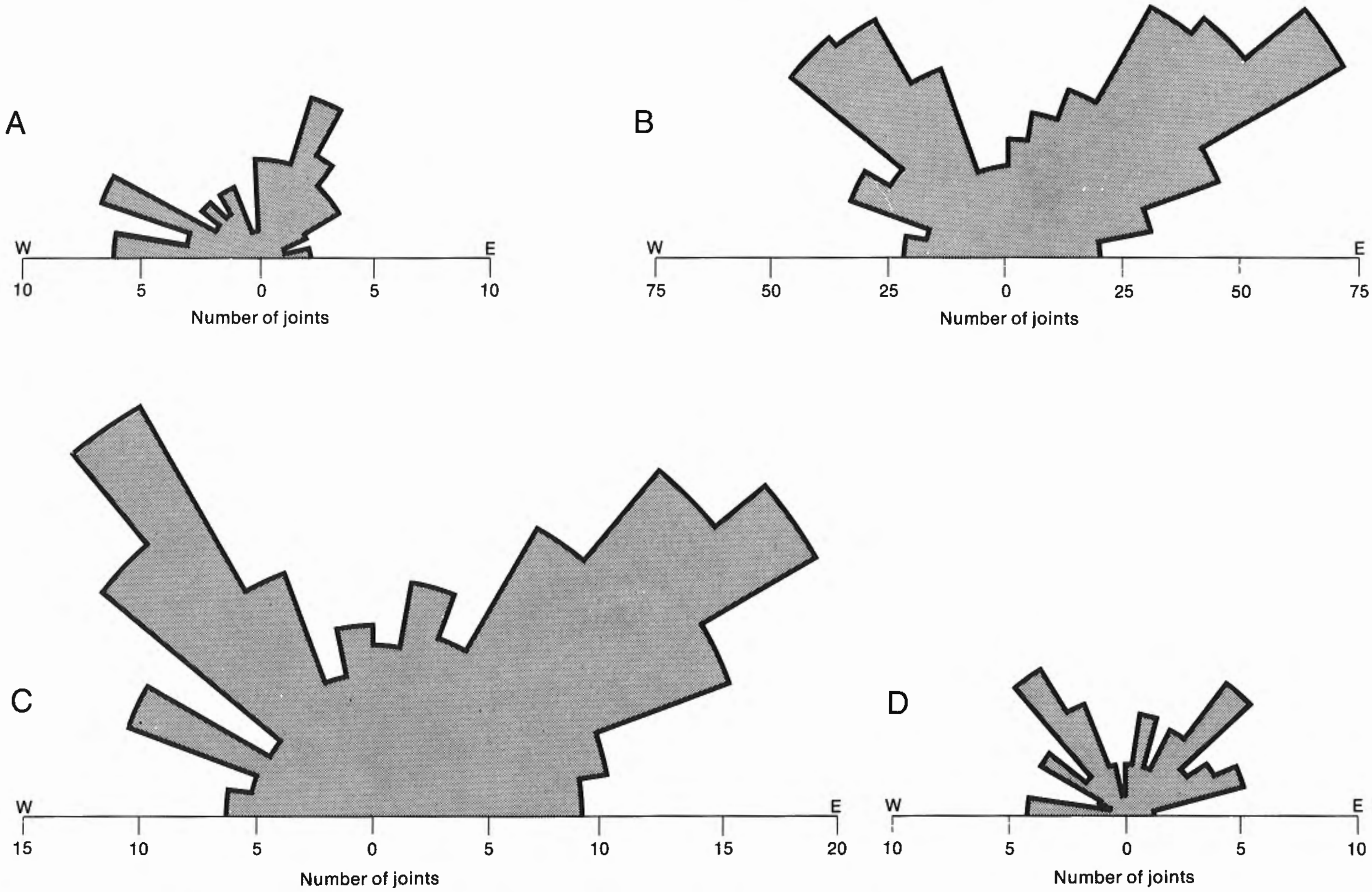
The lower Paleozoic platformal rocks are intersected by a considerable number of northeast-trending fault systems. Most are normal faults which appear to be related to the Great Northern Uplift and gentle upwarping of the southeastern margin of the Canadian Shield (Grenville Province).

Perhaps the most prominent and extensive are two relatively parallel fault systems, some 130 km long, which parallel the Great Northern Highlands and Highlands of St. John respectively and continue into the adjacent coastal lowlands to the northeast. The eastern fault marks the prominent fault-scarp boundary between areas containing the Highlands of St. John and Precambrian terrane of the Great Northern Highlands. Some 10 km to the west is another prominent fault scarp which forms the western boundary of the Highlands of St. John with the narrow coastal lowlands farther west. These faults display a composite vertical displacement of some 1300 m in the southern part of the Strait of Belle Isle area, but appear to decrease considerably in displacement as they trend across the coastal lowlands to the northeast.

Associated with the above faults are a considerable number of lesser subsidiary faults which are en echelon or splay out from the main systems. Either of such faults would apparently account for the local presence of a narrow inlier of Precambrian rocks near Ten Mile Lake. The evidence of the amount of vertical displacement at this locality is lacking although the presence of Lower Cambrian Forteau Formation in juxtaposition with Precambrian rocks would suggest a vertical displacement in the order of 610 m.

Southwest-trending subsidiary faults intersect rocks of the St. George Group and Table Head Formation near Port au Choix. Of principal interest are two systems, one of which extends through Back Arm, and the other lying offshore immediately to the northwest of the twin peninsulas at Port au Choix. Increasing dolomitization in the vicinity of the fault planes, extensive shattering of the strata and steeply plunging slickensides are the principal evidence for assuming the presence of two faults.

In the same general area as described above are two other faults of strike-slip nature. One is located near Port au Choix and the other at Barbace Cove, and both trend northwesterly. They have a left-lateral movement which would appear to postdate the northeast-trending fault at Back Arm. Dolomitized bands about 1.5 m wide are vertical and displaced left-laterally. Shallow-plunging to horizontal slickensides near the fault plane suggests movement of essentially strike-slip nature. The actual displacement is 2 to 2.5 m. Recent investigations in the northern Gulf of St. Lawrence by Haworth and Sanford (1976) demonstrated the seaward extension of one or more of the above described faults in the adjacent offshore areas to the south and west of Pointe Riche Penninsula. Some 95 km northeast of Port au Choix, a minor fault with a variable northeast to east trend was observed along the south shore of the Strait of Belle Isle, 5 km north of Green Island Brook. On the north side of the fault the beds are flat and relatively undisturbed, whereas on



A. Joint directions representing 61 measurements from the Bradore and Forteau formations in southern Labrador.

B. Joint directions representing a summary of 681 measurements in Cambro - Ordovician carbonate and sandstone strata of selected areas in St. Lawrence Platform, 51 undeformed and faulted.

C. Joint directions representing 204 measurments near Flower's Cove and Brig Bay.

D. Joint directions representing 59 measurements in the area between Main Brook and Round Lake.

Figure 81. Rosette diagrams of joint directions in the Strait of Belle Isle area.

the south the beds are uplifted and tilted 10°. A short distance to the southwest, on the largest of the Seal Islands, near Flower's Cove, a minor fault was observed where Eddies Cove Formation is highly brecciated with the matrix containing considerable pyrite (Fig. 82).

Numerous faults have been mapped in the Hawke Bay-Port Saunders area along the southern margin of the Highlands of St. John where the latter have been uplifted above the coastal lowlands. These faults trend in several directions, but most trend northeast. They, in turn, are intersected by faults trending north to northeast. The area is thus dissected into a number of fault blocks. Along the fault planes, slightly tilted but otherwise undeformed strata locally are tightly folded, essentially as a result of drag along the fault planes.

Along the shoreline between Cape Norman and Pistolet Bay, two northeast-trending faults are exposed. The strata adjacent to the fault planes are highly shattered, broadly folded, and locally tightly folded. Steeply to vertically plunging slickensides are well developed along adjacent joint planes filled with carbonate veins and suggest vertical movements. Many veins near minor faults with an average trend of 115° in turn show lateral movement both sinistral and dextral. Locally along the same shoreline, horizontal slickensides suggest that some blocks have also undergone strike slip movements, whereas most faults are of the normal variety.

Of major structural significance is an arcuate fault which extends northeasterly from the Great Northern Highlands near Cloud Hills across the coastal lowlands to Hare Bay. The actual location and displacement of the fault are unknown, but its presence is implied from distribution of the formations and its emplacement appears to be at least 600 m vertically with the east side down in the Salmon River area. There are undoubtedly a great number of additional northeast-trending faults in the lowland area north of the Great Northern Highlands that have not been identified either in the field or from air photographs. Many faults in the lowlands are implied from lineaments on air photographs.

Gentle upwarping of the southern margin of the Canadian Shield in Quebec and Labrador in the late Paleozoic and later times resulted in local faulting of the Lower Paleozoic strata at several localities. The more prominent is a northeast-trending fault that extends for some 32 km from Bradore Bay to the longitude of L'Anse-au-Loup. At the west end near Bradore Bay, the fault-line scarp is 150 m high. The Cambrian strata have been down-dropped to the south, and the displacement diminished toward the east. So that at its east end its effects are only minor cataclastic deformation along a network of veinlets 2 to 15 cm apart in Bradore arkose and sandstone with no vertical displacement apparent. Several other vertical and steeply dipping normal faults, with vertical displacements of 15 m or less, cut the Bradore and Forteau formations in various other localities in

Figure 82. View to the northwest from the largest of Seal Islands near Flower's Cove; the fault breccia in the foreground at 51°17'23"N; 56°45'38"W, is composed of angular fragments of white-weathering dolomite set in a pyritic matrix. This breccia marks a minor, northeast-trending, normal fault which cuts the flat-lying dolomite of the Eddies Cove Formation. (GSC 6-3-71 LMC).

Labrador. Three sets of such normal faults occur as follows: one trending at 20°; another at 55°; and a third at 80°. In addition, two parallel, closely spaced faults trend northerly from L'Anse-au-Clair. They are expressed topographically as a narrow bedrock canyon incised in the Paleozoic terrane.

In the Strait of Belle Isle area east of a line from Belle Isle to the vicinity of Cloud Hills, deformation in the form of normal and thrust faulting and folding becomes more pronounced and is undoubtedly the result of compression along the eastern margin of the platform sometime between Middle Ordovician and Middle Devonian. Strata of the Labrador Group on Belle Isle are considerably more intensely faulted and folded than their counterparts in adjacent areas of insular Newfoundland and Labrador. The unconformity between the Lower Paleozoic and/or Hadrynian Bateau Formation and Precambrian basement rocks along the northeast coast of Belle Isle, has been tilted steeply to the east. In addition, the Lower Paleozoic strata have been intersected by high angle normal faults and locally by small scale thrust faults. These strata are locally overturned towards the west by high angle, westerly directed thrust faults.

At the southwest end of Belle Isle a small overturned syncline contains shale of the Forteau Formation and nearby Precambrian basement gneiss have been thrust westward over sandstone of the Bradore Formation. Northeast- trending normal faults can be observed at Scotswood Cove, Lighthouse Cove, and Blandfords Cove extending west to Round Head and White Point. The basal contact of the Bradore Formation is offset by faults at Lighthouse and Scotswood coves, where the faults have a right-lateral, strike-slip component. Other faults are either normal or westerly directed, high-angled thrust faults.

The deformation of platformal rocks as a result of emplacement of the Hare Bay Allochthon during an early phase of the Taconic orogeny in the early Middle Ordovician time is displayed in the form of local folding of the Goose Tickle Formation over which the allochthon was transported, in addition to extensive faulting which undoubtedly accompanied the transport of the thrust slices. Further, more intensive thrust faulting along the easternmost margin of the map area (Pistolet Bay to Canada Bay) took place between Middle Ordovician and Middle Devonian time where allochthonous terrane along the western margin of White Bay was further deformed and compressed against the platform. The local and regional structural geology of this more eastern part of the Strait of Belle Isle area can be found in Williams and Smyth (this memoir).

ECONOMIC GEOLOGY

Introduction

A number of small quarry operations in the Strait of Belle Isle area have provided materials for road construction using shale of the Forteau Formation and limestone rubble of the Table Head Formation. In addition, two small quarries at Canada Harbour have been opened in white marble deposits of the Table Head Formation. No other mining or quarrying operations occur in the area, although future potential exists for development of the extensive limestone deposits and for the occurrence of Mississippi Valley-type lead-zinc deposits. The offshore region of the Gulf of St. Lawrence is a possible future source of oil and natural gas.

Limestone and dolomite

High quality limestone deposits occur in abundance in many parts of the area. The following localities offer the best potential for development, their descriptions are largely summarized from DeGrace (1974).

1. St. John Island. Extensive carbonate deposits occur on St. John Island in St. John Bay, 13 km north of Port au Choix. The island, with an area of 10 km² rises to an elevation of about 60 m above sea level. In this locality, reserves of high-calcium limestone of the Table Head Formation are in the order of 690 megatonnes plus 1.2 megatonnes of dolomite, and at least 0.9 megatonnes of dolomitic limestone both from the St. George Group (DeGrace, 1974).

The western half of St. John Island is entirely underlain by brownish grey, medium- to thick-bedded, fossiliferous, fine to medium crystalline limestone of the Table Head Formation. White crystalline calcite is common as infillings of vugs and veins. Due to the irregular fracture pattern the rock, upon weathering, typically disintegrates to a limestone rubble. Black bituminous material occurs frequently along fracture systems, locally with minor argillaceous material. The bedding attitude varies from nearly flat-lying to gently dipping. The limestone is uniform in lithology throughout the section, with only slight variations in colour, hardness, and bituminous content.

To the southeast of St. John Harbour is a neck of land on which there is about 30 m of similar Table Head limestone with favourable conditions for quarrying. This area has a reserve potential of 54 megatonnes.

2. Cooks Harbour. Immediately adjacent to the wharf, just north of Cooks Harbour village, Table Head limestone is abundant and has an average composition of 53.1 per cent CaO, 1.1 per cent MgO, 2.4 per cent impurities (DeGrace, 1974). This limestone is medium to dark grey with a brown cast, micro- to finely crystalline, and weathers medium grey. The limestone resembles that of St. John Island, but is more deformed. The limestone also outcrops along the road from Cooks Harbour to and beyond the intersection with the road to St. Anthony.

3. Burnt Island. Burnt Island, located immediately west of the village of Raleigh (the latter containing wharf facilities) is underlain by relatively pure limestone and is nearly barren of vegetation. Although typical of the Table Head Formation, the limestone is brownish grey, hard, recrystallized, fractured, and contains abundant white calcite veins and stringers. The upper 15 m of limestone has an average composition of 53 per cent CaO, 1.0 per cent MgO, and 2.0 per cent impurities. The remainder contains 6 to 10 per cent MgO and is slightly higher in impurities. Total limestone reserves of the island are 363 megatonnes (Harris, 1962; McKillop, 1962).

4. Hare Island. On Hare Island, in northwestern Hare Bay, 46 m of Table Head limestone overlies 9.1 m of interbedded dolomite and limestone, and is in thrust fault contact with dark grey shale of the Northwest Arm Formation. A grab sample at this locality tested 54.1 per cent CaO, 0.6 per cent MgO, and 1.5 per cent impurities. High-calcium limestone reserves at this site are approximately 8.2 megatonnes (DeGrace, 1974).

5. Other Localities. Much of the region bordering Eddies Cove West and Port au Choix is underlain by relatively pure beds of Cambrian and Ordovician limestone that are well exposed in the coastal regions. Most of the limestone dips consistently and gently to the southeast.

Immediately northeast of Gargamelle and east of Back Arm is a prominent ledge underlain by at least 9 m of Table Head limestone, which is in turn underlain by dolomite. Quarrying of the weathered limestone has been carried out at this locality for local use as road material. Southeast of Gargamelle, another limestone bed is exposed, which is relatively free of glacial overburden. Immediately southwest of Gargamelle is another ledge of limestone on a neck of land that forms the southern side of Gargamelle Cove. In addition to the above, relatively pure St. George limestone is well exposed on the west shore north of Boat Harbour.

Zinc occurrences

Surface indications of mineralized carbonate rocks in the St. George Group to the west of Hare Bay are similar to mineralized showings to the south of the present map area, near Daniels Harbour, where ore reserves of at least 4.9 megatonnes averaging 7.7 per cent zinc are now being mined (Collins and Smith, 1975).

The sphalerite mineralization in the Zinc Lake area, 11 km northeast of Daniels Harbour, comprises a number of vug-filling and intercrystalline orebodies in a mottled dolomite host rock which lies stratigraphically in the uppermost beds of the St. George Group. The ore is of the stratabound Mississippi Valley-type and is genetically related to a pre-Middle Ordovician erosion surface (Cumming, 1968). Further information pertaining to the detailed stratigraphy and habitat of the ore is contained in Collins and Smith (1972, 1973, 1975).

One minor constituent that occurs commonly in western Newfoundland zinc deposits is cadmium. It substitutes for zinc in the sphalerite lattice, and its abundance as a solid solution is directly dependent upon the zinc content of the ores (Sangster, 1968). The low-iron sphalerite of the Newfoundland deposits commonly contains a relatively high proportion of cadmium and provides an economically important by-product obtained during the smelting of the zinc ores.

Zinc and cadmium concentrations are mainly restricted to a particular diagenetic facies of a limestone unit near the top of the St. George Group and consists of honey-coloured sphalerite surrounding brown dolomite in a pure white dolomite gangue. This host rock, referred to pseudobreccia (Cumming, 1968), is laterally persistent but is mineralized only locally where it is adjacent to areas of solution collapse breccia. The solution collapse breccias are laterally discontinuous features developed mainly in the Catoche Formation (Kluyver, 1975). These zones vary in size but average 900 m in length and 90 m in width. They are composed of clasts and fragments of limestone, dolomite, chert, organic material, and iron oxides or iron sulphides.

The collapse breccias consist of angular to subrounded fragments and clasts, up to 1 m or more in diameter which are set in a matrix of fine-grained dolomite, and iron oxides or sulphides. Chert pebbles are common in the matrix and pisolitic aggregates in a dolomite-spar matrix occur near the base of some of the breccias. The matrix is usually darker than most fragments, reflecting its residual nature with much higher iron oxide, sulphide, and hydrocarbon content.

The collapse breccias are interpreted as paleokarst deposits, filling caves and underground channels produced during the time of erosion between the St. George and the Table Head depositional periods (Collins and Smith, 1972, 1975). These paleokarst deposits indicate a minimum uplift of 100 to 130 m above sea level after the deposition of the Port au Choix Formation.

The pseudobrecciation associated with the sphalerite deposits results from selective dolomitization of the limestone beds of the Catoche Formation, followed by leaching of limestone during karst development, and then by open-space filling by white sparry dolomite, probably during subsidence preceding deposition of the basal Table Head limestone. The final product consists of patches of brown dolomite surrounded by coarse crystalline white dolomite.

Adjacent to collapse breccia masses (i.e. areas of greatest porosity or micrite dissolution), zinc sulphide was deposited before white dolomite and grew in colloform fashion around dolomite patches. Where not recrystallized, the sphalerite has maintained its colloform habit. Sphalerite is everywhere coated by white dolomite, which filled the remaining open spaces (Collins and Smith, 1973).

During uplift at the end of Early Ordovician, karstification of the carbonate platform developed major underground or cavern drainage systems within the Catoche limestone, with attendant channelways and minor openings along faults and joints in this and overlying strata (Collins and Smith, 1972). More selective solution of limestone beds immediately adjacent to caverns developed a porous anastomosing network of dolomite mottles and dolomite interbeds. Within this underground hydrologic system, colloform sphalerite grew around the remaining dolomite patches (Collins and Smith, 1973).

Beales et al. (1974) showed from paleomagnetic studies of the Daniels Harbour orebodies that the pole positions were similar for both the St. George Group host rocks and the ore itself. They concluded that the ore and the host rocks are of similar age.

Mineral deposits in Ordovician carbonate in western Newfoundland are similar to those in eastern Tennessee (S.W. Mather, Tennessee Division of Geology, pers. comm., 1974). There the Knox Group (St. George equivalent) and Chickamauga (Table Head equivalent) Group are separated by a major erosional interval. The Knox Group is the host rock for Pb-Zn ore deposits and its upper surface is a paleokarst with local relief as great as 60 m or more. Some of the sinkholes related to this surface in Tennessee have fillings of tuffaceous matter, in others the filling is quartzite pebbles and sand, and in others the filling is carbonate breccia blocks in carbonate matrix. In enough instances to make coincidence an unlikely choice, sphalerite, or barite, or pyrite occur stratigraphically below major sinks of Early - Middle Ordovician age. The orebodies are localized in carbonate breccias.

The morphology and internal features of the west coast Newfoundland zinc deposits are also similar to those of eastern Tennessee. That is, zinc deposits in both areas are localized in Lower Ordovician carbonate rocks which occur below a regional disconformity. The collapse breccias in the ore horizon are similar in both areas. The internal structure of the zinc ore in both areas is strikingly similar. The deposits are nearly monomineralic, with a low-iron sphalerite being the main ore mineral.

Such zinc deposits appear to have wide lateral extent in western Newfoundland. The known deposits near Daniels Harbour correspond to deposit type B-1 and B-2 of Callahan (1974). Additional favourable stratigraphic horizons may occur within and at the base of the autochthonous sequence (Fig. 83).

Oil and gas possibilities

No oil or gas shows have been discovered in the Strait of Belle Isle area. However, the region has a marginal potential in the nearby offshore areas of the Anticosti Basin.

Few indications of hydrocarbons were noted during the present surveys although, in general, the porous dolomite of the St. George Group gives off a petroliferous aroma when freshly broken. Also, tarry bitumen was noted in vugs in the dolomite of the upper 1.5 m of the St. George Group on Port au Choix Peninsula.

Johnson (1941a) and Fleming (1970) summarized the numerous occurrences of hydrocarbons reported from western Newfoundland including an occurrence of oil shale in the Carboniferous succession near Conche in White Bay.

Data from wells drilled on Anticosti Island indicate oil shows in the Ordovician section there. In particular, these oil shows have been largely from the Romaine Formation, which is a correlative of the St. George Group.

Geology of proposed tunnel crossings

One of the principal objectives of the writer's field work in the Strait of Belle Isle area was to establish the stratigraphic and tectonic framework of the lower Paleozoic sequences of northern Newfoundland and Labrador. Such information, which could be extrapolated beneath the sea floor of the Strait of Belle Isle, is useful to engineering studies of a possible tunnel for cables to carry hydroelectric power from Labrador to the Island of Newfoundland.

In 1973, medium penetration reflection seismic surveys were carried out under the direction of D.E.T. Bidgood of the Nova Scotia Research Foundation in the Strait of Belle Isle, and this was followed up by the drilling of core holes at Pointe Amour (Labrador) and Yankee Point (Newfoundland).

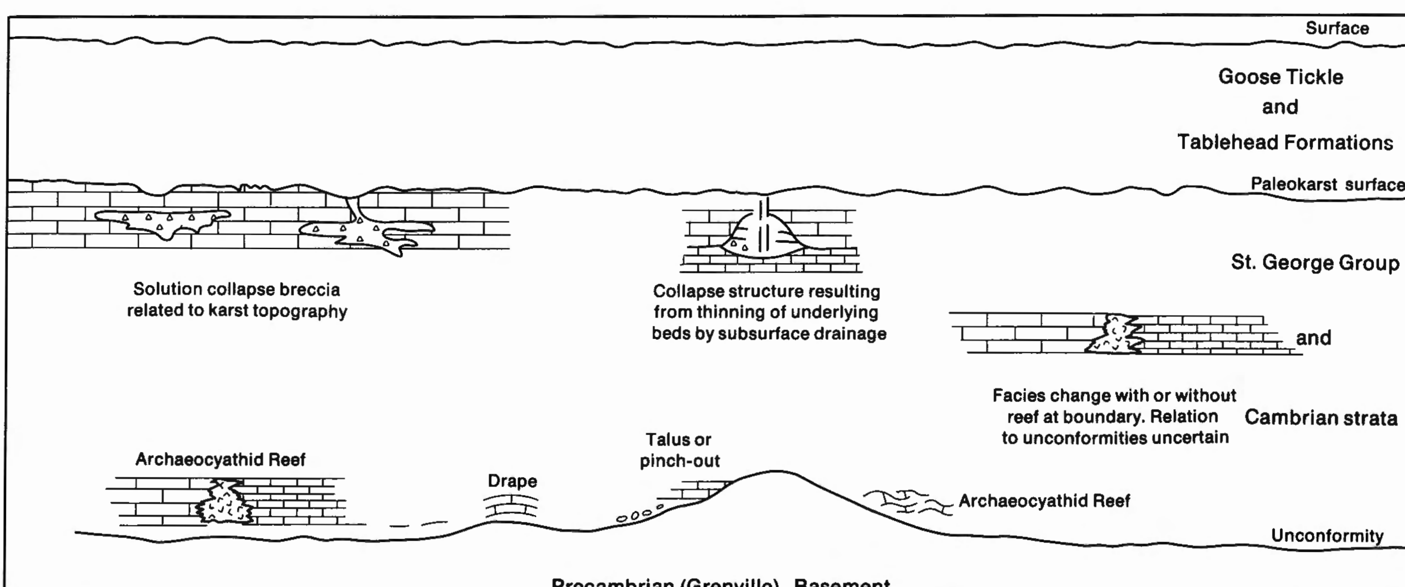


Figure 83. Idealized vertical section through the platformal strata of western Newfoundland showing features along which Mississippi Valley-type deposits may be formed (modified from Callahan, 1974).

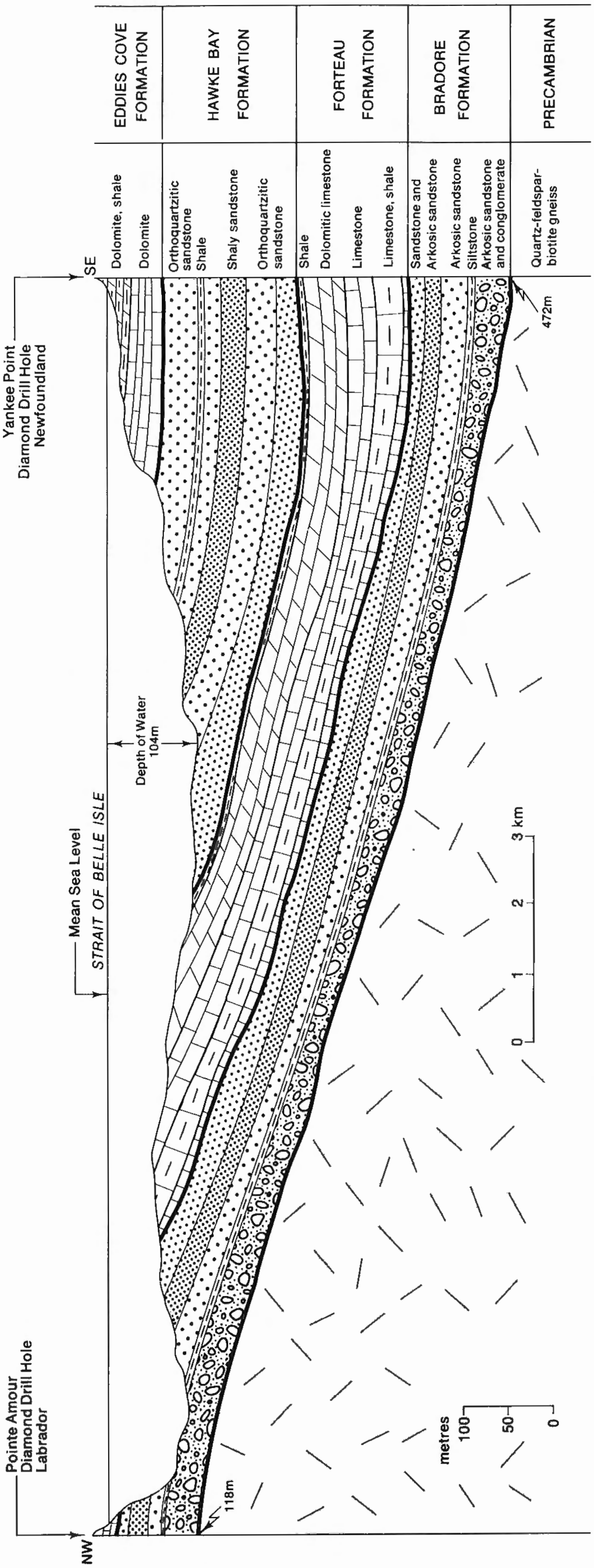


Figure 84. Geological cross section of the narrowest part of the Strait of Belle Isle.

The drilling was carried out by Teshmont Consultants Ltd., Winnipeg, and Patrick Harrison and Company, Ltd., Toronto. The geological logs of the core from these holes appear in the appendix to this report.

In November 1975 additional deep penetration reflection seismic surveys, in the narrow part of the Strait, were performed by Geoterrex, for Harrison Bradford and Associates Limited, St. Catharines.

A plan view showing the distribution of Lower Cambrian formations in the area of the proposed crossing beneath the western approaches of the Strait of Belle Isle is illustrated by Haworth and Sanford (1976). Much of the sea floor beneath the Strait is underlain by carbonate and shale of the Forteau Formation. Immediately offshore from Labrador the bedrock surface is composed of arkosic and orthoquartzitic sandstone and conglomerate of the Bradore Formation, whereas immediately offshore of insular Newfoundland, younger Hawke Bay orthoquartzitic sandstone and Eddies Cove dolomite form the sea floor terrain in a narrow belt bordering the coastline.

At Pointe Amour in Labrador, 118 m of Forteau and Bradore strata were penetrated by drilling, whereas on the coast of Newfoundland 472 m of Paleozoic strata were encountered, the youngest being the Eddies Cove Formation underlain in turn by strata of the Hawke Bay, Forteau, and Bradore formations.

On the basis of core hole data and seismic reflection surveys, the subsurface mapping of a corridor between Pointe Amour and Yankee Point has provided considerable new information pertaining to the stratigraphy and structure of the Paleozoic rocks that can be expected to be encountered in the construction of a tunnel between Labrador and insular Newfoundland (Nova Scotia Research Foundation, 1973).

Figure 84 is a generalized cross-section between Pointe Amour Labrador and Yankee Point, insular Newfoundland, showing the succession and lithology of the rock units comprising the Paleozoic strata beneath the Strait of Belle Isle and adjacent onshore areas. The cross-section does not show faults, which undoubtedly cut the strata in the Strait.

APPENDIX

GEOLOGICAL LOGS OF CORES FROM DIAMOND DRILL HOLES ON STRAIT OF BELLE ISLE

1. *Yankee Point hole 74-B-D1*

Location: Near south shore of Strait of Belle Isle at Yankee Point, insular Newfoundland 51°19'35"N, 56°43'08"W.

Elevation: 21.3 feet. Inclination: vertical.

Total depth: 1550 feet. Completed: 1975.

Purpose: To determine rock character, stratigraphy and depth to basement along proposed tunnel beneath Strait of Belle Isle for cables to carry hydroelectric power from Labrador to insular Newfoundland. Drilling for Teshmont Consultants Ltd., Winnipeg, and Patrick Harrison and Company, Ltd., Toronto.

Log by P.T.N. Hum, N. James, and L.M. Cumming from core stored at Donovan's Industrial Park, St. John's, Newfoundland.

Middle and (?) Upper Cambrian Eddies Cove Formation	(11 - 239 ft.)	228 ft. (incomplete) (69.5 m)
Middle Cambrian Hawke Bay Formation	(239 - 749 ft.)	510 ft. (155.5 m)
Lower Cambrian Forteau Formation	(749 - 1148 ft.)	399 ft. (121.7 m)
Lower Cambrian and (?) Hadrynian Bradore Formation	(1148 - 1547 ft.)	399 ft. (121.7 m)
Helikian and older granitic gneiss	(1547 ft.)	

Depth (feet)	Description	Thickness (feet)
0 - 2	Gravel and boulders	2
2 - 11	Sand and gravel	9
	EDDIES COVE FORMATION (228 feet, incomplete)	
11 - 15	Dolomite and shale, grey to dark grey; 80 per cent dolomite 20 per cent shale; shale beds vary in thickness from micro to 0.5 foot.	4
15 - 38	Dolomite, grey to light grey; occasional small vugs infilled with dolomite and calcite crystals; interbedded shale varying in thickness from micro to 1.2 feet with average spacing of 0.3 foot; 90 per cent dolomite, 10 per cent shale.	23
38 - 59	Dolomite, grey to light grey; occasional small vugs infilled with calcite and dolomite crystals; interbedded dark grey shale varying from micro to 1.0 foot thick and average spacing of 0.5 foot; 90 per cent dolomite, 10 per cent shale.	21
59 - 71	Dolomite and shale, light to dark grey; shale varies in thickness from micro to 0.4 foot and spaced at 0.3 foot; 50 per cent dolomite, 50 per cent shale.	12
71 - 79	Shale; beds vary in thickness from micro to 2.1 feet, with average spacing of 0.4 foot; 50 per cent dolomite,50 per cent shale.	8
79 - 87.5	Dolomite light grey; thin interbedded shale partings varying in thickness from micro to 0.01 foot and spaced at 0.4 foot; 90 per cent dolomite, 10 per cent shale.	8.5
87.5 - 90.7	Dolomite and shale, light grey to grey; 50 per cent dolomite, 50 per cent shale.	3.2
90.7 - 98.5	Dolomite, light grey; shale partings of micro thickness spaced at 0.6 foot.	7.8
98.5 - 98.9	Dolomite and shale interbedded; shale beds vary in thickness from micro to 0.2 foot; 75 per cent dolomite, 25 per cent shale.	0.4
98.9 - 117.7	Dolomite, light grey to grey; interbedded shale, micro to 0.2 foot thick with average spacing of 0.2 foot; 85 per cent dolomite, 15 per cent shale.	18.8
117.7 - 139	Shale and dolomite, dark grey to grey; shale beds have an average thickness of 0.5 foot; 80 per cent shale and 20 per cent dolomite.	21.3
139 - 239	Dolomite, grey, fine to medium grained with shale partings, grey to dark grey, mostly of hairline thickness but occasionally 0.1 foot. These partings have a subhorizontal trend and are irregular with an average spacing of 0.05 foot; 12 joints infilled with fibrous gypsum 0.01 foot thick and occasional vugs filled with dolomite and gypsum. Joint spacing 2.0 feet; 98 per cent dolomite, 2 per cent shale.	100
	HAWKE BAY FORMATION (510 feet)	
239 - 379	Sandstone, orthoquartzitic, white to grey and interbedded shale black to dark grey. Bed thickness varies from micro to 1.3 feet, average 0.2 foot. Orthoquartzite has thin laminae of shale of micro thickness and load casts and crossbedding; occasional fibrous gypsum veinlets 0.01 foot thick; 60 per cent sandstone and orthoquartzite, 40 per cent shale.	140
379 - 411.5	Shale, dark grey to black, occasional orthoquartzite beds from 387.0 to 399.0; 90 per cent shale, 10 per cent orthoquartzite.	32.5
411.5 - 419	Orthoquartzite and shale partings; 95 per cent orthoquartzite, 5 per cent shale.	7.5
419 - 421.7	Sandstone, orthoquartzitic, and shale; 50 per cent orthoquartzite, 50 per cent shale.	2.7
421.7 - 425	Sandstone, orthoquartzitic, white to grey. Shale partings micro to 0.03 foot in thickness spaced at 0.3 foot.	3.3
425 - 429	Shale, subarkosic, red.	4
429 - 449	Sandstone, orthoquartzitic, fine to medium grained white to pink to grey. Average bed thickness 1.0 foot and varies micro to 2.5 feet thick; interbedded shale 0.1 foot to micro in thickness; 95 per cent orthoquartzite, 5 per cent shale.	20
449 - 459	Shale with interbedded orthoquartzite, fine to medium grained, black to light grey. Bed thickness micro to 0.8 foot, average 0.06 foot; 70 per cent shale, 30 per cent orthoquartzite.	10

Depth (feet)	Description	Thickness (feet)
459 - 489	Sandstone, orthoquartzitic, white, occasional grey shale laminations of micro thickness, spaced at 2.0 feet average. There are two subvertical fractures 0.5 foot long and 0.1 foot wide filled with fibrous gypsum; 99.9 per cent orthoquartzite, 0.1 per cent shale and gypsum.	30
489 - 507.2	Sandstone, and shale interbedded average thickness 0.1 foot. Sandstone is grey to green to white; arkosic to orthoquartzitic. Bedding is subhorizontal; 75 per cent shale, 25 per cent sandstone.	18.2
507.2 - 509	Sandstone, orthoquartzite, micro shale laminations at 0.2 foot spacing.	1.8
509 - 519	Sandstone orthoquartizitic, white to grey to pinkish. Occasional shale lamination micro to 0.01 foot thick, spacing 0.2 foot; 90 per cent sandstone, 10 per cent shale.	10
519 - 529	Sandstone, orthoquartzitic, to subarkosic, pink to red shale partings micro to 0.1 foot thick, spaced at 0.8 foot; 80 per cent sandstone, 20 per cent shale.	10
529 - 539	Sandstone, subarkosic, red. Bedding is horizontal and displays slumps, load casts and sand volcanoes. Average bed thickness is 0.1 foot; 95 per cent sandstone, 5 per cent shale.	10
539 - 542.5	Sandstone, red and red shale interbedded.	3.5
542.5 - 559	Shale and sandstone, black, grey to dark green. There are many small scale sedimentary features as in 529 - 539. Bedding is horizontal; 65 per cent shale, 35 per cent sandstone.	16.5
559 - 560.8	Sandstone and shale, grey to black as above.	1.8
560.8 - 589	Sandstone, orthoquartzitic, white to pinkish, medium to fine grained. There are subvertical fractures 2 to 3 feet long. These are cemented with soft, white gypsum. The fractures are micro to 0.01 foot thick. There are subhorizontal shale laminae of micro thickness at an average spacing of 1 foot; 99.9 per cent orthoquartzite, 0.1 per cent shale.	28.2
589 - 592.7	Sandstone, orthoquartzitic, as above.	3.7
592.7 - 629	Shale, red to black to dark grey, and orthoquartzitic sandstone; bed thickness varies from micro to 2.0 feet; 50 per cent shale, 50 per cent sandstone.	36.3
629 - 650	Sandstone, orthoquartzitic, white to grey with occasional bands of dark grey shale. Shale bands vary in thickness from micro to 0.5; 90 per cent orthoquartzite, 10 per cent shale.	21
650 - 668	Sandstone, orthoquartzitic, white to grey, in beds averaging 0.6 foot thick.	18
668 - 718.5	Sandstone, orthoquartzitic, and interbedded shale, sandstone beds are up to 3.5 feet thick and shale beds average 0.2 foot thick; 50 per cent sandstone, 50 per cent shale.	50.5
718.5 - 736	Fault breccia composed of 30 per cent sandstone fragments and 70 per cent shale fragments; also fault gouge in two zones each with a width of 1 foot.	17.5
736 - 749	Sandstone, orthoquartzitic interbedded with shale; sandstone beds up to 2 feet thick; 75 per cent sandstone, 25 per cent shale.	13

FORTEAU FORMATION
(399 feet)

Depth (feet)	Description	Thickness (feet)
749 - 772	Shale, dark grey, making up greater than 95 per cent of the section; cut by a network of veinlets and stringers of white calcite up to 0.07 foot thick.	23
772 - 802	Dolomite with shaly partings; contains carbonate clasts up to 0.5 inch in width; 80 per cent dolomite, 20 per cent shale.	30
802 - 810	Shale interbedded with limestone; 70 per cent shale, 30 per cent limestone.	8
810 - 818	Dolomite with carbonate veinlets up to 0.2 inch thick; contains vugs which make up 30 per cent of the section. Artesian salt water encountered at this horizon.	8
818 - 832	Dolomite, massive with shale interbeds and partings; 95 per cent dolomite, 5 per cent shaly material.	14
832 - 840	Shale and interbedded limestone containing limestone clasts; 70 per cent shale, 30 per cent limestone.	8
840 - 859	Limestone and shale interbedded; basal 5 feet contains dolomite clasts; 80 per cent limestone, 20 per cent shale.	19
859 - 882	Limestone, dolomitic and fossiliferous, interbedded with shale; limestone 80 per cent, shale 20 per cent.	23
882 - 889	Limestone and shale interbedded in equal amounts.	7
889 - 894	Limestone and shale interbedded; 80 per cent limestone.	5
894 - 905	Limestone, oolitic with calcite veinlets up to 0.2 inch thick.	11
905 - 912	Limestone and shale interbedded; limestone makes up 70 per cent of the section and contains fossils.	7
912 - 919	Limestone, coarsely crystalline, massive.	7
919 - 928	Limestone and shale interbedded; 65 per cent limestone.	9
928 - 933	Shale and interbedded limestone; 70 per cent shale.	5
933 - 990	Limestone and shale interbedded; limestone is oolitic and fossiliferous; shale is friable and contains occasional clasts of limestone up to 1 inch wide; 80 per cent limestone, 20 per cent shale.	57
990 - 1005	Limestone, fossiliferous with minor shale partings.	15
1005 - 1030	Limestone and interbedded shale; massive limestone 50 per cent, fossiliferous limestone 10 per cent, shale 40 per cent.	25
1030 - 1115	Shale, laminated, contains rounded clasts of carbonate and gypsum up to 0.2 inch in diameter; interbedded limestone; 80 per cent shale, 20 per cent limestone.	85
1115 - 1132	Shale, laminated, and interbedded limestone; 60 per cent shale, 40 per cent limestone.	17
1132 - 1148	Limestone and shale interbedded; limestone is fossiliferous and shale beds average 0.2 foot thick and contain occasional rounded clasts of limestone; 55 per cent limestone, 45 per cent shale. These are the basal beds of the Forteau Formation.	16

BRADORE FORMATION
(399 feet)

Depth (feet)	Description	Thickness (feet)
1148 - 1189	Sandstone, red, arkosic, interbedded and interlaminated with red shale, contains scattered gypsum clasts.	41
1189 - 1196	Sandstone, red, arkosic, friable.	7
1196 - 1201	Sandstone red, arkosic massive, poorly cemented in places.	5
1201 - 1266	Sandstone, orthoquartzitic, grey-green, with subordinate amounts of arkosic sandstone; orthoquartzite contains 80 to 90 per cent white quartz grains in silica cement.	65
1266 - 1300	Sandstone, red-brown, arkosic, medium to coarse grained.	34
1300 - 1308	Sandstone, arkosic interbedded with shale in beds up to 0.2 foot thick. This was the end of the hole in 1974.	8
	In 1975 the hole was re-entered and deepened to basement. From 1308 feet downward, a uniform section of red arkosic sandstone of the Bradore Formation. At 1547 feet the arkose is in knife-sharp unconformable contact with granitic gneiss of the Precambrian basement. Total depth of hole 1550 feet.	239

2. *Pointe Amour hole 73-B-D2*

Location:	Near north shore of Strait of Belle Isle 1240 feet northeast of lighthouse at Pointe Amour, Labrador. 51°27'35"N; 56°51'25"W.
Elevation:	70 feet. Inclination: vertical.
Total Depth:	489 feet. Completed 1974.
Purpose:	To determine rock character, stratigraphy and depth to basement along proposed tunnel beneath Strait of Belle Isle for cables to carry hydroelectric power from Labrador to insular Newfoundland. Drilling for Teshmont Consultants Ltd., Winnipeg, and Patrick Harrison and Company Ltd., Toronto.

Log by P.T.N. Hum, N. James and L.M. Cumming from core stored at Donovan's Industrial Park, St. John's, Newfoundland.

Lower Cambrian Forteau Formation	(1 - 73.6 ft.)	72.6 ft. (incomplete) (22.1 m)
Lower Cambrian and (?) Hadrynian Bradore Formation	(73.6 - 387.5 ft.)	313.9 ft. (95.7 m)
Helikian and older granitic gneiss	(387.5 - 489 ft.)	101.5 ft. (incomplete) (31.0 m)

Depth (feet)	Description	Thickness (feet)
0 - 1	Moss and peat.	1
	FORTEAU FORMATION (72.6 feet, incomplete)	
1 - 13.5	Shale fractured, with mud partings 1 to 3 inches thick.	12.5
13.5 - 19.0	Limestone and shale, grey to dark grey, bedding horizontal; joints with 0.5 foot spacing are mud-coated.	5.5
19.0 - 49	Limestone and shale, black to grey; joints with 2 foot spacing; crystalline calcite pods at 22.0 to 22.6 feet and a crystalline calcite horizon 0.1 foot thick at 40.1 feet.	30
49 - 73.6	Limestone and shale, grey to red; occasional cavities lined with limonite-coated calcite crystals.	24.6
	BRADORE FORMATION (313.9 feet)	
73.6 - 79	Sandstone, arkosic, red; a few shale partings and calcareous blebs; bedding horizontal.	5.4
79 - 84.3	Sandstone, orthoquartzitic, pink, fine grained.	5.3
84.3 - 85.3	Siltstone and sandstone, micro fractures along bedding.	1
85.3 - 89	Sandstone, orthoquartzitic, medium grained.	3.7
89 - 91	Sandstone, orthoquartzitic to subarkosic, reddish to purple.	2
91 - 93.6	Sandstone, red, medium grained with scattered red siltstone rock fragments.	2.6
93.6 - 95.2	Sandstone, orthoquartzitic, reddish to purple.	1.6
95.2 - 97	Conglomerate, red, composed of rounded fragments 0.25 inch in diameter of quartz, feldspar and scattered metamorphic rock fragments.	1.8
97 - 99	Sandstone, subarkosic to orthoquartzitic, reddish to purple with liesegang bands.	2
99 - 109	Sandstone, orthoquartzitic to subarkosic, dark red to purple to pinkish, medium grained, partings of soft yellow clay mineral.	10
109 - 129	Sandstone, arkosic, red to dark red, medium grained.	20
129 - 139	Sandstone, arkosic to subarkosic, dark red to grey, coarse to medium grained. Joint spacing is at 3.0 feet varying from horizontal to 45° dip.	10
139 - 169	Sandstone, arkosic to orthoquartzitic, red to grey, medium grained.	30
169 - 189	Sandstone, arkosic to subarkosic, reddish to dark red. There are occasional rounded quartz and feldspar fragments 0.25 inch in diameter.	20
189 - 199	Sandstone, subarkosic, red, coarse to medium grained; predominance of rounded quartz and feldspar fragments, with siliceous cement.	10
199 - 209	Sandstone, subarkosic, red, coarse to medium grained, occasional bands of thinly laminated red and green siltstone.	10
209 - 219	Sandstone, subarkosic, medium to fine grained, occasional bands of thinly laminated siltstone, dark red to red.	10
219 - 226	Sandstone, subarkosic, reddish brown, medium grained, contains four bands of thinly laminated siltstone 0.05 to 0.2 foot thick.	7
226 - 229	Sandstone, subarkosic, reddish brown, medium to coarse grained.	3
229 - 239	Sandstone, subarkosic, reddish brown to light grey.	10
239 - 249	Sandstone, subarkosic to orthoquartzitic, reddish brown to light grey, fine to coarse grained, contains seven layers of siltstone up to 0.8 foot thick. Siltstone, is reddish brown, light grey and purple.	10
249 - 259	Sandstone and siltstone, subarkosic, reddish brown to grey, crossbedded; primary bedding is horizontal.	10
259 - 269	Sandstone, subarkosic to orthoquartzitic, reddish brown to grey, medium to fine grained; five laminated siltstone bands 0.02 to 0.4 foot thick.	10
269 - 279	Sandstone, subarkosic, dark red to red, coarse to fine grained, crossbedded; primary bedding is horizontal.	10
279 - 289	Sandstone, subarkosic to arkosic, red to pink coarse to fine grained, four bands of siltstone 0.05 to 0.2 foot thick.	10
289 - 329	Sandstone, arkosic, red to pink, fine to medium grained; siltstone interbeds vary in thickness from micro to 0.2 foot	40
329 - 339	Sandstone, arkosic, reddish to brown; primary bedding is horizontal; some crossbedding.	10
339 - 348.5	Sandstone, arkosic, reddish to brown, fine to medium grained.	9.5
348.5 - 349.3	Conglomerate, pebbles of quartz and feldspar 0.03 inch in diameter, in red siltstone matrix.	0.8
349.3 - 359	Sandstone and pebble conglomerate, arkosic, reddish brown; some beds of siltstone. Siltstone beds average 0.4 foot thick, varying from micro to 1.0 foot thick.	9.7

Depth (feet)	Description	Thickness (feet)
359 - 369	Sandstone and conglomerate, arkosic, reddish to brown pebbles of feldspar and quartz.	10
369 - 372.8	Sandstone and conglomerate, arkosic, brownish red to pink, thin laminations of magnetite.	3.8
372.8 - 373.6	Sandstone, arkosic, red, medium grained.	0.8
373.6 - 379	Conglomerate	5.4
379 - 382	Conglomerate, light green.	3
382 - 382.8	Sandstone, red, fine grained with some rounded fragments of feldspar, quartz and siltstone.	0.8
382.8 - 383.9	Siltstone, red, laminated.	1.1
383.9 - 387.5	Conglomerate, brown to pink with some layers of red siltstone 0.02 to 0.2 foot thick.	3.6

Depth (feet)	Description	Thickness (feet)
	PRECAMBRIAN GRANITIC GNEISS (101.5 feet, incomplete)	
387.5 - 389	Gneiss with biotite, fine to medium grained; foliation subhorizontal; kaolinite, particles of glauconite.	1.5
389 - 399	Gneiss with biotite, dark brown to pink, fine to medium grained; foliation subhorizontal to dip 10°.	10
399 - 409	Gneiss, feldspathic, micaceous (biotite), pink to brown; foliation subhorizontal to dip 20°.	10
409 - 417.5	Gneiss, feldspathic (orthoclase), slightly biotitic; foliation subhorizontal to dip 20°.	8.5
417.5 - 419	Gneiss, biotitic, reddish brown, with rusty stain, fine to medium grained.	1.5
419 - 429	Gneiss, biotitic, feldspathic, reddish brown, traces of kaolinite; fine to coarse grained; some layers of feldspar and quartz 0.01 to 0.1 foot thick; foliation (gneissosity) subhorizontal to dip 20°.	10
429 - 439	Gneiss, feldspathic, slightly biotitic, fine to coarse grained; reddish brown to pink, gneissosity subhorizontal to dip 20°.	10
439 - 449	Gneiss, feldspathic, pink, fine to coarse grained.	10
449 - 469	Gneiss, feldspathic, some biotite, pink, coarse grained, gneissosity subhorizontal.	20
469 - 489	Gneiss, feldspathic, gneissosity subhorizontal to dip 20°.	20
	Total depth of hole 489 feet.	

PART 3

GEOLOGY OF THE HARE BAY ALLOCHTHON

Harold Williams and W.R. Smyth

INTRODUCTION

The Hare Bay Allochthon occurs in western Newfoundland at the northern tip of the Great Northern Peninsula. The Allochthon extends for approximately 130 km along the eastern shore of the Great Northern Peninsula and up to 32 km inland. Hare Bay exposes an excellent cross section of the Allochthon and affords a marine passage through its central part. Canada Bay affords another well-exposed cross section at its narrower southern part.

St. Anthony is the largest community in the area and is the site of the Grenfell Mission and a new modern hospital that serves northern Newfoundland and Labrador. It is accessible by land-based aircraft, gravel road, and Canadian National coastal service. All of the coastal communities north of Hare Bay are now connected by secondary gravel roads. South of Hare Bay, most coastal communities are also connected by gravel roads, except for Grandois, St. Julien's Island, and Fishot Islands.

Geological setting

The geology of western Newfoundland is typical of that all along the western side of the Appalachians and consists of three major tectonic elements.

(1) A Precambrian crystalline basement deformed during the Grenvillian orogenic cycle.
(2) An autochthonous mainly Cambrian - Ordovician sequence that unconformably overlies the Precambrian rocks. It includes a thin basal clastic-volcanic unit, a middle carbonate unit, and an overlying easterly-derived clastic unit.
(3) Transported assemblages that were emplaced above the autochthonous sequence during the Middle Ordovician. The transported rocks comprise two large allochthons in western Newfoundland: the Humber Arm Allochthon in the south and the Hare Bay Allochthon in the north. The lower structural slices of the allochthons consist mainly of Cambrian to Lower Ordovician clastic sedimentary rocks. These are overlain by higher structural slices that consist of igneous and metamorphic rocks. The highest slice of both allochthons contains rocks of the ophiolite suite with an attached metamorphic aureole. All of the slices are separated from one another and from the autochthon by thin zones of black shale mélange.

Previous work

The earliest geological investigations in the Hare Bay area were reconnaissance surveys by Murray (Chapt. 2 in Murray and Howley, 1881), Richardson (in Logan, 1863) and Howley (in Murray and Howley, 1918, p. 484 - 501). The first comprehensive map and geological report of the Hare Bay area was produced by Cooper (1937). Shortly afterward, Betz (1939) published a map and report that summarized his studies in the Canada Bay area.

The implications of transported rock sequences in western Newfoundland were first worked out by Rodgers and Neale (1963) and these ideas prompted recent studies. Gillis (1966) and Stevens (1968) mapped most coastal exposures of the Allochthon in Hare Bay, and Tuke (1968) mapped the area north of Hare Bay. Smyth (1971) and Williams (1971) summarized the geology of the Hare Bay Allochthon and reviewed the history of changing ideas that bear upon transported sequences in western Newfoundland. Williams and Smyth (1973), reviewed the structural setting of the White Hills Peridotite and interpreted its metamorphic aureole as a transport phenomenon.

Baird (1966) described the Carboniferous rocks of the Conche - Grey Islands area and Kennedy et al., (1973) described the geology of Grey Islands.

Williams (1975) has summarizead the structural succession and nomenclature of all the rocks in the Hare Bay Allochthon and made comparisons with rocks of the Humber Arm Allochthon. Jamieson (1977) has conducted petrologic and geochemical studies on some of the igneous rocks of the Hare Bay Allochthon.

Present study

The present memoir is based upon field mapping carried out during the summers of 1972 and 1973. It also includes the results of the second author's Ph.D. studies carried out in the area south of Hare Bay during 1969 - 71 (Smyth, 1973). The report is concerned mainly with the transported rocks, although the autochthonous rocks are also described where they have been mapped around the western margin of the Allochthon and delineated as inliers within the Allochthon. The Grey Islands were mapped in 1969 (Kennedy et al., 1973) and the Carboniferous rocks at Conche were remapped during the 1973 field season.

The authors wish to thank Alvin Crocker of Trout River, Newfoundland, for his assistance and jovial companionship during the summers of 1972 and 1973 and they wish to thank the many residents of Conche, Goose Cove, St. Anthony, and Fishot Islands for field assistance and continued hospitality.

Physiography and glaciation

The major tectonic elements of western Newfoundland are clearly reflected in the present topography. Precambrian crystalline rocks form the central core of the Great Northern Peninsula and these are exposed in the highlands in the southern part of the map area. The autochthonous sequence forms coastal lowlands that flank the crystalline highlands to the west and north and these are covered by muskeg and a myriad of small ponds. The Hare Bay Allochthon forms a rolling upland terrain at the northeast part of the peninsula. A prominent scarp marks the western edge of the Allochthon and the structural slices that comprise the Allochthon are in most places morphologically distinct, especially the ophiolite slice. The slice is surrounded by a prominent low ridge and its structural base is a clear line of demarcation along the north side of Hare Bay slightly above tide level. The volcanic rocks at Raleigh and the ultramafic rocks at Howe Harbour (Deer Barren) are other examples of morphologically distinct transported slabs.

A prominent ridge that runs from Great Islets Harbour to Croque Harbour is the topographic expression of resistant basalt, and similar ridges occur north of Coles Pond. The long, linear depression within the uplands south of Great Islets Harbour is underlain by carbonate and represents an erosional window through the Allochthon, the Whites Arm Window (Smyth, 1973). A prominent northeast-trending topographic grain reflects cleavage, faults, and fold axes. Locally at St. Anthony Bight and eastward, northwest-trending ridges are controlled by earlier cleavage and fold structures that are cut at high angles by the regional northeast-trending structures. The topography at Milan Arm, where rounded islands and hills of various size are surrounded by lower terrain, results from the chaotic nature of underlying mélange deposits. Similar topography characterizes coastal exposures of similar mélanges throughout Newfoundland, for example, Dildo Run, Carmanville, Woods Island, etc.

Most of the Allochthon comprises Maiden Point greywacke that constitutes rounded hills with stunted tree growth and intervening depressions with swamps, ponds, or more thickly wooded areas. The White Hills Peridotite is barren of trees and its contact with underlying amphibolite coincides with the abrupt appearance of spruce and fir.

Glacial till is patchy and thin. Erratics are common. Glacial striations and grooves trend east to southeast and this direction of ice movement is supported by the occurrence of conspicuous erratics of anorthosite and complexly deformed pink gneiss that must have originated to the northwest in Labrador. Milan Arm, St. Anthony Bight, St. Anthony Harbour, Northwest Arm, Howe Harbour, and Croque Harbour are all aligned in the direction of glacial ice movement.

The shoreline is gently sloping in most places and is easily accessible by boat. In Sacred Bay and in much of Pistolet Bay, tidal benches are formed across easily erodable shale and mélanges so that boat work is difficult where these are strewn with large glacial erratics. A recent emergence of the land is shown by raised beaches and wave-cut terraces that reach more than 60 m above present sea level. A prominent wave-cut terrace is preserved in the mountain slopes around Howe Harbour and coarse cobble beach deposits are evident at several levels on the south side of Bell Island. The straight shoreline along the east side of the Allochthon continues to White Bay and is controlled by steep northeast-trending faults.

GENERAL GEOLOGY

The rocks of the Hare Bay area can be subdivided into three contrasting groups as follows: (i) crystalline basement rocks that are inliers of the Grenville Structural Province; (ii) a Cambrian - Ordovician sequence that unconformably overlies the crystalline basement; and (iii) a variety of transported sequences in separate slices that structurally overlie the autochthonous sequence and that collectively constitute the Hare Bay Allochthon. Metamorphic rocks at Grey Islands and Carboniferous cover rocks form additional groups.

The basement rocks are known as the Basement Gneiss Complex. They form the highlands in the southern part of the map area and their present mapping and investigation has been the chief concern of H.H. Bostock (this memoir). We have mapped the Basement Gneiss Complex only south of Canada Bay where it has been involved in local thrusting. The basement rocks there are mostly retrograded gneiss that occurs structurally within the autochthonous sequence.

The autochthonous sequence surrounds the Basement Gneiss Compex to the north and it is exposed mainly to the west of the Hare Bay Allochthon. These rocks have been studied by L.M. Cumming (this memoir), although we have also mapped the exposures along the west margin of the Allochthon and the inliers within the Allochthon southwest of Great Islets Harbour, near Griquet, Noddy Bay and Foirou Island. This sequence can be divided into a Hadrynian to Lower Cambrian basal clastic-volcanic unit (Bateau, Lighthouse Cove, and Bradore formations); a Lower Cambrian to Middle Ordovician carbonate unit (Devils Cove, Forteau, Cloud Rapids, Treytown Pond, St. George and Table Head formations); and a Middle Ordovician upper clastic unit (Goose Tickle Formation and metamorphic equivalent Sugarloaf Schist). All these rocks form an essentially continuous sequence.

The Grey Islands are underlain by polydeformed and metamorphosed mainly sedimentary rocks that are correlated with the Fleur de Lys Supergroup of the Burlington Peninsula to the south of the map area (Kennedy et al., 1973). These rocks are interpreted as Hadrynian - Cambrian eastward correlatives of the basal clastic-volcanic unit of the autochthonous sequence. At Bell Island, the metamorphic rocks are cut posttectonically by the Bell Island Granite that is dated isotopically by K-Ar (muscovite) at 368 ± 16 Ma (Wanless et al., 1973, p. 110).

The Hare Bay Allochthon comprises six contrasting rock groups in separate structural slices that range in age from Hadrynian - Lower Cambrian to Lower Ordovician. From structurally lowest to structurally highest, the transported rocks are as follows: Northwest Arm Formation, mainly shale and sandstone of Early Ordovician age; Maiden Point Formation, mainly greywacke, volcanic rocks and dykes of inferred Hadrynian or Early Cambrian age; Grandois Group that includes the Irish Formation (sandy limestone) and St. Julien Island Formation* (polymictic conglomerate) of inferred Early Ordovician or earlier age; Milan Arm Mélange comprises a mixture of blocks and shale matrix of Hadrynian to Early Ordovician age; Cape Onion Formation, mainly pillow lava and minor black shale of Early Ordovician age; St. Anthony Complex that includes the Ireland Point Volcanics, Goose Cove Schist, and Green Ridge Amphibolite, all possible Maiden Point or Cape Onion correlatives, and the White Hills Peridotite of probable Early Ordovician age.

The transported slices are separated in most places by mélange zones that vary from a metre to tens of metres in thickness. The mélanges are everywhere similar and consist of a variety of boulders and larger blocks in a black or black and green shale matrix. The most common blocks are sandstone of the Maiden Point and Northwest Arm formations and the mafic volcanic rocks of the Maiden Point, Ireland Point, and Cape Onion formations. The matrix is everywhere similar to shale of the Northwest Arm Formation. The Milan Arm Mélange is distinctive in that it has a preponderance of abnormally large ultramafic, amphibolitic, and volcanic blocks, but its matrix shale is like that of the other mélanges. In a few places where relatively high structural slices lie directly upon the upper clastic unit of the Authochthon, the contact is a hard thrust and mélange is sparse or absent, for example Cape Onion Formation above Goose Tickle Formation at Raleigh, Maiden Point Formation above Goose Tickle Formation at Sugarloaf Cove and Croque.

West of Locks Cove, carbonate of the Table Head Formation is interpreted to occupy a small slice among the transported rocks. This carbonate is viewed as a parautochthonous tectonic sliver ripped up and incorporated among the farther travelled slices during emplacement. At Hare Island, carbonate of the Table Head Formation and St. George Group are thrust westward over the Northwest Arm Formation. This thrusting postdates emplacement of the Allochthon. Similar thrusts brought the Table Head Formation or St. George Group above the younger Goose Tickle Formation at Raleigh and Brent Islands.

* The St. Julien Island Formation is now correlated with conglomerates at Sops Arm to the south and assigned a Silurian age. This change has been made on the accompanying map.

TABLE OF FORMATIONS

Era	Age	Formation		Lithology
Paleozoic	Carboniferous	Cape Rouge Formation		Grey to brown sandstone, plant-bearing siltstone, shale
Paleozoic	Carboniferous	Crouse Harbour Formation		Brown to red conglomerate, minor sandstone
	Unconformably upon Fleur de Lys Supergroup; faulted against Maiden Point Formation			
Paleozoic	Early Carboniferous to Late Devonian			Intermediate to acid dykes and lamprophyres
	Cuts Goose Cove Schist, mélange, and Fleur de Lys Supergroup			
Paleozoic	Devonian	Bell Island Granite		Grey to pink medium grained microcline granite
	Posttectonically cuts Fleur de Lys Supergroup			
Paleozoic	Early Ordovician or Earlier	St. Anthony Complex	White Hills Peridotite	Serpentinized hartzburgite, dunite, lherzolite
		St. Anthony Complex	Tectonic contact welded by metamorphic recrystallization	
Paleozoic	Early Cambrian-Hadrynian to Early Ordovician	St. Anthony Complex	Green Ridge Amphibolite	Pyroxene and garnetiferous amphibolite, minor marble, gabbro
		St. Anthony Complex	Metamorphic gradation	
Paleozoic	Early Cambrian-Hadrynian to Early Ordovician	St. Anthony Complex	Goose Cove Schist	Greenschist, agglomerate and tuff, greywacke, shale, limestone, gabbro
		St. Anthony Complex	Metamorphic gradation	
Paleozoic	Early Cambrian-Hadrynian to Early Ordovician	St. Anthony Complex	Ireland Point Volcanics	Purple, red and green agglomerate, green pillow lava
	Structurally upon Cape Onion Formation, Milan Arm Mélange, Maiden Point and Goose Tickle formations			
Paleozoic	Early Ordovician	Cape Onion Formation		Pillow lava, mafic pyroclastic rocks, black shale
	Structurally upon Maiden Point and Goose Tickle formations			
Paleozoic	Early to Middle Ordovician	Milan Arm Mélange		Plutonic, volcanic and greywacke blocks in shale matrix
	Structurally upon Maiden Point and Goose Tickle formations			
Paleozoic	Early Ordovician or Earlier	Grandois Group	Irish Formation	Sandy limestone, minor quartzite and greenschist
		Grandois Group	Fault Contact	
Paleozoic	Early Ordovician or Earlier	Grandois Group	St. Julien Island Formation	Polymictic red to purple conglomerate, greywacke, minor gabbro
	Structurally upon Maiden Point Formation			
Paleozoic	Early Cambrian to Hadrynian	Maiden Point Formation		Greywacke, pebble conglomerate, mafic pyroclastic rocks and pillow lava, mafic dykes, cordierite-andalusite hornfels
	Structurally upon Northwest Arm and Goose Tickle formations			
Paleozoic	Early Ordovician	Northwest Arm Formation		Black and green shale, sandstone, limestone
	Soft tectonic contact with mélange			
Paleozoic	Middle Ordovician	Goose Tickle Formation		Grey siltstone, shale, greywacke; pelitic and calc-silicate schist of Sugarloaf Schist Member
Paleozoic	Middle Ordovician	Table Head Formation		Grey hackly weathering limestone
	Local erosional disconformity			
Paleozoic	Early Ordovician	St. George Group		Light to dark grey and buff dolomite and limestone
Paleozoic	Middle Cambrian	Eddies Cove Formation	Treytown Pond Formation	Limestone and dolomite
Paleozoic	Middle Cambrian	Eddies Cove Formation	Cloud Rapids Formation	Blue-black, fine grained limestone
Paleozoic	Early Cambrian	Forteau Formation		Grey shale and nodular limestone
Paleozoic	Early Cambrian	Devils Cove Formation		Purple and white limestone
	Conformable contact with Bradore Formation			
Paleozoic and Precambrian	Early Cambrian and Hadrynian	Fleur de Lys Supergroup		Psammitic to pelitic schist, actinolite-chlorite schist
		Contact not exposed		
Paleozoic and Precambrian	Early Cambrian and Hadrynian	Bradore Formation		Arkosic sandstone, pebble conglomerate
Paleozoic and Precambrian	Early Cambrian and Hadrynian	Lighthouse Cove Formation		Mafic volcanic rocks and related dykes
Paleozoic and Precambrian	Early Cambrian and Hadrynian	Bateau Formation		Quartzite, conglomerate
	Unconformity			
Precambrian	Helikian and (?) Earlier	Basement Gneiss Complex		Mainly granitic gneiss

Some of the transported rocks of the Hare Bay Allochthon were deformed and metamorphosed before emplacement. The structural styles are thought to relate to initial uprooting and to the early transport processes. Indirect geological reasoning indicates that the Allochthon was emplaced during Middle Ordovician. Since emplacement, all of the autochthonous and allochthonous rocks were involved in upright folds about northeast-trending axes.

Intermediate to mafic dykes and a few lamprophyres cut the transported rocks and mélanges posttectonically. A small lamprophyre dyke at the northwest tip of Groais Island cuts the Fleur de Lys Supergroup and similar lamprophyre occurs as clasts in nearby Carboniferous conglomerate. The Groais Island lamprophyre dyke is dated by K-Ar (biotite) at 353 ± 16 Ma (Wanless et al., 1973, p. 109-110).

Carboniferous rocks at Conche and Cape Rouge peninsulas are sandstone and conglomerate referred to the Crouse Harbour and Cape Rouge formations by Baird (1966). The basal Crouse Harbour Formation unconformably overlies the Fleur de Lys Supergroup at Groais Island and it is faulted against the Maiden Point Formation toward the west. The Carboniferous rocks are only mildly deformed in broad open folds and lack penetrative cleavage in most places.

BASEMENT GNEISSES

The basement gneisses were studied mainly from a structural viewpoint where they were thrust and incorporated within the autochthonous sequence at the south end of the Hare Bay Allochthon. These gneisses are retrograded with a superimposed fabric so that they are atypical of the basement gneisses regionally.

Basement Gneiss Complex

Basement gneisses exposed in the large inlier at the southern margin of the map area were referred to the Long Range Complex (Clifford and Baird, 1962; Smyth, 1973) and Basement Gneiss Complex (Bostock, this memoir). South of the Hare Bay Allochthon, and southwest of Sugarloaf Cove, the complex consists of pink massive to well-foliated quartz-plagioclase-potash feldspar-biotite gneiss that is intruded posttectonically by pink microgranite. Mafic gneiss, which represents pretectonic dykes and sills, occurs locally among the pink gneiss. Similar rocks, which include retrograded metagabbro, occur along the west side of Canada Bay.

Southwest of Sugarloaf Cove a small area of gneiss has been involved in thrusting or reverse faulting, for a northeast-dipping unconformity between the Basement Gneiss Complex and the Bradore Formation is followed northeastward by retrograded crystalline rocks. The retrograded rocks have a gneissic fabric and are cut by microgranite similar to that within the Basement Gneiss Complex farther south. Near Burnt Point on the south side of Canada Bay, another occurrence of retrograded gneiss and microgranite forms a thin slice within a thrusted belt of Bateau and Bradore rocks.

Gneissic banding in the Basement Gneiss Complex at Sugarloaf Cove is cut by a subhorizontal schistosity that increases in intensity eastward toward the structural top of the retrograded rocks. Original feldspar is completely sericitized and saussuritized and the sericite, where aligned, defines the superimposed fabric. Biotite flakes are kinked, broken, and partly chloritized, and hornblende is altered to chlorite. In the most deformed and retrograded examples, the acid gneisses consist of isolated quartz and chloritized biotite crystals set in a sericitic groundmass. These rocks resemble deformed sediments, though the large kinked biotite crystals are distinctive.

The thrusted gneisses at Burnt Point are similarly retrograded and the fabrics are largely the result of superimposed shearing and recrystallization.

Locally west and northwest of Sugarloaf Cove, the superimposed subhorizontal fabric is folded about upright northeast-trending axial surfaces. The sequence of events suggests that the subhorizontal fabric represents a penetrative deformation (late Taconic) that immediately followed or accompanied the final emplacement of the Hare Bay Allochthon. The later folding may correlate with the regional postemplacement folding about northeast axes (Acadian).

AUTOCHTHONOUS ROCKS

The autochthonous rocks were mapped in all coastal exposures from Pistolet Bay to south of Canada Bay, including all of Hare Bay. They were also mapped inland along the western margin of the Allochthon and where they occur as inliers within the Allochthon. Structural features and relationships to the transported rocks were stressed throughout, and no attempt was made to separate the main lithic units of the Ordovician carbonate sequence.

Basal clastic-volcanic unit

Three formations are recognized within the basal clastic-volcanic unit. The lower two are known only from this northern area of Newfoundland and nearby Labrador. The upper formation is more widespread in western Newfoundland beneath the carbonate sequence. Thickness variations are erratic but the regional pattern suggests the appearance of older and thicker units beneath the carbonate sequence from west to east across western Newfoundland.

Bateau Formation

A thick sequence of white quartzite exposed at White Islands and White Rocks east of Quirpon was assigned to the Bateau Formation by Williams and Stevens (1969). These rocks were previously assigned to the Bradore Formation (Gillis, 1966; Tuke, 1968), as the lithology and thickness are typical of the Bradore Formation. The rocks, however, more closely resemble the thick quartzite sequence of the Bateau Formation that unconformably overlies the basement gneiss at Belle Isle (Williams and Stevens, 1969). The thickness of the sequence is estimated at approximately 460 m with neither top nor base exposed. A coarse plutonic boulder conglomerate that occurs between Long Range microgranite and mafic volcanic rocks just west of Burnt Point, Canada Bay, resembles the Bateau conglomerate at Belle Isle and is assigned to the Bateau Formation. The conglomerate is about 6 m thick and its contacts with nearby rocks are poorly exposed or modified by faults. A metre or more of conglomerate also intervenes locally between crystalline basement and overlying mafic volcanic rocks at Otter Cove of Canada Bay

In its type area, the Bateau Formation is conformably overlain by the Lighthouse Cove Formation, followed by the Bradore Formation that is in turn conformably overlain by the Lower Cambrian Forteau Formation.

The quartzite at White Islands is well sorted with well-rounded grains. The bedding is regular with individual beds ranging in thickness from 15 cm to 1 m. Some beds contain magnetite-rich lenses; pink potash feldspar is evident in a few places. A 15-m unit of grey and purple slate occurs within the quartzite at the north end of the largest of the White Islands. The quartzite strikes northerly and most of the White Islands occupy the east limb of an upright anticline. The rocks have a pronounced steeply east-dipping cleavage.

The conglomerate near Burnt Point has well-rounded cobbles and boulders that consist of pink granite gneiss, amphibolite, and quartzite. The matrix is a fine schistose mudstone.

Lighthouse Cove Formation

Thin altered basalt flows above the Basement Gneiss Complex at the south shore of Otter Cove and those in a thrust slice west of Burnt Point are assigned to the Lighthouse Cove Formation. This name was first introduced by Williams and Stevens (1969) to designate volcanic rocks that overlie basement gneiss at Belle Isle. It was later extended to include volcanic rocks that occupy similar stratigraphic levels in southeast Labrador and Canada Bay (Strong and Williams, 1972). The main outcrops of the Lighthouse Cove Formation in Canada Bay occur at Cloud Hills (Clifford, 1965; Bostock, this memoir).

At Otter Cove, about 9 m of lava with the top of the formation unexposed outcrops in two small separate areas. The lava either rests directly on the Basement Gneiss Complex or else is underlain by up to 3 m of coarse sandstone and pebble conglomerate. The southernmost exposure consists of two flows separated by 1 m or more of pebble conglomerate with quartzite clasts. Vesicular pillows occur locally at the base of the flows and hexagonal cooling joints toward their top.

West of Burnt Point the Lighthouse Cove Formation is represented by a thin, altered, green unit that outcrops between retrograded basement gneiss or coarse boulder conglomerate and Bradore sandstone. Most of the contacts in this area have been modified by faults, although these rocks west of Burnt Point occur in normal stratigraphic order from west to east.

Bradore Formation

Sandstone that unconformably overlies the Basement Gneiss Complex and follows the Lighthouse Cove Formation is assigned to the Bradore Formation (Schuchert and Dunbar, 1934). These rocks were referred to the Cloud Mountain Formation (Betz, 1939), although they were correlated with the Bradore Formation.

The sandstone outcrops along the north shore of Otter Cove, the type locality of Betz's Cloud Mountain Formation, and southwards along the west shore of Wild Cove. From Wild Cove the beds continue inland and reappear on the shore near Sugarloaf Cove. A small area of Bradore sandstone occurs at Burnt Point above the Lighthouse Cove Formation. Unconformable relations with the Basement Gneiss Complex are exposed at Otter Cove, Fly Point, northwest of Wild Cove, and southwest of Sugarloaf Cove. The sandstone is conformably overlain by Lower Cambrian limestone of the Devils Cove (Forteau) Formation at Otter Cove and Wild Cove.

Betz (1939) reported 178 m of strata near Dieppe Point but the formation is slightly less than 60 m thick west of Wild Cove. The Wild Cove thickness compares with 67 to 70 m at the Bradore type section in southeast Labrador (Schuchert and Dunbar, 1934) and with 91 to 122 m at the abnormally thick Belle Isle section (Williams and Stevens, 1969). Thrust faults near Dieppe Point, which repeat the top of the section, may account for Betz's figure.

The Bradore Formation consists of grey to white and pale brown to purple arkosic sandstone with pebble conglomerate beds toward the base. Thin discontinuous coarse basal lenses that are evident southwest of Sugarloaf Cove and west of Wild Cove probably represent infillings of local depressions in the Basement Gneiss Complex. Succeeding parts of the sections display alternating grey and green to reddish sandstone with local pebble beds, and upper parts of the sections contain distinctive purple to black hematitic sandstone. West of Fly Point near the top of the exposed Bradore section, a 7-cm-thick bed of purple crystalline limestone is interbedded with medium-grained, purple to black sandstone.

Megaripples occur near the base of the formation and large scale tabular and trough crossbeds and graded beds are common in places. South of Fly Point, the orientation of crossbeds indicates paleocurrents from the southwest.

The Bradore Formation is unfossiliferous, except for the vertical tubes *Scolithis linearis*, in its type section at Labrador. It is conformably overlain by fossiliferous Lower Cambrian beds and it is therefore probably of Early Cambrian age.

Metamorphic equivalents of basal clastic-volcanic unit at Grey Islands

Metamorphic rocks at Grey Islands are the northeast extremity of a belt of similar rocks that borders the carbonate sequence of western Newfoundland (Zone C of Williams et al., 1972). The metamorphic rocks are underlain by rejuvenated gneissic basement on Burlington Peninsula (de Wit, 1974) and most reasoning suggests that they are of Hadrynian and Cambrian age. Their main period of metamorphism and deformation is certainly pre-Silurian and possibly Late Cambrian (Church, 1969; Kennedy, 1973), but an Early to Middle Ordovician age is favoured here. Lithologic comparisons, coupled with age and geological setting, suggest that the rocks are essentially autochthonous and metamorphosed equivalents of the basal clastic-volcanic unit of the Autochthon farther west.

Fleur de Lys Supergroup

The metasedimentary rocks that constitute most of Groais Island and the western part of Bell Island are correlated with the lithologically, structurally, and metamorphically similar Fleur de Lys Supergroup of the Burlington Peninsula (Kennedy et al., 1973). The metamorphic rocks are predominantly coarse psammitic schist and semipelitic to pelitic schist that were derived from quartzose greywacke, fine-grained siltstone and shale. Original bedding is evident in some places and grading is locally preserved in the coarser clastic beds. A distinctive unit of actinolite-chlorite schist on Groais Island was probably derived from mafic tuff and agglomerate. The metamorphic rocks were divided into five lithic units and arranged in a tentative stratigraphic order using a few top determinations in widely separated areas (Kennedy et al., 1973). The lower three units consist of pelitic to psammitic schist and pebbly quartzite that are restricted to Bell Island. The upper two units are chlorite-actinolite schist and overlying pebbly psammitic schist. These units are restricted to Groais Island. Intensity of small-scale folding within each unit makes estimates of original thickness impossible.

Lithology

The Fleur de Lys Supergroup of Bell Island is predominantly albite-muscovite schist with some chlorite, biotite, zoisite, and minor garnet. Local impure marble beds and small pods of actinolite schist occur in the psammitic schist. Albite porphyroblasts are common and in places comprise more than 50 per cent of the rocks. Tourmaline, magnetite, epidote, and sphene are common accessory minerals.

The chlorite-actinolite schist of Groais Island is thinly bedded and contains diagnostic small magnetite octahedra. Buff-weathering carbonate-rich schist, graphitic biotite schist, albite-muscovite-chlorite schist, and epidote-actinolite schist are all represented in small amounts either within or near the contacts of this unit.

The pebbly psammitic schists of Groais Island have pelitic interbeds and thin layers of graphitic-muscovite-albite pelite. Clasts in the psammite are grey to blue quartz with lesser plagioclase.

Thin sheets of hornblende-albite amphibolite at Bell Island probably represent metamorphosed mafic sills, and pods and lenses of fuchsite-actinolite schist in the chlorite-actinolite unit may represent deformed and metamorphosed ultramafic rocks.

Structure

Two penetrative schistosities and two strain-slip fabrics are present in the metamorphic rocks. The relationships between these fabrics, metamorphic mineral growth, and associated structures led to the recognition of a sequence of four deformational events (Kennedy et al., 1973). The first schistosity (S_1) is subparallel to bedding and pebbles in psammitic beds are flattened parallel to this fabric. No small scale F_1 folds are apparent. The second schistosity (S_2) is the prominent penetrative foliation. It is a micaceous banding formed by folding and transposition of the earlier fabric. S_2 forms the axial plane foliation of tight to isoclinal, microscopic to macroscopic F_2 folds that fold the S_1 schistosity. The F_2 folds plunge gently between northeast and northwest. On Bell Island they are reclined to recumbent structures that show vergence upward toward the west, indicating that the metamorphic rocks there occupy the lower limb of a major overturned fold that closes westward. On Groais Island the F_2 folds are upright to inclined and the systematic variation in vergence and local facing directions indicate that the chlorite-actinolite unit occupies the cores of F_2 anticlines (Kennedy et al., 1973). Variations in the attitude of the S_2 schistosity and second phase folds are attributed to a third structural event. Variations in the attitude of a third phase schistosity (S_3) on Groais Island are interpreted, in turn, as the result of open F_4 folds.

Metamorphism

The metamorphic mineral assemblages present in the Fleur de Lys Supergroup of Grey Islands fall within the greenschist facies. The textural relationships of the individual mineral phases indicate that growth of albite was prolonged at Bell Island, where garnet and inclusions of calcic plagioclase in albite suggest local amphibolite facies conditions. Retrogression is indicated by replacement of biotite by clorite and local chloritization of garnet.

Correlation

The chlorite-actinolite schist of Groais Island and its overlying psammitic unit are comparable to the Birchy Schist and overlying Mings Bight Group of the Fleur de Lys Supergroup in the type area. The metamorphic rocks of Bell Island are tentatively correlated with the White Bay Sequence of the Fleur de Lys Supergroup (Kennedy et al., 1973). Graded psammites with grey to blue quartz grains on Groais Island are lithologically similar to greywacke of the Maiden Point Formation and the Fishot Islands Member of the St. Anthony Complex on the west side of Fishot Islands. The chlorite-actinolite unit is lithologically similar to the Goose Cove Schist.

Carbonate sequence

A carbonate sequence is exposed across Canada Bay where it dips eastward away from the basement gneisses and their cover of the clastic-volcanic unit. Northward toward Hare Bay the dips become gentle so that only the upper part of the sequence is exposed along the western margin of the Hare Bay Allochthon. The upper part of the carbonate sequence is exposed in the central part of the Whites Arm Window through the allochthonous cover, and it is exposed beneath the Allochthon at Noddy Bay, Griquet, and Foirou Island.

The carbonate sequence is repeated by numerous thrust faults in Canada Bay, especially toward the east and nearest the Allochthon. Stratigraphic studies are difficult in this deformed area. Elsewhere, we have not studied all of the carbonate succession, notably the Middle Cambrian Cloud Rapids and Treytown Pond formations (Betz, 1939), both of which are replaced by the new name Eddies Cove Formation (Cumming, this memoir).

In describing the carbonate rocks, we use the well-established names of Schuchert and Dunbar (1934) where possible, rather than the local names of Betz (1939) in Canada Bay and Cooper (1937) in Hare Bay. Lower Cambrian carbonate and shale above the Bradore and below Middle Cambrian strata in the Strait of Belle Isle map area are referred to the Forteau Formation. But along Canada Bay, Betz (1939) named a lower limestone unit the Devils Cove Formation and an upper shale and limestone the Forteau Formation. Thus these two formations in Canada Bay, described below, are equivalent to the Forteau Formation elsewhere in the map area.

Devils Cove Formation

Bedded purple and white limestones that overlie the Bradore Formation with sharp conformable contact along the west shore of Canada Bay are referred to the Devils Cove Formation (Betz, 1939). The type section is at the northeast shore of Otter Cove where about 15 m of limestone, with top unexposed, intervenes between the Bradore Formation and shale and limestone mapped by Betz (1939) as the Forteau Formation. At the west shore of Wild Cove, approximately 9 m of purple limestone and thin minor shale conformably overlies the Bradore Formation. Southward to Wild Cove, similar limestone occurs in fault contact with the Goose Tickle Formation. There, the limestone is sheared and recrystallized and forms a mylonite up to 1 m from the fault. A similar relationship was observed west of Burnt Point where a 3-m zone of sheared limestone overlies the Bradore Formation and is in fault contact with the Goose Tickle Formation.

The Devils Cove Formation is fossiliferous, and poorly preserved gastropods were noted in pink limestone west of Dieppe Point. At Wild Cove, the pink limestone contains microfragments of algae filaments and brachiopod shells. Betz (1939) collected a variety of gastropods and trilobites from localities farther north and outside the present area of study that indicate an Early Cambrian age.

Forteau Formation

Grey shale and sandy limestone of the Forteau Formation (Schuchert and Dunbar, 1934; Betz, 1939) occur near the western margin of the Hare Bay Allochthon along the northwest shore of Canada Bay. Along the south side of Canada Bay the formation is cut out by thrust faults. At Dieppe Point the lowermost beds parallel the Devils Cove Formation across a small exposure gap and consist of thin interbeds of grey limestone, mudstone, and shale. These are followed by blue sandy limestone and interbedded grey mudstone. More than 90 m of strata with top unexposed outcrop east of Dieppe Point, and a thickness of 213 m of strata has been estimated at Castor Cove (Betz, 1939).

The formation contains a variety of brachiopods, trilobites, gastropods, and cephalopods that indicate an Early Cambrian age (Betz, 1939) and support the correlation of the beds at Canada Bay with strata of the type locality in southeastern Labrador.

St. George Group

Lower Ordovician shallow water dolomite and minor limestone along the western margin of the Hare Bay Allochthon were referred to the Chimney Arm Formation at Canada Bay (Betz, 1939) and the Southern Arm Limestone at Hare Bay (Cooper, 1937). Subsequent workers assigned these rocks to the St. George Formation (Gillis, 1966; Tuke, 1968) and St. George Group (Cumming, this memoir). The St. George Group underlies the western half of the peninsula between Bide Arm and Chimney Bay and occurs locally in the southwest part of Hare Bay and in a thrust slice at Raleigh. It is also present just west of the Allochthon at Schooner and Burnt islands in Pistolet Bay. Along the south shore of Canada Bay, the St. George Group occurs in a thrust slice at Wild Cove Point and equivalents are probably represented by recrystallized limestone in thrust slices farther east. Megaripples, desiccation cracks, and algal mounds occur in the dolomite.

At Canada Bay, Betz (1939) estimated the thickness of the group at 549 m. The lowest beds are exposed along the east shore of Chimney Bay and consist of 45 m of thin-bedded, black hackly limestone. These are followed by about 457 m of alternating dolomite and sandy dolomite in beds from 30 cm to 1 m thick. The dolomite weathers white to buff or brown but is grey to pink on fresh surfaces. Thin chert beds and nodules occur towards the top of the sequence near Handy Harbour. The uppermost chert horizon is overlain by a few metres of dolomite followed conformably by black and grey limestone with only minor dolomite. The grey limestone marks a distinct lithic change and is assigned to the overlying Table Head Formation.

Copper (1937) estimated 457 m of St. George correlatives at Southern Arm of Hare Bay; about 100 m of beds, with base unexposed, occur at Pistolet Bay (Tuke, 1968).

Fossils have been collected from most occurrences of St. George dolomite and all indicate an Early Ordovician age (Cooper, 1937; Betz, 1939; Tuke, 1968; Cumming, this memoir).

Table Head Formation

The uppermost formation of the carbonate sequence consists of grey rubbly-weathering limestone and shale. The limestone was named the Bide Arm Formation (Betz, 1939) in Canada Bay and the Hare Island Limestone (Cooper, 1937) in Hare Bay and these were correlated with the Table Head Formation (Schuchert and Dunbar, 1934). The Table Head Formation occurs along the western side of the Hare Bay Allochthon and its upper part is exposed in the Whites Arm Window and smaller inliers at Noddy Bay, Griquet, and Foirou Island.

A complete section of the formation is represented on Burnt Island of Pistolet Bay where about 305 m of beds conformably overlie the St. George Group and are overlain by the Goose Tickle Formation (Tuke, 1968). Similar thicknesses of strata were reported in Hare Bay (Cooper, 1937) where the limestone is overlain conformably by Goose Tickle shale and siltsone. In Canada Bay the Table Head Formation is cut by thrusts and it is locally involved in recumbent folds so that no complete section is known, making thickness estimates difficult.

The Table Head Formation records a gradual change in lithology with more argillite toward its top. At Bide Head and Wild Cove Point, a basal member about 30 m thick consists of black to blue to white crystalline limestone with some interbedded dolomite. A much thicker middle member consists of a well-bedded sequence of black to grey to white limestone and marble. An upper member consists of thin-bedded, black hackly limestone with minor thin chert beds and dark grey to black shale. The upper member is best exposed at Big Spring Inlet and southwest of Great Islets Harbour. At Canada Bay, much of the upper member is shale with interbedded thin limy siltstone. Intraformational breccia occurs in thin beds near the top of the formation at the north shore of Coles Pond.

The Table Head Formation is folded and cleaved in the Griquet and Noddy Bay inliers but at Foirou Island the beds are subhorizontal and relatively undeformed. The beds are gently east-dipping in the thrust slice at Raleigh and striated bedding surfaces indicate tectonic transport toward the northwest. In Hare Bay, the beds are openly folded and locally cut by thrusts. The Hare Island thrust and its southward continuation to Fournier Point brings the Table Head Formation above the Goose Tickle and Northwest Arm formations. A similar thrust at Brent Islands and Marechal Island brings the Table Head above the Goose Tickle, although movement is limited along this structure, for at Direction Island it is only a zone of steep bedding inversion without detachment. Increasing intensity of deformation and recrystallization southward to Canada Bay precludes stratigraphic study where the carbonate is mostly fine-grained crystalline marble. At Englee Island, the Table Head Formation is involved in recumbent folding.

The Table Head Formation is fossiliferous but preservation in the area along the west side of the Hare Bay Allochthon is generally poor. Gastropods were noted in the middle unit on the north side of Bide Head and on the west shore of Bide Arm. The upper black limestone unit is abundantly fossiliferous and shelly fossils were noted at Big and Little Spring inlets, at the river gorge about 1 km west of the arm entering Great Islets Harbour from the southwest, at Raleigh and Hare Island, and at the southwest side of Coles Pond. Conodonts from the top of the black limestone member at the west side of Little Spring Inlet indicate an Arenig - Llanvirn age (Fåhraeus, 1970). The Table Head is dated throughout western Newfoundland as early Middle Ordovician (Whittington and Kindle, 1963).

Upper clastic unit

The upper clastic unit consists of easterly to northeasterly derived greywacke and shale, marking the end of carbonate deposition that existed in the Autochthon from Early Cambrian to Middle Ordovician. It forms a continuous narrow belt that parallels the western margin of the Hare Bay Allochthon from Pistolet Bay to Canada Bay and it is also exposed below the Allochthon in Whites Arm Window and in inliers in the vicinity of Quirpon. South of Canada Bay the clastics are repeated by thrust faults with the most easterly thrust slice dipping east beneath the Hare Bay Allochthon. South of Canada Harbour, the greywacke and slate appear to grade into semipelitic and calc-silicate schists, which are well exposed around Sugarloaf Cove in White Bay. The Sugarloaf schists (Smyth, 1973) are interpreted as a highly deformed and metamorphosed equivalent of the upper clastic unit. However, because of their special structural setting and lithological differences, and as other correlations are possible, these schist are described separately.

Goose Tickle Formation

The upper clastic unit was named the Goose Tickle Slate at Hare Bay (Cooper, 1937; Gillis, 1966) and the Englee Formation at Canada Bay (Betz, 1939). The name Goose Tickle Formation (Tuke, 1968) is used here for the entire area as the rocks are best exposed and contacts are most easily defined in the Hare Bay area.

The top of the Goose Tickle Formation is defined by a tectonic contact with the Hare Bay Allochthon. In most places the contact is marked by 10 m or more of black shaly mélange, but at Canada Bay the contact is a sharp thrust and mélange is absent. About 457 m of strata are exposed in Pistolet Bay (Tuke, 1968), about 240 m at Big Spring Inlet (Smyth, 1973), and 305 m at the type section at the head of Hare Bay (Stevens, 1970).

At Pistolet Bay and Hare Bay, the Goose Tickle Formation consists of medium-grained greywacke beds from several centimetres to 30 cm or more thick with interlayered thinner dark shale and siltstone beds. Graded beds are common with internal convolutions and ripple drift lamination at the top. Load casts and sole markings are also present locally. At the west shore of Northwest Arm, conglomerate with unsorted limestone pebbles and cobbles in a shale matrix occurs in thick units separated by shale. Other conglomerate units up to 10 m thick are interlayered with greywacke and siltstone at Howe Harbour and Northern Arm (Fig. 85). These contain clasts of limestone, black shale, a variety of buff-weathering limy siltstone and sandstone, and volcanic rocks. Most of the sedimentary clasts can be matched directly with the nearby Northwest Arm Formation, and the volcanic clasts have counterparts in the Allochthon as well. Local conglomerate lenses and coarse sandstone beds that contain similar detritus from the Allochthon occur southwest of Great Islets Harbour. Southward to Canada Bay, the Goose Tickle Formation displays a decrease in grain size, sand content, and bed thickness. There, dark grey slate to brown siltstone predominate with minor greywacke beds towards the top of the section.

Directional features at Big and Little Spring inlets indicate paleocurrents from the northeast, and a few determinations at American Tickle indicate currents from the east.

Rock fragments identified in thin section include diabase, felsite, myrmekite, granophyre, chert, limestone, ooliths, sandstone, shale, and chlorite grains with hematite and chromite inclusions. Heavy minerals include chromite, hematite, zircon and rare tourmaline. The chromite is translucent, reddish, and distinctive. Chromite in serpentinite grains occurs in some examples from the north of Hare Bay (R.K. Stevens, pers. comm. 1973).

In most places the Goose Tickle Formation is folded about northeast-trending axis and is cut by a single steep penetrative cleavage. The intensity and complexity of deformation increases eastward, from open folds without penetrative cleavage on the west side of Hare Bay and Pistolet Bay to tight upright folds with steep penetrative cleavage along the east side of the Allochthon. The most intense and complex deformation is at Canada Bay where the formation is mainly slate and phyllite with two cleavages. At Croque, the Goose Tickle Formation is schistose and contains extensive muscovite as a result of prograde metamorphism at the base of the Hare Bay Allochthon.

Poorly preserved brachiopods were noted in siltstone beds west of White Hump Pond and graptolite fragments occur in outcrops on the western shore of Coles Pond and at Little Spring Inlet. The best preserved and most abundant graptolites in the formation occur at Pistolet Bay (Tuke, 1968) and at the western shore of Hare Bay (Erdtmann, 1971a,b). These assemblages indicate an early Middle Ordovician (Llanvirn) age for the formation.

Sugarloaf Schist Member

Semipelitic and calc-silicate schists exposed on the coast at Sugarloaf Cove south of Canada Bay were referred to the Sugarloaf Schist Member of the Goose Tickle Formation (Smyth, 1973). The schists are thought to grade into the Goose Tickle Formation but inland exposure is poor and the contact was not observed. At the coast, the schists are adjacent to retrograded gneisses of the Basement Gneiss Complex across a wide mylonite zone, and they are overlain with sharp contact by less deformed greywacke of the allochthonous Maiden Point Formation.

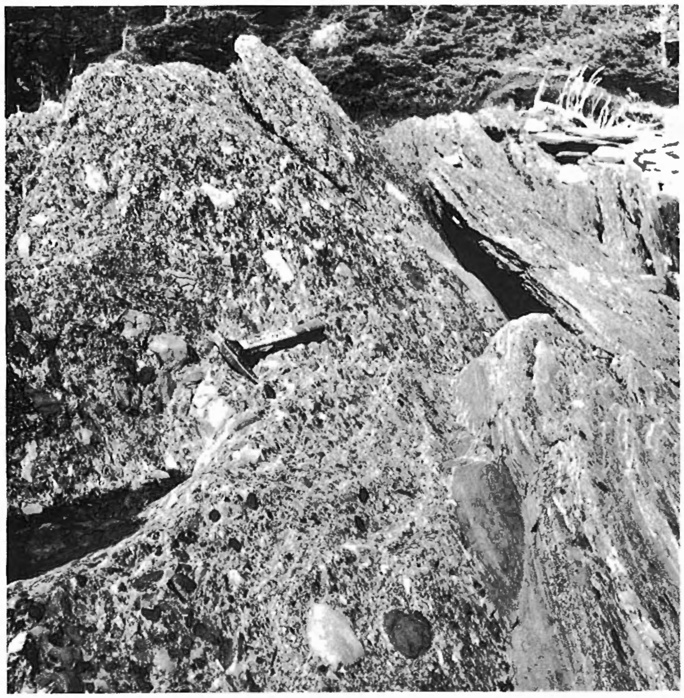


Figure 85. Conglomerate of Goose Tickle Formation comprises mainly detritus from the Northwest Arm Formation, Northern Arm, Hare Bay. (GSC 202609-D).

The Sugarloaf Schist is especially enigmatic as its rocks occur within an area of major thrusting and it is uncertain whether they represent a highly deformed slice of the thrusted autochthonous sequence, or whether they are a separate structural slice beneath the Maiden Point Formation and an integral part of the Hare Bay Allochthon. Most lines of reasoning support the authochthonous or parautochthonous nature of the schists as follows: (i) they occupy a similar stratigraphic and structural position to the Goose Tickle Formation that is stratigraphically above the Table Head and structurally beneath the transported Maiden Point Formation; (ii) the Maiden Point Formation is not underlain stratigraphically or structurally by similar schists elsewhere within the Hare Bay Allochthon; (iii) the Maiden Point Formation does not rest on any part of the Autochthon other than the Goose Tickle Formation; (iv) the structural base of the Maiden Point Formation is also a sharp apparently insignificant contact without mélange at Croque where the Goose Tickle Formation is schistose and exhibits prograde metamorphism; and (v) tectonic contacts of profound stratigraphic omission are common in the highly telescoped thrust belt along the south side of Canada Bay.

The Sugarloaf Schist comprises low-grade, regionally metamorphosed rocks with two cleavages. The rocks are fine grained, thin bedded, dark purple to dark grey, pelitic and semipelitic schist with thin white-weathered calc-silicate layers. The calc-silicate layers are conspicuous and diagnostic of the schists in inland exposures. Deformed marble beds up to 30 cm thick are interbedded with semipelite near the contact with the Maiden Point Formation. Graded psammite with pebbles of blue quartz and interbanded basic schist occurs below banded semipelitic and calc-silicate schist in a cliff face near the northwest corner of Sugarloaf Cove. The psammitic-basic schist horizon is about 6 m thick and appears to occupy the nose of a southwest-facing recumbent fold.

The semipelitic and pelitic schists contain biotite, muscovite, quartz, and oligoclase with minor epidote and zoisite, and accessory apatite, zircon, tourmaline, and iron ores. The calc-silicate beds are rich in carbonate and epidote and the basic schists are rich in biotite, actinolite, and magnetite.

The dominant fabric element of the Sugarloaf Schist is a penetrative S_1 schistosity that is refolded by tight inclined F_2 folds with associated S_2 crenulation cleavage. Minor F_3 folds locally affect the earlier structures.

The Sugarloaf Schist also exhibits the affects of posttectonic thermal metamorphism that is evident in the nearby Maiden Point Formation in this area south of Canada Bay. Posttectonic prophyroblasts include biotite, epidote, and cordierite in pelitic and semipelitic schist, and diopside and tremolite-actinolite in calc-silicate schist and marble.

The lithological differences between the Goose Tickle Formation and Sugarloaf Schist are largely attributed to metamorphism and deformation. The pelitic and semipelitic schists are thought to represent original Goose Tickle slate and greywacke, with the calc-silicate schists and marble representing thin limy siltstone and limestone interbeds.

The mylonite zone between the Basement Gneiss Complex and the Sugarloaf Schist is exposed in the cliffs about 305 m southwest of Sugarloaf Cove. It contains 30 m of phyllonitized granite overlain by a similar thickness of protomylonite. The phyllonitized granite is separated from the underlying retrograde Basement Gneiss Complex by an intervening thin faulted wedge of gritty psammite and dark green quartz-biotite schist. The fabric in the phyllonitized granite is locally refolded by minor folds, indicating that the mylonitization was an early event. The contact with the overlying protomylonite is sharp and is parallel to the fabric in both units. It is interpreted therefore as a fault of the same age as the mylonitization. Protomylonite is petrographically similar to schists of the Sugarloaf Schist Member and displays similar posttectonic thermal metamorphic growth of diopside, tremolite-actinolite, and minor sphene and vesuvianite in the calc-silicate foliae, and biotite and tremolite-actinolite in the semipelitic foliae. Tectonic banding in the schists is folded around minor second phase folds indicating that it is of the same early generation as the fabric in the phyllonitized granite. The mylonitization in this wide zone is interpreted as the result of intense shearing associated with the justapositioning of the Sugarloaf Schist Member of the Goose Tickle Formation and the Basement Gneiss Complex.

Paleographic setting of the autochthonous rocks

The stratigraphic and sedimentologic features of the Autochthon have been interpreted to reflect the evolution of a continental margin (Stevens, 1970; Smyth, 1973; Williams and Stevens, 1974). According to this model, the basal clastic-volcanic unit relates to Hadrynian rifting with high relief and coarse clastic deposition and volcanism followed by lower relief, marine transgression, and deposition of mature beach sands in an intertidal environment in Early Cambrian time. The Bateau and Lighthouse Cove formations are overstepped westward by the Bradore Formation, which overlies unweathered gneiss of the Basement Gneiss Complex. The maturity of the sandstone therefore probably reflects prolonged transport and intertidal abrasion rather than a long period of weathering. A shoreline is further supported by the nature of underlying mafic volcanic rocks that are locally pillowed and subaqueous at Burnt Point but are subareal columnar flows nearby to the west at Cloud Hills.

The carbonate sequence with its shelly faunas and shallow water depositional characteristics was thought to have built up as a shallow-water carbonate bank (Rodgers, 1968; Stevens, 1970; Swett and Smit, 1972). Shale units toward the top of the Table Head Formation and intraformational conglomerates such as those near the top of the sequence at Coles Pond suggest subsidence and instability of the bank. Further deepening resulted in the cessation of lime deposition and the formation of black pyritiferous shale.

The upper clastic unit contains greywacke with sedimentary structures typical of turbidites, and graptolitic shale that indicates deposition in deep water. As the upper clastic unit was derived from the east and northeast, and as it contains detritus from the Hare Bay Allochthon that later overran it, then it is clearly related to the emplacement of the Allochthon.

ALLOCHTHONOUS ROCKS

The six distinct rock groups within the Hare Bay Allochthon form one or more separate slices that occur together at consistent structural levels. Each slice or assemblage of slices bears the name of its comprising rock group or formation. The same stratigraphic names that define the contrasting rock units therefore suffice to identify structural slices or slice assemblages. This system of structural-stratigraphic nomenclature is consistent with that applied to the Humber Arm Allochthon in the Bay of Islands area (Williams, 1973; 1975).

No single and continous vertical section exhibits all six contrasting transported rock groups that make up the Allochthon, so that the order of structural stacking must be built up from observations throughout the Allochthon. One or more omissions in the stacking order is commonplace, but reversals in the order are unknown.

There is no obvious age pattern in the sequence of structural stacking within the Hare Bay Allochthon, as for example, the structurally highest slices containing the stratigraphically oldest rocks. However, facies and provenance considerations coupled with the order of structural stacking and the mode of assembly and emplacement of the Allochthon (Williams, 1975) all combine to indicate that successively higher slices originated successively farther east at deposition, so that the highest structural slice is the farthest travelled. As a broad generality (and excluding the Cape Onion slice) the structurally highest slices contain the most deformed rocks.

The Maiden Point and St. Anthony slices are by far the most areally extensive, the Maiden Point being 128 km in total length. However, both are exceptionally thin, less than 610 m structural thickness, compared to their lateral extent. North of Hare Bay the slices are subhorizontal, although all were affected by postemplacement deformation. South of Hare Bay, the Maiden Point slice assemblage has been domed and eroded to expose the authochthonous Goose Tickle and Table Head formations in the Whites Arm Window.

Northwest Arm slice

The Northwest Arm slice comprises sedimentary rocks of the Northwest Arm Formation. These rocks are of limited areal extent and occur only at the western margin of the Hare Bay Allochthon where now-widely-separated occurrences probably represent erosional remnants of a once more-continuous single slice.

The Northwest Arm Formation

The name Northwest Arm was first assigned to a sequence of distinct sedimentary rocks at Northwest Arm of Hare Bay by Cooper (1937) and it has subsequently been applied more regionally to either the same outcrops or nearby lithologic correlatives at Hare Bay and Pistolet Bay (Gillis, 1966; Tuke, 1968; Williams et al., 1973; Williams and Smyth, 1974). Copper (1937) originally viewed the rocks as authochthonous and interpreted them to occupy a stratigraphic position above Middle Ordovician carbonates of the Table Head Formation and below Middle Ordovician autochthonous sediments of the Goose Tickle Formation. Tuke (1968) collected the graptolite *Staurograptus dichotomus* from the Northwest Arm Formation in Pistolet Bay, thus defining its Early Ordovician age. The formation is therefore transported and must structurally overlie the Goose Tickle Formation where the two are in contact.

Cooper (1937) estimated the thickness of the Northwest Arm Formation at 152 m; Tuke (1968) calculated it to be 65 m. Its stratigraphic top and bottom are not exposed in the Hare Bay area and the chaotic internal structure of the rocks precludes accurate thickness estimates.

The Northwest Arm Formation consists of finely bedded black and green shale, in most places graphitic and pyritic, with interbeds of grey- and buff-weathering limy siltstone, grey to brown sandstone, white to grey limestone, limestone breccia, and buff to dark grey and green chert and siliceous argillite. The rocks are almost everywhere chaotic with the shale beds forming a smeared and streaky matrix around boudins or blocks of the more resistant rock types (Fig. 86). The chaotic nature of the formation presents problems in field mapping for it is locally difficult to distinguish the Northwest Arm Formation per se from mélanges that occur between higher structural slices and that have Northwest Arm shale as their matrix. The shaly mélanges are distinguished chiefly by the presence of mafic volcanic or other exotic blocks that are unknown in the Northwest Arm Formation.

The contact between the Northwest Arm Formation and underlying autochthonous rocks of the Goose Tickle Formation is best exposed on the west side of Triangle Point in Pistolet Bay and at several localities along the north shore of Northwest Arm in Hare Bay. In all localities, well-bedded greywacke of the Goose Tickle Formation is in fairly abrupt, although locally irregular contact with chaotic dark shaly rocks that contain sandstone and limestone blocks. The contact is nowhere a planar hard tectonic thrust; rather it is an uneven surface where chaotic Northwest Arm rubbly shale overlies bedded greywacke.

Figure 86. Sedimentary blocks in chaotic black and green shale of the Northwest Arm Formation, south shore of Pistolet Bay. (GSC 202609).

The chaotic nature of the Northwest Arm Formation indicates emplacement in a semiconsolidated condition and transport by gravity sliding. Locally it contains sandstone blocks that resemble sandstone in the underlying Goose Tickle Formation. These were probably ripped up and incorporated within the overlying slide mass during transport.

The Lower Ordovician Northwest Arm Formation in the Hare Bay map area is the correlative of such lithologically contrasting sequences as the St. George Group and the Cape Onion Formation. South of the map area, lithologically identical correlatives are the transported Green Point Formation at Green Point and Port au Port Peninsula, and the transported Middle Arm Point Formation of the Humber Arm Supergroup at Humber Arm.

Maiden Point slice assemblage

Two separate slices are recognized in the Maiden Point slice assemblage. The larger is the most extensive slice within the Hare Bay Allochthon. A much smaller slice locally overlies the larger one at Windy Point (Croque Head slice of Smyth, 1973) and the two are separated by a black shaly mélange zone. Small areas of mélange locally separate continuous sections of the Maiden Point Formation in the larger slice at St. Carols, Deep Bay, and Little Canada Harbour. These mélange exposures are interpreted as the mélange beneath the main Maiden Point slice that is exposed in either tight anticlinal fold cores or brought up along late high-angle faults.

The main slice of the Maiden Point assemblage overlies the autochthonous Goose Tickle Formation in most places, with black shaly mélange present at the base of the slice. In a few places, for example, in the vicinity of Howe Harbour and Northwest Arm on the north side of Hare Bay, and at Indre Point on the south side of the bay, it overlies the Northwest Arm slice. Mélange is locally absent where the Maiden Point overlies the Goose Tickle Formation at Croque and at Canada Bay, and metamorphic rocks (Sugarloaf Schist Member of Goose Tickle Formation) form a distinct unit at the base of the Maiden Point slice assemblage south of Canada Bay.

The Maiden Point slice assemblage consists entirely of rocks assigned to the Maiden Point Formation, which is composed mainly of coarse greywacke, slate, pebble conglomerate, mafic volcanic rocks, and in addition includes a diorite-gabbro intrusive member. Limestone conglomerate and a thin marble unit are associated locally with the Maiden Point volcanic rocks at Croque Harbour. The Maiden Point Formation, although thick and the most areally extensive of the transported rocks, has defied all attempts at stratigraphic subdivision because of a lack of continuous marker horizons combined with complex internal structure.

Maiden Point Formation

The name Maiden Point 'Sandstone' was first used by Cooper (1937) to designate a sequence of clastic sedimentary rocks that are well exposed on the south shore of Hare Bay. Tuke (1968) proposed the term Maiden Point Formation for the rocks north of Hare Bay because of the wide variety of rock types included in the lithologic map unit. In the Canada Bay area, Betz (1939) proposed the name Canada Head Formation for identical greywacke and slate. These are clearly correlatives of the Maiden Point Formation to the north, and as the rocks occur in the same structural slice, the name Canada Head Formation has been dropped. Cooper (1937), Gillis (1966), and Tuke (1968) all included volcanic rocks within the Maiden Point Formation. Some of these are interlayered with the Maiden Point greywacke and slate, for example at St. Lunaire, Sacred Bay, Milan Arm, west of White Hills, and south of Great Islets Harbour; others at Ireland Point, Milan Arm, and Cape Onion clearly belong to structurally higher slices and they are redefined.

Diorite-gabbro sills and plugs that are spatially associated with sedimentary and volcanic rocks of the Maiden Point Formation in the same structural slice are all considered part of the Maiden Point Formation. These rocks are most prominent in places where the Maiden Point slice is in contact with structurally higher slices, as earlier noted by Cooper (1937). The intrusive rocks can be used to define the Maiden Point slice assemblage in places where other lithologies are absent. For this reason the intrusive rocks are included as part of the Maiden Point Formation.

Local occurrences of serpentinized ultramafic rocks near Grandois are not included in the Maiden Point Formation although their spatial distribution suggests a relationship to Maiden Point mafic volcanic rocks. The ultramafic rocks may represent unrelated detached blocks.

Mafic volcanic rocks, chiefly fragmental, with chert and slate interbeds along the south shore of Milan Arm are locally included in the Maiden Point Formation, whereas nearby occurrences of similar rocks to the west are depicted as blocks within the Milan Arm Mélange. If these rocks are integral parts of, or derived from, the Maiden Point Formation, then the abundance of chert in the local exposures is atypical of the formation elsewhere.

Cooper (1937) chose the well-exposed section of sedimentary rocks along the south shore of Hare Bay as the type section and estimated the thickness of the formation at 1830 m. Betz (1939) estimated the same thickness for correlatives at Canada Bay. Tuke (1968) estimated 150 m of volcanic rocks near the base of the formation and a total thickness of 2000 m. Recumbent folding makes thickness estimates hazardous, but the areal extent of the formation implies a thickness in the order of several kilometres.

Monotonous greywacke and slate make up about 90 per cent of the Maiden Point Formation. The greywacke is typically dark grey and weathers light grey to buff. Graded beds from 30 cm to 1 m thick are common in most sections. These rocks comprise mainly grey to blue quartz (60%) and plagioclase feldspar (25%) with minor rock fragments, muscovite, and clay minerals. Slump structures and sole markings are rare. Black to grey and red and green slate is interbedded with the greywacke, notably at St. Anthony and south of St. Julien Island. These are thinly laminated rocks in units up to 9 m thick. Quartz pebble conglomerate is present in the type section and is common in most places on the north side of Hare Bay. Quartz pebble conglomerate is less conspicuous south of Grandois, and at Canada Bay the greywacke is finer and more thinly bedded.

Greywacke of the Maiden Point Formation in the separate small slice from St. Juliens to Windy Point is finer grained and more thinly bedded than elsewhere. These rocks also exhibit a more complex internal structure and in most places have been converted to a semischist. Smyth (1971) referred them to the Croque Head Formation, but that name has now been dropped (Smyth, 1973; Williams and Smyth, 1974; Williams, 1975).

A 6-m-thick conglomerate unit that has boulders and cobbles of basalt and marble occurs within mafic volcanic rocks of the Maiden Point Formation at the head of Croque Harbour. Nearby at Croque village a 9-m marble unit occurs at the steep contact between Maiden Point volcanics and the autochthonous Goose Tickle Formation. The Croque Harbour conglomerate is intraformational to the Maiden Point valcanic rocks, and the marble unit is included within the Maiden Point slice as this lithology is unknown at the top of the Goose Tickle Formation. The contacts between the Maiden Point slice and the Goose Tickle Formation are poorly defined at Croque and mélange is absent.

Altered mafic agglomerate, tuff, and *basaltic lava* that occur within the Maiden Point Formation form discontinuous horizons up to a kilometre thick. These are most common in the north and St. Lunaire and thinner units occur along the east side of Whites Arm Window where they are at or near the structural base of the Maiden Point slice. Other smaller occurrences at Howe Harbour, Coles Pond, a large pond west of Great Islets Harbour, and Little Cormorandier Island are also at or near the base of the main structural slice. Agglomerate predominates in the thickest sections at St. Lunaire and north of Howe Harbour whereas altered basaltic lavas are more common south of Hare Bay. The lavas are green to grey and purple, fine grained, and locally amygdaloidal. Pillow lava occurs locally at St. Lunaire, and at Little Cormorandier Island limestone fills pillow interstices.

Petrochemical studies indicate that the Maiden Point volcanics are transitional between tholeiitic and alkali basalts (Smyth, 1973), and while showing definite chemical similarities to the Lighthouse Cove Formation (Strong and Williams, 1972), they are chemically unlike the oceanic tholeiites of the Lushs Bight Group of central Newfoundland (Smitheringale, 1972).

Medium- to coarse-grained, massive *gabbro* and *diorite* of the Maiden Point Formation form prominent intrusions near St. Anthony and inland east of Raleigh. Dykes and sills of similar rocks are common at the structural top of the Maiden Point slice along the north side of Hare Bay and a few occurrences are known on the south side of the bay. The intrusions are pretectonic with respect to the steep prominent cleavage in bordering greywacke, and in places along the north shore of Hare Bay the intrusions are structurally overlain by higher slices of the Hare Bay Allochthon. The common occurrence of these intrusions at the structural contact between the Maiden Point and St. Anthony slices suggested to Cooper (1937) that the bodies were emplaced during thrusting. Possibly they are related to volcanism within the Maiden Point Formation and now mark a competent structural level.

The Maiden Point Formation was involved in two deformational episodes that varied in intensity from place to place so that the rocks exhibit marked differences in outward appearance and internal fabric. The first deformation is evidenced by recumbent folds with subhorizontal cleavage. Excellent examples are apparent at St. Anthony Bight (Fig. 87, 88) and Croque Harbour, and their presence is indicated in other places by bedding-cleavage relationships. South of Hare Bay the early recumbent folds are upward-facing toward the northwest whereas north of Hare Bay the best examples at St. Anthony Bight face slightly downward to the southwest. These early folds are clearly the result of penetrative deformation in well-indurated rocks and their presence in the thin, subhorizontal Maiden Point slices indicates that the rocks were emplaced as already deformed, rigid and hard slices. The second deformation is evidenced by upright folds with steeply southeast-dipping cleavage (Fig. 89). These folds affect autochthonous and allochthonous rocks alike so that they postdate the emplacement of the Hare Bay Allochthon. Excellent examples are evident along the type section. East of St. Anthony Bight, the subhorizontal early cleavage associated with recumbent structures is folded about northeast-trending axes and steep axial surfaces (Fig. 88, 90).

Locally where extensively cleaved, the Maiden Point greywacke is a semischist with reduced grain size and the clastic texture is partially obliterated by a composite tectonic fabric. At Cat Cove and vicinity, at the south end of the Allochthon, the Maiden Point sediments are converted to hornfels that has well-developed biotite, cordierite, and andalusite porphyroblasts. These postdate the latest steep cleavage and apparently reflect the presence of a nearby intrusion.

The age of the Maiden Point Formation is uncertain but indirect geological evidence suggests an Early Cambrian or Hadrynian age. A preponderance of blue quartz indicates a crystalline source and a westerly derivation from nearby

Grenvillian rocks before the establishment of the Cambrian carbonate cover sequence (Stevens, 1970; Williams, 1971). A direct correlative of the Maiden Point in western Newfoundland is the Summerside Formation of the Humber Arm Supergroup at Bay of Islands (Stevens, 1970; Williams, 1973).

Thick clastic sequences like the Maiden Point Formation that occur along the western side of the entire Appalachian System are interpreted as continental margin deposits associated with rifting and opening of the proto-Atlantic Ocean (Williams and Stevens, 1974). The Maiden Point volcanics are probably the result of the same rifting process, and this interpretation is enhanced by their petrochemical affinities to the Hadrynian Lighthouse Cove Formation, which together with an associated thick clastic sequence, is interpreted as a continental margin tholeiite suite (Strong and Williams, 1972).

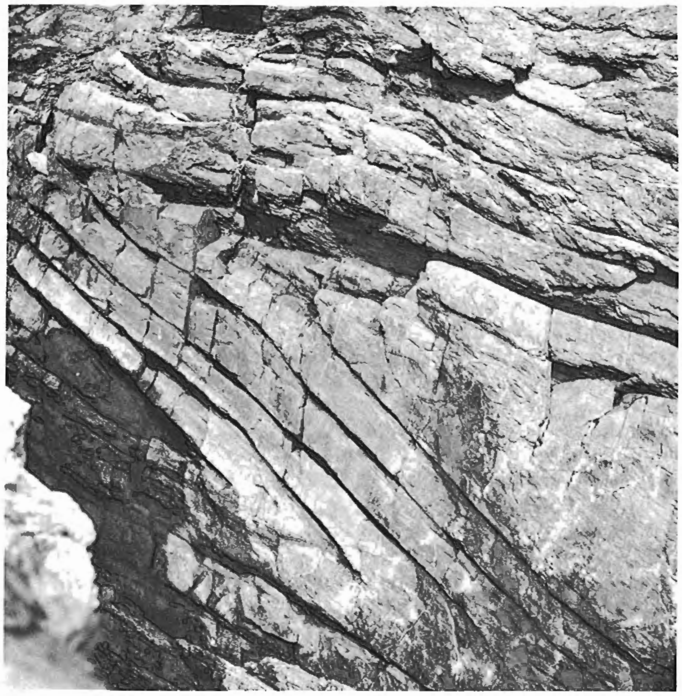


Figure 87. Synemplacement folds in Maiden Point Formation, Swoilers Cove, Cape St. Anthony. (GSC 202609-G).

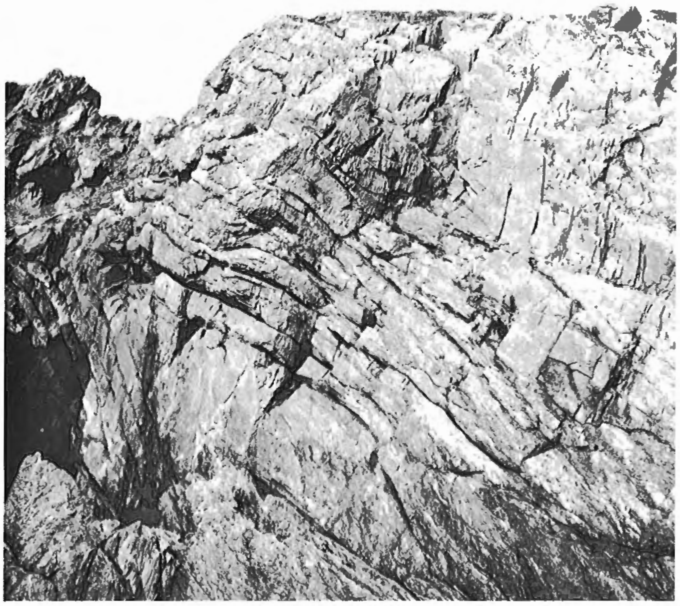


Figure 88. Synemplacement folds in the Maiden Point Formation, Swoilers Cove, Cape St. Anthony. A younger cleavage appears best near the top of the outcrop. (GSC 202609-H).

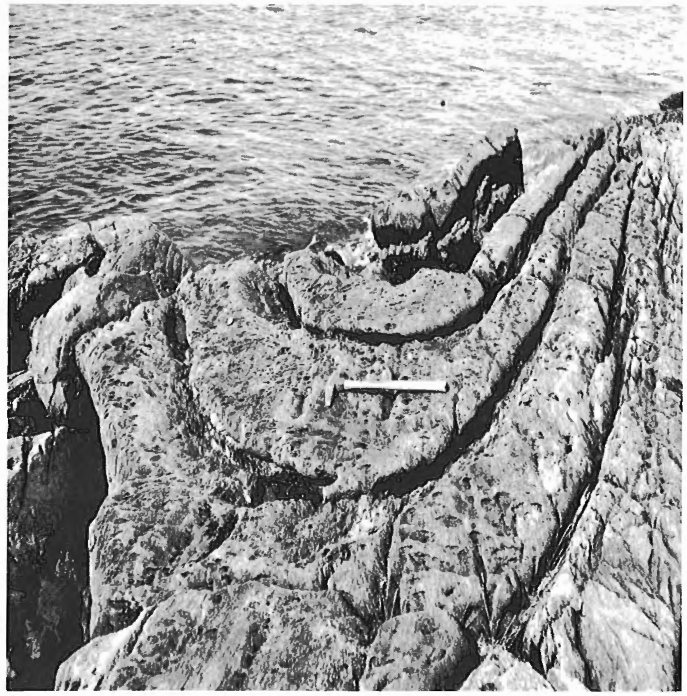


Figure 89. Northeast-plunging upright postemplacement fold in Maiden Point Formation, near Maiden Point, Hare Bay. (GSC 202609-E).

Figure 90. Cleavage associated with synemplacement folds in the Maiden Point Formation folded by postemplacement upright folds with steep axial plane cleavage, Swoilers Cove, Cape St. Anthony. (GSC 202609-F).

Grandois slice

The Grandois slice outcrops only on St. Julien Island and nearby Black Island, and it is areally the smallest slice of the Hare Bay Allochthon. It overlies the Croque Head slice of the Maiden Point slice assemblage and its basal contact is marked by a mélange zone that is well exposed at the south end of St. Julien Island. The top of the slice is not exposed but it presumably underlies the St. Anthony slice to the east.

The Grandois slice comprises the Grandois Group. These rocks were at first thought to form part of the uppermost slice of the Hare Bay Allochthon during preliminary studies (Williams et al., 1973); however, they are unknown elsewhere within the Allochthon and they are atypical of the St. Anthony Complex. Smyth (1973) and Williams and Smyth (1974) first used the name St. Julien Island slice for this distinct and separate rock group, but since the rocks are now called the Grandois Group, the slice name is changed in accord with the scheme of nomenclature previously outlined for the slices of the Hare Bay Allochthon.

The Grandois Group

The Grandois Group contains two formations, the Irish Formation to the west and the St. Julien Island Formation to the east. The formations are juxtaposed on St. Julien Island by a high-angle, northeast-trending fault. The relative ages of the two formations are unknown.

St. Julien Island Formation *

The St. Julien Island Formation (Smyth, 1971) consists of polymictic conglomerate with interbeds of red to green and purple coarse greywacke. It outcrops on the eastern side of St. Julien Island and on Black Island but neither top nor bottom is exposed. Its minimum exposed thickness at St. Julien Island is 60 m (Smyth, 1973). Pretectonic small gabbro sills and irregular intrusions cut the conglomerate both at Black Island and at the northwest end of St. Julien Island.

Cobble conglomerate is the predominant lithology although the beds are poorly sorted and clasts range from pebbles to boulder size. Thick beds without apparent grading are typical. Large scale trough crossbeds are evident locally. Some of the larger clasts are well rounded but most have been affected by tectonic flattening. The clasts comprise red quartzite (40%), vein quartz (30%), acidic volcanic rocks (15%), granite (5%), jasper (2%), and less commonly shale, mafic volcanics, sericitic phyllite, mafic tuff, gneiss, epidote, and quartz-chlorite schist.

The St. Julien Island Formation is strongly deformed and the conglomerate matrix and interbedded greywacke are converted to foliated quartz-sericite-chlorite semischist. The folilation surrounds augenshaped clasts that are flattened and elongated in the plane of the foliation. The degree of flattening varies with the competency of the clasts.

The wide variety of clasts in the St. Julien Island conglomerate implies a provenance other than the stable platform toward the west. Rather, the clastics must have been derived from a nearby volcanic-plutonic terrane that included areas of sedimentary rock. Possibly the conglomerate was derived from a volcanic milieu that lay to the east of the ancient continental margin.

The St. Julien Formation is now interpreted as Silurian and is therefore no longer viewed as an integral part of the Hare Bay Allochthon)

Irish Formation

Sandy limestone that outcrops along the western side of St. Julien Island was originally named the Irish Limestone (Smyth, 1971). The sequence consists of three units, each of which comes in contact with mélange at the base of the Grandois slice. The basal unit consists of sheared and brecciated quartzite with minor interbedded greenschist. It is overlain with gradational contact by brown thinly bedded siliceous limestone and slate. These in turn pass upwards into lighter coloured and thicker bedded sandy limestone. The limestone is brecciated in places near the contact with underlying mélange and locally the breccia is mixed with mélange. Both the mélange and breccia are involved in upright folds that postdate slice emplacement.

The sandy limestone contains up to 30 per cent terrigeneous material that is mainly quartz (70%), feldspar (30%), detrital muscovite, and accessory minerals. The clastic grains show a gradation from medium-sand size at the base of the beds to fine-sand size at the tops of beds. The quartz and feldspar grains are subangular and the clastic limestone is immature. Possibly they represent distal turbidites from a carbonate-rich source (Smyth, 1973).

The age of the Irish Formation is unknown. The most reasonable source of carbonate detritus is the autochthonous platformal sequence that evolved farther west. The admixture of carbonate and sand detritus suggested to Smyth (1973) that the formation may be equivalent to the Cambrian impure limestone and shale of the Forteau, and Cloud Rapids and Treytown Pond (Eddies Cove) formations of the Autochthon (Betz, 1939).

Milan Arm Mélange

The Milan Arm Mélange, although formally named herein and given the status of a separate structural slice, is in most respects similar to the mélanges that separate all structural slices of the Hare Bay Allochthon. However, it is areally extensive, it occupies a distinct structural postion, and its blocks or knockers are so huge as to be easily resolvable at even the present scale of mapping. In that it is impractical to name each of the large mappable blocks within the mélange, a single name is given to designate all of these rocks (Williams, 1975).

The Milan Arm Mélange outcrops on the north and south shores of Milan Arm and it extends north along the east side of Pistolet Bay to Carpon Cove. It is interpreted to structurally overlie the Maiden Point slice assemblage to the east and structurally overlies the autochthonous Goose Tickle Formation along its western margin on the shore of Pistolet Bay. At Croix Point on the south side of Pistolet Bay, the mélange is overlain structurally by the topographically higher St. Anthony slice assemblage. Mélange that is included as part of the Milan Arm continues from Carpon Cove to West Road of Sacred Bay, where it separates the Cape Onion and Maiden Point slices. The topography at West Road does not clearly reflect the structural stacking order, but the mélange there is interpreted to structurally overlie the Maiden Point slice and to lie beneath the Cape Onion slice. Alternately the Cape Onion slice may be viewed as an exceptionally large and continuous exotic at the same structural level as other exotics within the mélange farther south.

The Milan Arm Mélange consists of various sized blocks and slabs, largely of plutonic and volcanic rocks, that are surrounded by a black and green shale matrix. It has a distinct topographic expression with the knockers forming sharp hills, smaller knolls, and offshore islands. The matrix shale is exposed only on the coast and occupies low bush-covered ground inland. The chaotic nature of the rocks at Milan Arm was not previously recognized. Prior to the concept of Taconic klippen in western Newfoundland, the largest exotic blocks were mapped either as separate intrusions or as integral parts of the Maiden Point Formation (Cooper, 1937). Following the recognition of the Hare Bay Allochthon, some rocks of the Milan Arm Mélange were mapped as part of the allochthonous Maiden Point Formation (Gillis, 1966; Tuke, 1968) and the ultramafic rocks, where separated, were thought to be part of the St. Anthony slice (Williams et al., 1973).

*See footnote on p. 2

The most common blocks within the Milan Arm Mélange are serpentinized peridotite, mafic volcanic rocks, amphibolite and foliated gabbro, greywacke, diorite, and exceptionally coarse grained pyroxenite and hornblendite that are associated with tonalite and hornblende-biotite schist. Most of these rocks can be matched directly to the same lithologies represented in other nearby structural slices. A few of the lithologies are unique to the mélange. The largest single exotic is a serpentinized peridotite slab at Milan Point that is at least 1 km wide. It structurally overlies amphibolilte east of Milan Point and these and other superimposed pieces are surrounded by matrix shale.

The ultramafic rocks have a foliation and they are similar to the White Hills Peridotite. One and a half kilometres north of Milan Point, a large ultramafic exotic contains finely laminated garnetiferous hard, thin, recrystallized layers that are comparable to banded rocks found in western Newfoundland at the contacts between transported ophiolites and their underlying metamorphic aureoles (Williams and Smyth, 1973). The Milan Point exotic is therefore a sampling of the basal part of an ophiolite and its contact metamorphic zone.

Mafic volcanic rocks are for the most part similar to those of the Cape Onion Formation. Some are highly altered and near Carpon Cove they are carbonatized and brown weathering. Amphibolite exotics are in some places similar to amphibolite of the St. Anthony Complex but others are foliated gabbro without counterparts in the Hare Bay Allochthon. Greywacke blocks are typical of the Maiden Point Formation, and at least some of the diorite blocks may be derived from the same formation.

Coarse-grained pyroxenite and hornblendite that occur on both sides of Milan Arm have no correlatives elsewhere in the allochthon. A single exotic that forms an island southeast of Milan Point has bands and lenses of pyroxenite, hornblendite, and foliated coarse-grained tonalite within hornblende schist and hornblende-biotite schist. Some of the rocks are exceptionally coarse grained. Single tabular pyroxene crystals up to 15 cm in length are locally common and hornblende prisms up to many centimetres in length occur in both the schist and massive hornblendite. Similar coarse-grained hornblendite occurs in the mélange on the south side of Milan Arm directly south of Micmac Island.

Many of the amphibolite and diorite exotics along the north shore of Milan Arm are encased in a relatively thin hard rind of light grey calc-silicate alteration products (rodingite). The tough alteration halos presently form wave washed outcrop surfaces and the surrounding matrix is not evident. In a few examples, the rodingite alteration halos are surrounded, in turn, by thinner serpentinite coatings, implying that the rodingite represents an alteration zone where amphibolite or diorite was earlier surrounded by serpentinite. The exotics may have once been inclusions in a serpentinized ultramafic intrusion. More likely, they were inclusions in a serpentinite mélange that disintegrated so that the resistant exotics are recycled where they now occur in the shaly matrix of the Milan Arm Mélange.

The Milan Arm Mélange occurs at the western leading edge of the Hare Bay Allochthon in a zone where the higher structural slices have a tendency to disrupt, repeat, and disintegrate. The complex geological relationships west of Howe Harbour reflect break-up in this same structural zone.

In the Humber Arm Allochthon, the Little Port slice assemblage occurs in a similar westerly position and has several features reminiscent of the Milan Arm Mélange. The Little Port Complex occurs in a number of structural slices that are locally separated by a thick serpentinite mélange west of Lark Harbour (Comeau, 1972; Williams, 1973). Slices and blocks of the Little Port Complex are in places separated by steep zones of both serpentinite and shaly mélange where the larger slices are disrupted and disintegrated at Little Port. The Little Port Complex includes a variety of amphibolite like that which occurs as exotic blocks at Milan Arm; on the north side of the harbour at Trout River, the Little Port amphibolite is rodingitized where it is in contact with serpentinized peridotite.

There is no fundamental difference between tabular exotics within the Milan Arm Mélange, the successive slices that make up the Little Port slice assemblage, and the largest slices that comprise the Hare Bay and Humber Arm allochthons. All are contained in the same black and green shale matrix. The only difference is that of scale and the relative paucity of shaly matrix material between the larger structural slices of the major allochthons.

Cape Onion slice

The Cape Onion slice comprises mainly mafic pillowed lava assigned to the Cape Onion Formation that extends northeastward from Raleigh to form the Cape Onion Peninsula. Similar rocks occur offshore at Sacred Islands and a smaller isolated mass occurs at the village of Raleigh. In most places the Cape Onion slice overlies the Maiden Point slice. The clearest relationships are seen along the west side of Great Sacred Island and along the southern side of Little Sacred Island. A coarse mélange at Onion Cove underlies the volcanic rocks and the mélange contains large blocks of Maiden Point greywacke, suggesting that the Maiden Point Formation occurs in subsurface. One and a half kilometres north of Raleigh on the east side of Ha Ha Bay, the Cape Onion slice overlies deformed greywacke that is assigned to the Maiden Point Formation. The small isolated mass at Raleigh directly overlies the Goose Tickle Formation. The contact there is a subhorizontal thrust without mélange.

In the vicinity of Howe Harbour, mafic pillowed lava occurs at Lock's Island and near Roland Point. At Lock's Island, the pillowed lava is dark green to black with limestone filling the pillow intersticies. The volcanics structurally overlie the Table Head Formation and they are overlain by the St. Anthony slice assemblage. At Roland Point, black amygdaloidal (calcite) pillowed lavas structurally overlie the Maiden Point slice and they are overlain by the St. Anthony slice assemblage. Both the Lock's Island and Roland Point volcanic rocks have been assigned to the Cape Onion Formation. Another occurrence of mafic volcanic rocks structurally between the Northwest Arm and St. Anthony slices on the west side of Howe Harbour is also tentatively assigned to the Cape Onion Formation.

Previous workers (Cooper, 1937; Gillis, 1966; Tuke, 1968) included the Cape Onion volcanics with the Maiden Point Formation. However the assemblage is distinctive and the abundant pillowed lavas contrast with the predominant pyroclastic volcanics of the Maiden Point Formation. In addition, no greywacke of Maiden Point type occurs in the Cape Onion slice.

Cape Onion Formation

The Cape Onion Formation consists of black to green basaltic pillowed lavas with local agglomerate and tuff units and minor black pyritic shale. The most continuous section is exposed along the east side of Ha Ha Bay and impressive exposures of continuous pillowed lavas occur on Great Sacred Island and Little Sacred Island. The formation is probably a kilometre or more thick.

Where agglomerate, tuff, and shale are interlayered with the mafic lava at Ha Ha Bay, the beds strike northwesterly and dip moderately toward the northeast. A similar attitude on a thin tuff horizon at Great Sacred Island suggests stratigraphic continuity, although at Onion Cove local attitudes are northeast-trending. A few diorite dykes cut the Cape Onion Formation. These are probably related to the mafic volcanism and are indigenous to the structural slice, for mafic dykes are absent in autochthonous rocks nearby. Carbonate is commonly associated with the volcanic rocks. It fills pillow interstices, occurs in amygdules, and forms the matrix of some fragmental volcanic rocks. Brown-weathering outcrops at Raleigh are carbonate-rich altered mafic volcanic rocks.

Black shale of the Cape Onion Formation occurs in thin units interlayered with the volcanic rocks. Most units are finely laminated, pyritic and graphitic; one horizon at Onion Cove bears graptolites.

Rocks of the Cape Onion Formation are relatively undeformed compared to the structurally underlying Maiden Point Formation and structurally overlying St. Anthony Complex. There is no indication of pre-emplacement penetrative deformation, and postemplacement deformation has affected the rocks only locally. Where the Cape Onion volcanics structurally overlie the Goose Tickle slate northeast of Raleigh, cleavage in the slate locally continues upward into the volcanic rocks, but in the small isolated sheet at Raleigh the cleavage does not appear to penetrate the overlying more competent volcanic rocks. At Onion Cove, graptolitic black graphitic shale interlayered with the volcanic rocks are essentially uncleaved whereas nearby mélange beneath the Cape Onion slice displays a northeast-striking and moderately southeast-dipping cleavage. Postemplacement deformation in the Cape Onion Formation is lacking also because the formation is at the western margin of the Hare Bay Allochthon, where the competent volcanic rocks are effectively outside the zone of intense postemplacement deformation.

Black shale interlayered with mafic volcanic rocks at Onion Cove contain the following Lower Ordovician (Tremadocian) graptolites (identification by B.D. Erdtmann, Indiana University at Fort Wayne, Report 1-69-BDE pers. comm., 1968):

Dictyonema flabelliforme cf. var. *flabelliforme* (Eichwald, 1840)
Dictyonema flabelliforme cf. var. *anglicum* Bulman, 1927
Dictyonema flabelliforme cf. var. *parabola*? Bulman, 1954
Staurograptus sp.
Anisograptus sp.

The fossiliferous rocks and associated volcanics appear to comprise a separate block, but the lithologies are typical of the formation nearby.

Volcanic rocks of the Cape Onion Formation, the Ireland Point Volcanics, and diorite-gabbro dykes of the Maiden Point Formation all appear to be closely related in petrographic and chemical characteristics. They are interpreted therefore as a single suite of hydrous alkali basalts, possibly once forming part of a seamount near a continental margin (Jamieson, 1977).

St. Anthony slice assemblage

The St. Anthony slice assemblage is the structurally highest of the Hare Bay Allochthon. Its comprising rocks are referred to the St. Anthony Complex that is further subdivided into four units. The structurally highest unit, the White Hills Peridotite, is typical of ultramafic rocks that occur at the stratigraphic base of ophiolite suites. The three underlying units are amphibolite (Green Ridge), greenschist (Goose Cove), and mafic volcanic rocks (Ireland Point) that form an aureole of decreasing metamorphic grade and structural complexity beneath the White Hills Peridotite. The St. Anthony slice assemblage was earlier referred to as the White Hills slice (Williams et al., 1973; Smyth, 1973; Williams and Smyth, 1974), but in accord with the nomenclature now adapted for structural slices and transported rock assemblages in western Newfoundland (Williams, 1973), and because the term 'White Hills' is also used to designate a formation within the slice, the name White Hills slice is dropped.

The rocks of the St. Anthony Complex have a complex structural history that predates slice emplacement. The structural contrast with underlying allochthonous or autochthonous rocks is everywhere pronounced, and locally east of Ireland's Bight and at Quirpon Island, boulders of foliated greenschist and volcanic rocks occur in black shale mélange beneath the St. Anthony slice assemblage.

The St. Anthony slice assemblage forms the prominent highland north of Hare Bay and smaller erosional slice remnants form Mount Mer at Howe Harbour and Galets Head at Quirpon Island. Other probable remmants of the slice north of Hare Bay occur 6 km east of Raleigh and north of Savage Cove near Cape Onion. South of Hare Bay, the St. Anthony slice assemblage constitutes Fishot Islands and nearby Great Verdon and Pigeon islands, and a small exposure of the same slice occurs east of Croque village. The St. Anthony slice assemblage is subhorizontal, or nearly so, in most places north of Hare Bay (Fig. 91). At Fishot Islands, it dips gently to moderately east and a regional aeromagnetic anomaly offshore suggests the presence of a steep peridotite sheet.

The St. Anthony slice assemblage overlies the Maiden Point slice assemblage in most places. Clear structural relationships are exposed along the north shore of Hare Bay from Starks Bight to Lock's Cove and south of the St. Anthony road from Pistolet Bay westward. Similarly, erosional remnants of the assemblage overlie the Maiden Point slice assemblage at Quirpon and 6 km east of Raleigh. South of Hare Bay, the St. Anthony assemblage overlies the Maiden Point assemblage at Fishot Islands and the outlier east of Croque village overlies the smaller Croque Head slice of the Maiden Point slice assemblage. West of the White Hills, the St. Anthony slice assemblage overlies the Goose Tickle Formation and at Howe Harbour it locally overlies the Northwest Arm Formation. The St. Anthony slice assemblage overlies the Milan Arm Mélange on the south shore of Pistolet Bay, and greenschist and amphibolite at Savage Cove indicate that it also lies upon the Cape Onion slice.

St. Anthony Complex

The St. Anthony Complex is a polygenetic assemblage of ultramafic rocks and mainly metavolcanic rocks. The ultramafic rocks are formally referred to the White Hills Peridotite and the underlying metavolcanic rocks are divided into three formations defined on metamorphic grade and structural style. The structurally lowest formation, the Ireland Point Volcanics, consists mainly of agglomerate and pillow lavas that are in most places cleaved and flattened but nowhere multideformed or highly metamorphosed. These rocks pass upwards into the polydeformed low-grade Goose Cove Schist that in turn grades upwards into the Green Ridge Amphibolite. The type section is north of Hare Bay where the formations form a metamorphic aureole beneath the White Hills Peridotite (Williams and Smyth, 1973). Metamorphic grade increases structurally upward toward the base of the peridotite and the transitions from relatively undeformed and metamorphosed mafic volcanics to greenschist and amphibolite are sufficiently distinct to allow separation of the Ireland Point, Goose Cove, and Green Ridge formations, respectively.

The concentric outcrop pattern of the Green Ridge Amphibolite and Goose Cove Schist around the peridotite at White Hills is a direct result of the present topography and subhorizontal attitude of the ultramafic rocks and underlying units within the same subhorizontal structural slice. Three

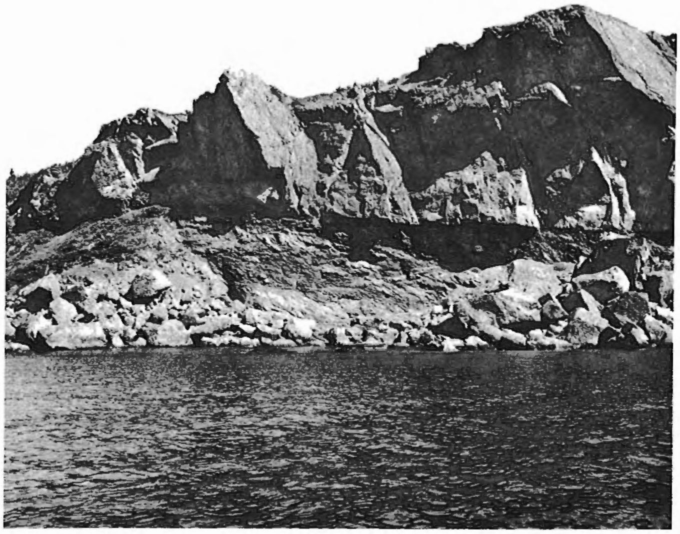


Figure 91. Subhorizontal tectonic contact between Ireland Point Volcanics of St. Anthony slice assemblage and underlying Maiden Point greywacke of Maiden Point Slice assemblage, Ireland's Point, Hare Bay. (GSC 202609-J).

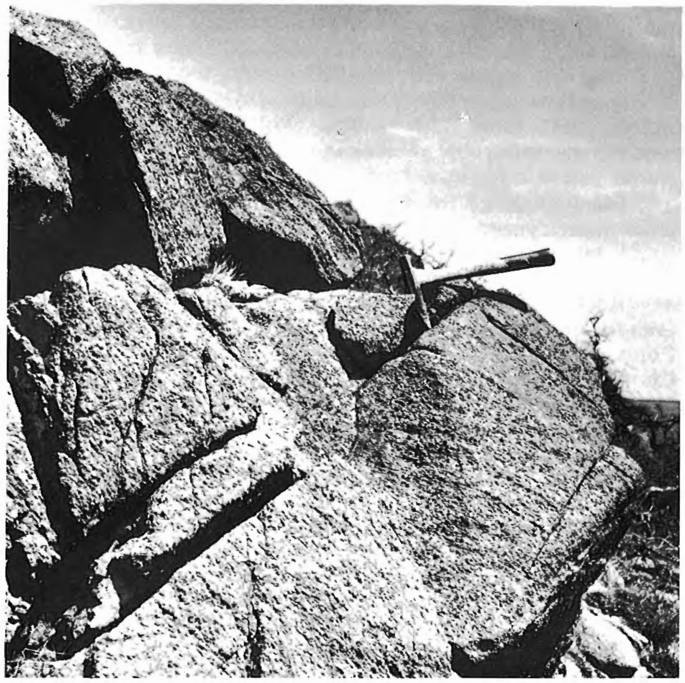


Figure 92. Foliated hartzburgite of the White Hills Peridotite, Mount Mer, Hare Bay. (GSC 202609-C).

kilometres west of St. Anthony, a small ultramafic outlier occurs within the Green Ridge Amphibolite and a topographic depression within the White Hills Peridotite nearby exposes an inlier of Green Ridge Amphibolite, thus confirming a subhorizontal attitude of the rock units. This geological setting at White Hills clearly indicates that greenschist and amphibolite remnants of the St. Anthony slice assemblage elsewhere in the map area were also once capped by peridotite that has been removed by erosion.

The structural base of the St. Anthony Complex is in most places coincident with the Goose Cove Schist but at Hare Bay it is below the relatively unmetamorphosed Ireland Point Volcanics and at Howe Harbour it transgresses upward to eliminate the aureole so that the White Hills Peridotite directly overlies lower structural slices. The relationship indicates that the aureole of the White Hills Peridotite bears no genetic relationship to the present tectonic base of the transported slice.

Ireland Point Volcanics

The term Ireland Point Volcanics was first used by Cooper (1937) to designate the mafic volcanic rocks on the north side of Hare Bay. The term was dropped by most subsequent workers and the volcanic rocks were included in the Maiden Point Formation (Gillis, 1966) or the Goose Cove Formation (Tuke, 1968). The volcanics occur at the base of the St. Anthony Complex along the north side of Hare Bay, where they structurally overlie the Maiden Point Formation. Cooper (1937) suggested a total thickness of 610 m but a lack of distinct layering makes estimates unreliable. If the thickness is about 1000 m, then the volcanic rocks must have moderate dips within the subhorizontal slice. This further implies that the subhorizontal formational boundaries defined on metamorphic grade sharply transgress the stratigraphic sequence.

Agglomerate is the most common rock type in the Ireland Point Formation. Undeformed green varieties with angular volcanic fragments from 2 to 15 cm or more in diameter are well exposed at Starks Bight. West of Ireland's Bight the agglomerate is green, red, and purple and consists of amygdaloidal red and green lava fragments in a green to grey matrix. In places the volcanic fragments are flattened and attenuated and the matrix is schistose. Pillow lava occurs locally in the agglomerate. The best preserved examples are along the west side of Starks Bight where limestone fills the interstices between bun-shaped pillows. East of Starks Bight, Ireland Point agglomerate becomes progressively more schistose and grades into the Goose Cove Schist.

Goose Cove Schist

The name Goose Cove Schist was used by Cooper (1937) to designate the greenschist that concentrically surrounds the White Hills Peridotite and that also occurs at Fishot Islands. Tuke (1968) and Smyth (1971) subsequently included all of the metamorphic rocks beneath the White Hills Peridotite at Hare Bay and correlatives at Fishot Islands within the Goose Cove Formation. The Goose Cove Schist as presently outlined is roughly coincident with the rock unit as originally defined by Cooper (1937). Also included in the formation are correlatives in outlying slice remnants at Quirpon, at the locality 6 km east of Raleigh, and near Cape Onion. The stratigraphic thickness of the formation is unknown but its structural thickness is estimated at 180 m in its type locality at Goose Cove. Smyth (1973) estimated 80 m of greenschist at Fishot Islands.

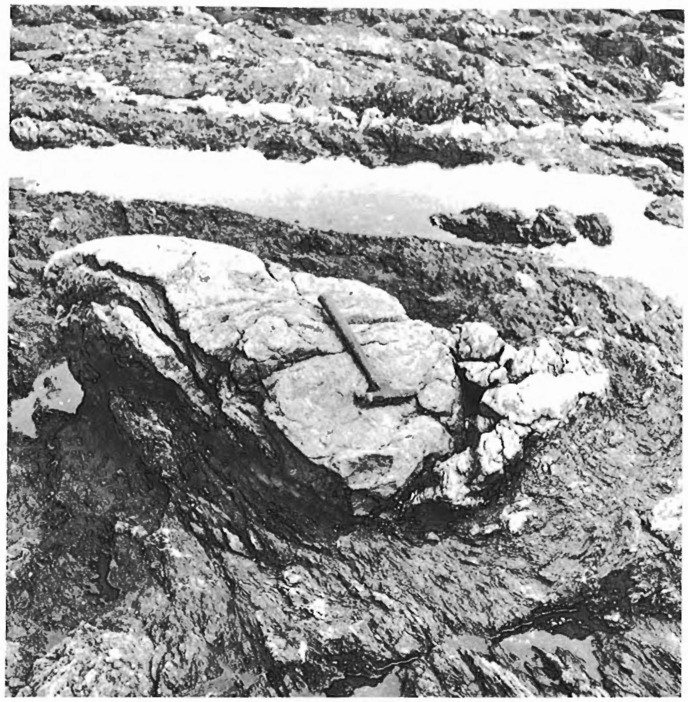


Figure 93. Block of Maiden Point greywacke within melange beneath the St. Anthony slice assemblage, west shore of Fishot Islands. (GSC 202609-A).

The Goose Cove Schist is a polydeformed and metamorphosed sequence of green tuff, agglomerate, and mafic pillow lava with thin units of greywacke, black pyritic slate, and limestone. Deformation is intense in certain zones within the formation so that the rocks vary from semischist to intensely foliated schist. Only in the least deformed areas are primary features of the rocks retained.

Finely laminated light to dark schist is predominant. The fine layering is thought to represent relict bedding in tuffaceous rocks although where deformation is intense it may reflect inhomogeneities within coarse pyroclastic rocks or flows. Graded beds less than 2 cm thick are recognizable within the finely layered rocks at Goose Cove South and nearby, confirming a tuffaceous origin. Three kilometres north of Goose Cove, the greenschist includes coarse tuffs and agglomerate, as clearly indicated by stretched volcanic fragments in a schistose greyish green matrix. Poorly preserved pillows are present in roadcuts north of Goose Cove, and nearby fine-grained massive altered rocks with plagioclase porphyroblasts or deformed phenocrysts are also probably flows.

At Goose Cove and Starks Bight, the greenschist has infolded black pyrite-bearing pelitic units, thin recrystallized limestone bands, and local psammitic units up to several tens of metres in structural thickness. The most prominent sedimentary unit within the formation is 10 m of greywacke that underlies the greenschist along the west side of Fishot Islands. It has beds about one metre thick at its base that are poorly graded and interbedded with red slate. Toward the top of the unit the greywacke beds are thinner and they are interbedded with purple and green agglomerate and tuff. Single greywacke beds up to 1 m thick occur interbedded with tuff and lava at Pigeon and Great Verdon islands, and schistose greywacke occurs within the Goose Cove Schist south of Milan Arm. Thin limestone beds, up to 30 cm thick, are interbedded with tuffs on the east side of Landing Cove at Fishot Islands and on the west side of Great Verdon Island. Hard, purple, well-foliated quartzite occupies a narrow zone of intense deformation bounded by less deformed mafic tuff at the southwest end of Fishot Islands.

Rocks of the Goose Cove Schist are characterized by greenschist metamorphic mineral assemblages that include chlorite, albite, epidote, muscovite, tremolite, and actinolite. The prismatic or platy metamorphic minerals have a strong preferred orientation that defines a prominent first schistosity (S_1). This schistosity is folded around recumbent second phase folds (F_2) that have large amplitudes compared to their wavelengths. Axial surfaces are gently dipping, and in a gross regional way, they parallel the S_1 schistosity and parallel the base of the White Hills Peridotite. Recumbent F_2 folds are well displayed at the coast south of Goose Cove and flat S_1 schistosity is prominent at Crémaillère Harbour and along the north side of Hare Bay to Lock's Cove. At Fishot Islands, second phase folds in the basal greywacke unit face west.

At Lock's Cove the greenschist is brecciated in local steep narrow zones that cross the subhorizontal schistosity at high angles. The brecciated rocks consist of schistose fragments in a coherent matrix of more finely broken schistose material. This brecciation is similar in aspect to that seen in igneous rocks at Bay of Islands (Williams and Malpas, 1972) although its origin is unknown.

The Goose Cove Schist is gradational over several tens of metres into the overlying Green Ridge Amphibolite by an increase in grain size of prismatic amphibole crystals, by a decrease in the amount of chlorite, and by the development of a second phase (S_2) hornblende fabric. The gradational relationship is well exposed and easily accessible east of the Goose Cove road on the flanks of Three Mountain Summits. It is also well exposed along the south side of Easter Tickle of Fishot Islands.

Green Ridge Amphibolite

The name Green Ridge Amphibolite is proposed here for the amphibolite and associated higher grade metamorphic rocks that occur structurally above the Goose Cove Schist and beneath the White Hills Peridotite. These rocks were not formally named by Cooper (1937) or Gillis (1966), and they

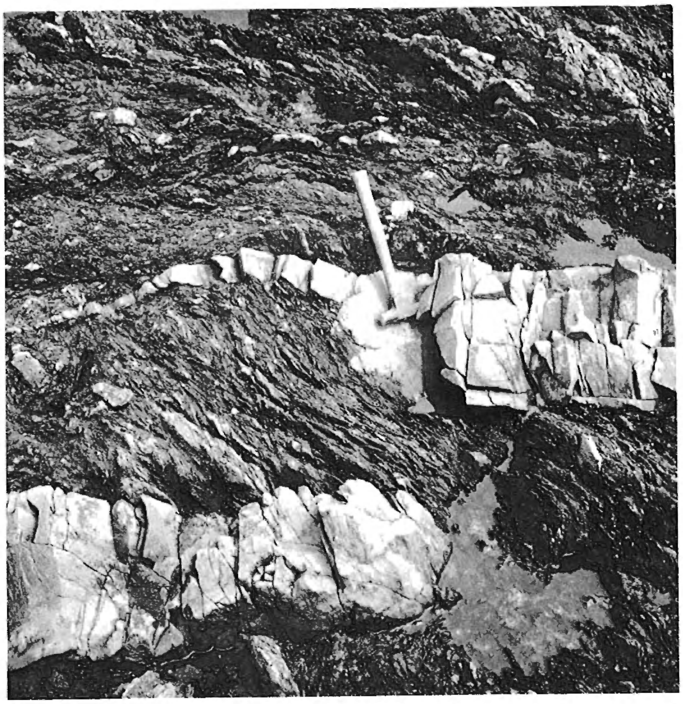


Figure 94. Late intermediate dykes cutting cleaved melange, west shore of Fishot Islands. (GSC 202609).

were included in the Goose Cove Formation by Tuke (1968). Smyth (1971, 1973) separated the amphibolite as a member of the Goose Cove Formation at Fishot Islands. The chief occurrence and type locality is at White Hills north of Hare Bay with smaller occurrences at Quirpon Island, at the locality 6 km east of Raleigh, at Cape Onion and on Fishot Islands. The structural thickness of the formation is estimated at 120 m.

Green to black hornblende-plagioclase schist, the most common lithology of the formation, is derived mainly from mafic volcanic rocks although no primary volcanic features are preserved. Some of the rocks are heterogeneous suggesting a coarse pyroclastic derivation and some banded varieties probably represent bedded tuffs. The hornblende-plagioclase schist is fine- to medium-grained, commonly with increasing grain size across the formation from structural base to top. Garnetiferous amphibolite is common toward the top of the formation.

Garnet-biotite schist occurs as thin local units; at Fishot Islands a few impure marble beds are interlayered with the hornblende schist. Metagabbro also occurs within the formation at Fishot Islands and near the top of the formation at White Hills.

The structural top of the Green Ridge Amphibolite is characterized by a 10- to 30-m-thick contact zone that can be defined in the rocks nearest the peridotite by the presence of deep brown hornblende and augite. From top to bottom across this zone, hornblende exhibits decreasing intensity of pleochroism from deep reddish brown to pale green and the amphibolite is in places banded. Rocks of the contact zone are also characterized by a strong schistosity with knots and augen of hornblende or pyroxene or both enclosed in a feldspar-rich matrix. The folilation parallels the base of the White Hills Peridotite and also the axial planes of recumbent second phase folds (F_2) that are well developed in underlying amphibolite. The fabric in the contact zone is interpreted as a second phase schistosity (S_2), since this schistosity is the dominant fabric of the Green Ridge Amphibolite.

White Hills Peridotite

The name White Hills Peridotite is applied to the ultramafic rocks that form the two tabular erosional remnants of a single peridotite sheet that correspond with the eastern and western White Hills. Two smaller bodies occur west of Howe Harbour. The southern occurrence is a continuation of the western White Hills ultramafic rocks and their aureole (Map 1495A, Section AB). The northern occurrence at Mount Mer is a separate mass that rests directly on lower sedimentary slices. The ultramafic rocks in the Milan Arm Mélange are also correlatives.

Cooper (1937) referred to the ultramafic rocks of the St. Anthony slice assemblage as the White Hills Peridotite Sheet and estimated their thickness at 610 m. Tuke (1968) suggested a thickness of 200 m and the accompanying map and cross sections indicate a similar value.

The White Hills Peridotite is brown weathering and devoid of vegetation. It consists of harzburgite, dunite, and minor pyroxenite. The harzburgite is medium- to coarse-grained and everywhere displays a foliation that is portrayed by flattened orthopyroxene crystals (Fig. 92). These deformed crystals are resistant and weather in high relief compared to surrounding partly serpentinized olivine, thus forming characteristic rough weathered surfaces. The dunites are finer and more even grained than the harzburgites and weather to a brownish yellow. The ultramafic rocks exhibit an irregular layering in some exposures and dunite and coarse pyroxenite form local layers in the harzburgite ranging from several centimetres to 2 m or more in thickness. The foliation and layering are parallel except in places where the layers are folded about tight isoclinal folds. Orthopyroxenite layers of two generations are contained in the harzburgite. The older is pretectonic and forms the most conspicuous primary layers. The younger is posttectonic and cross-cuts the layering and tectonic fabric in the harzburgite.

Cooper (1937) interpreted the White Hills ultramafic occurrences as two structural basins with the layering within the rocks being parallel to the basin floors. The present mapping does not suggest pronounced basinal structures for the sheet and attitudes of layering are irregular and bear no clear relationship to the shape of the peridotite sheet.

Finely banded, hard, recrystallized ultramafic rocks occur at the base of the White Hills Peridotite on the east side of Howe Harbour. The thinnest layers are lenticular and 0.3 cm thick and alternate with wider and more persistent layers up to 5 cm thick. Approximately 1.5 m of layered rocks are exposed, and their common occurrence as beach cobbles in this area suggests a greater extent. The rocks have a distinct foliation parallel to the layering that is portrayed by the parallel alignment of both metamorphic minerals and relict orthopyroxene. Leucocratic layers consist of deep brown amphibole, colourless clinopyroxene, brown biotite, and garnet and these bands alternate with darker serpentine-rich layers. The rocks were at first thought to be a part of the Green Ridge Amphibolite but the occurrence of attenuated relict orthopyroxenes with glide lamellae and relict chromite and ceylonite indicate that they are a metamorphosed basal part of the White Hills Peridotite. Similar rocks occur at the base of a small ultramafic outlier west of St. Anthony, and they are associated with ultramafic rocks in the Milan Arm Mélange on the east side of Pistolet Bay.

The base of the White Hills Peridotite is cataclastically deformed in the vicinity of Daniels Lookout. The tectonic fabric is brecciated and locally refolded on minor discontinuous flat-lying folds. The cataclastic effects die out upwards in about 15 m from the base of the peridotite sheet. This cataclasis is interpreted as a late detachment feature that postdated the main emplacment of the Allochthon.

Correlation, age, provenance, interpretation of St. Anthony Complex

The White Hills Peridotite is lithologically comparable to the basal unit of the Bay of Islands Complex (Williams, 1973) and both ultramafic assemblages overlie similar dynamothermal aureoles. The Bay of Islands Complex is a complete ophiolite suite and it has been interpreted as oceanic crust and mantle (Stevens, 1970; Church and Stevens, 1971; Dewey and Bird, 1971; Williams and Malpas, 1972). Accordingly, the White Hills Peridotite is interpreted as transported mantle. As in the case of the Bay of Islands Complex, the deformed and foliated ultramafic rocks are cut by pyroxenite dykes that are an integral part of the ophiolite milieu. The deformation in the ultramafic rocks therefore predated the final solidification of the complete ophiolite suite so that the deformed ultramafic rocks may be mantle tectonites. Their age is unknown, but if the Cape Onion volcanic rocks, now in a separate slice, were once the upper part of the same ophiolite suite, then the White Hills Peridotite was capped by Lower Ordovician volcanic rocks.

Finely banded, hard recrystallized rocks at the base of the White Hills Peridotite have analogues in the Bay of Islands Complex. All are interpreted as ultramafic rocks that originated as mantle tectonites and that were later mylonitized and recrystallized during transport (Williams and Smyth, 1973).

The Green Ridge Amphibolite and underlying formations within the St. Anthony Complex form a metamophic aureole beneath the White Hills Peridotite. Similar metamorphic rocks occur beneath the Bay of Islands Complex, and they are common associates of world-wide examples of transported ophiolite suites. Williams and Smyth (1973) suggested that the metamorphism and structural style of these rocks are the result of actual obduction of hot mantle material.

In the Hare Bay Allochthon, the downward gradation from the Green Ridge Amphibolite through the Goose Cove Schist to the Ireland Point Volcanics gives some idea of the rock types from which the metamorphic aureole was formed. Some indirect lines of geological reasoning and lithic comparisons suggest that the aureole protoliths are correlatives of the Maiden Point Formation. The greywacke at the base of the Goose Cove Schist along the western side of Fishot Islands is lithologically similar to semischist of the Maiden Point Formation. Likewise, the dominant pyroclastic lithologies of the Goose Cove and Ireland Point formations are more akin lithically to the type of volcanic rocks in the Maiden Point Formation than to any other nearby volcanic sequence. Cooper (1937) and Gillis (1966) suggested correlation of the Ireland Point Volcanics with the Maiden Point Formation and Smyth (1971, 1973) correlated the Fishot Islands greywacke with semischist of the Maiden Point Formation in the small slice near Croque.

The Fishot Islands greywacke, like the Maiden Point Formation, must have been derived from Grenville rocks to the west, and therefore was probably deposited at a continental margin during Early Cambrian or late Hadrynian. Similarly, the Ireland Point and Goose Cove volcanic rocks, if Maiden Point equivalents, may represent mafic volcanism at a distended and rifted margin.

Recent petrographic and chemical studies, which suggest correlation of the Ireland Point Volcanics with the Lower Ordovician Cape Onion volcanics (Jamieson, 1977), contradict the correlation of volcanic rocks of the St. Anthony Complex with volcanic rocks of the Maiden Point Formation. If the Fishot Islands greywackes of the St. Anthony Complex are indeed Maiden Point correlatives, then conceivably the St. Anthony Complex includes rocks of a wide age spectrum, all structurally telescoped and remoulded by later metamorphism related to ophiolite transport.

Williams and Smyth (1973) estimated that the White Hills Peridotite together with its attached aureole in the St. Anthony slice assemblage, moved westward for a minimum distance of 80 km.

Mélange zones

Mélange zones at the soles of the various structural slices within the Allochthon consist of chaotic blocks in a black, or black and green shaly matrix. In a few places, slice contacts are hard thrusts and mélange is sparse or absent, for example, the basal contact of the Maiden Point slice assemblage at Canada Bay and Croque, the basal contact of the Cape Onion Formation outlier at Raleigh, and the basal contact of the St. Anthony slice assemblage in some places above the Maiden Point slice assemblage. Mélange zones that appear to be within the Maiden Point slice assemblage occur at Little Canada Harbour, at Deep Bay on the south shore of Hare Bay, and at St. Carols. All of these occurrences are thought to represent mélange that is basal to the Maiden Point slice assemblage and either now exposed in elongate anticlinal cores or else brought up along steep faults.

The mélange zones vary in thickness from 1 to 30 m or more and their upper and lower contacts vary from sharp to gradational. They are all lithologically similar in that they contain isolated clasts in the same shaly matrix. In most places, the matrix shale is converted to slate or phyllite that contains secondary nodular concretions and euhedral crystal aggregates of pyrite. Clasts of Maiden Point greywacke (Fig. 93) are most common but mélanges above the Goose Tickle Formation contain abundant clasts of buff-weathering limy siltstone and sandstone, and mélanges above the Maiden Point slice assemblage contain abundant volcanic clasts. Diorite or gabbro blocks occur in the mélanges at Fishot Islands and beneath the Croque Head slice of the Maiden Point slice assemblage. Ultramafic blocks are unknown, although schistose volcanics from the St. Anthony slice assemblage occur in mélange on the north side of Hare Bay and at Quirpon Island.

The abundance of Maiden Point greywacke clasts is probably a reflection of both the widespread occurrence of this formation in the Allochthon and the fact that it forms a bounding surface to most mélanges. However, Maiden Point greywacke clasts are common even in places where the formation is absent, such as north of Hare Bay where the St. Anthony slice assemblage or Cape Onion slice directly overlies the Goose Tickle Formation, or at Savage Cove where the St. Anthony slice assemblage overlies the Cape Onion slice.

Clasts within the mélanges show a variation in size from pebbles to boulders and blocks more than 30 m across. All are chaotically intermixed without internal grading. Cobbles and boulders from 30 cm to a metre or more in diameter are the most common. These vary from angular to rounded with oval shapes being most common. Larger blocks tend to be tabular and angular, like the larger clasts in the Milan Arm Mélange. At St. Julien Island, thin limy siltstone slabs of Irish Formation up to 3 m long can be traced into broken angular blocks.

In most places north of Hare Bay and in the western part of Hare Bay, the mélanges possess a single steep cleavage. At Fishot Islands, southwest of Great Islets Harbour, and localities to the south, the mélanges exhibit two fabrics. The earlier is arranged parallel to the mélange contacts and is folded about upright folds with an associated steep crenulation cleavage. The intensity of the early fabric increases gradually from west to east across the southern part of the area and is most intense below the St. Anthony slice assemblage at Fishot Islands. The second phase folds and fabric are also most pronounced in the east. The early fabric in the mélange matrix at Big Spring Inlet is a fissility that surrounds augen-shaped clasts. Eastward, southwest of Great Islets Harbour in the same mélange zone below the Maiden Point slice assemblage, the early fabric is defined by flattened clasts and a slaty cleavage in the matrix. Farther east at Fishot Island, the early fabric is intense with quartz pebbles completely flattened into thin planar forms, and minor recumbent folds are associated with the first schistosity.

Any hypothesis that explains the mélange zones must consider the following points:

(1) Clasts can be matched directly to the Allochthon. Volcanic and diorite blocks are derived mainly from the Maiden Point Formation and buff-weathering limy siltstone and sandstone can be matched with lithologies of the allochthonous Northwest Arm Formation north of Hare Bay.

(2) All mélanges have the same black and green shale matrix that is identical to chaotic shale of the Northwest Arm Formation.

(3) The mélange zones display a similar structural thickness and uniformity regardless of structural position; there are no exceedingly thick zones.

(4) They bear a structural relationship to the bases of the overlying structural slices.

The nature of the clasts and matrix combined with the structural position of the mélanges beneath the transported slices, clearly indicate that their formation relates to the assembly and transport of the Hare Bay Allochthon. Exotic clasts within the mélanges are indigenous to structurally higher slices so that they are viewed as the products of surficial disruption and mass wastage, rather than deep seated tectonic mixing. Other clasts such as limy siltstone and sandstone are indigenous to the Northwest Arm Formation matrix and are mostly boudins and fragments of tectonically disrupted beds. Similarly, some Maiden Point clasts in mélanges above the Maiden Point slice assemblage must have been incorporated by structural overriding rather than by erosion.

Most of the transported rocks of the Hare Bay Allochthon are of the same age or older than the Lower Ordovician shale matrix of the Northwest Arm Formation. This formation cannot therefore represent a level of initial detachment, and this is further supported by the lack of coincidence between the present structural bases of the slices and stratigraphic levels within the slices. The Northwest Arm Formation therefore must represent an independent lithology across which each slice traversed during some stage of its transport. Structural analyses, coupled with the occurrence of large serpentinite and gabbro exotics in the structurally lowest mélanges of the Humber Arm Allochthon (Stevens and Williams, 1973) indicate progressive assembly of slices from east to west for the Newfoundland Allochthons. This mode of assembly and the presence of Northwest Arm lithologies in mélanges above the Maiden Point slice implies that the Northwest Arm Formation must have been deposited partly above and partly to the east of the Maiden Point Formation.

Structures of the St. Anthony slice assemblage clearly predate the assembly of the Allochthon and mélange formation, for schistose amphibolite occurs above much less metamorphosed and deformed mélange and schistose volcanic blocks occur in mélange. Structures of the Maiden Point slice assemblage are rather similar to those of the more deformed mélanges southwest of Great Islets Harbour and at Fishot Islands, so that its structures may relate to the assembly of the Allochthon. The Northwest Arm Formation in the lowest structural slice has the same structural style and chaotic nature as nearby mélange.

Most of the mélanges between the slices of the Hare Bay Allochthon contain well over 50 per cent shale matrix compared to clasts. The coarser mélanges with large tabular clasts contain much less matrix, and in the case of the spectacularly coarse Milan Arm Mélange, the percentage of shale matrix is well below 50 per cent. In fact some of the clasts in the latter example could be viewed as separate slices, and the Cape Onion slice may be a case in point. Because all of the slices of the Hare Bay Allochthon are separated by mélange, then the entire Allochthon can be viewed as a mega-mélange with very little matrix and exceptionally large tabular clasts.

Paleogeographic setting of the allochthonous rocks

The allochthonous rocks, like the underlying autochthonous rocks, are all interpreted to relate to the development of a continental margin that was eventually destroyed by the obduction of oceanic crust and mantle westward upon the continent. This model fits well with the lithologies represented within the Allochthon. Also, the facies and order of structural stacking indicate that the highest slices are the farthest travelled. The White Hills Peridotite is interpreted as mantle, and its underlying metamorphic rocks of the St. Anthony Complex, which now form an integral part of the same slice assemblage, are thought to represent supracrustal rocks that were deformed, metamorphosed, and structurally attached to the sole of the hot moving sheet at the time of its initial expulsion from an oceanic domain. The Cape Onion Formation probably originated in the same oceanic domain where at deposition it represented part of the upper volcanic layer of oceanic crust or a seamount at the continental margin. The Milan Arm Mélange, which occurs at the same high structural level as the Cape Onion slice, resulted from the mass break-up of an earlier intact slice whose rocks must have resembled those of the St. Anthony Complex. As well, it must have included tectonic serpentinite mélange as part of its structural milieu.

The mixed lithologies of the Grandois Group are more difficult to interpret. It is uncertain whether the conglomerates of the St. Julien Island Formation represent the detritus of a volcanic terrane at a rifted margin, or the detritus of an island arc that developed in response to compression and ocean closing. The limy sandstone of the Irish Formation may be offshore, deep water equivalents of early carbonate deposits along the continental shelf, or it may represent the local detritus of limestone reefs that surrounded volcanic islands.

The Maiden Point greywacke is the erosional product of a Precambrian metamorphic terrane. The age, thickness, and regional distribution of these kinds of rocks all along the zone of Grenville inliers in the western Appalachians strongly imply deposition in a continuous, though irregular belt that is most reasonably interpreted as a rifted continental margin (Williams and Stevens, 1974). Volcanic rocks and some intrusions within the Maiden Point Formation relate to rifting and the formation of such a margin (Smyth, 1973), whereas other intrusions within the Maiden Point Formation may relate to the construction of seamounts of the evolving continental margin (Jamieson, 1977).

The Northwest Arm Formation, in the lowest slice, lay to the east of the morphological shelf-edge and an evolving carbonate bank. There, it represented a shaly, deeper water, eastern facies of the Lower Ordovician carbonate that is so prominent and well developed along the western side of the Appalachian System.

The mélanges, which separate all of the slices both from one another and underlying autochthonous rocks, represent the mass wastage products of the slices that accumulated in marine muds like those of the Northwest Arm Formation. That formation is itself everywhere internally chaotic and in most places difficult to separate from mélange at the base of the Maiden Point slice assemblage. Mélanges with typical Northwest Arm Formation matrix above the Maiden Point slice assemblage indicate that the Lower Ordovician shale must have extended well to the east at deposition where it was a marine reservoir for olistostrome rubble that was continually overrun in the progressive westward assembly of the Allochthon.

POSTEMPLACEMENT INTRUSIONS

Intrusions that postdate the emplacement of the Hare Bay Allochthon consist of widely distributed dykes along the eastern shore of the Great Northern Peninsula. These cut allochthonous and autochthonous rocks alike, as well as mélange zones that separate the transported slices. Igneous intrusion was more widespread farther east where a large Devonian granite body constitutes most of Bell Island.

A few altered, green mafic dykes that cut the Table Head and Goose Tickle formations at Canada Harbour, Englee, southwest of Great Islets Harbour, and southwest of Indre Point are deformed by the earliest deformations in the surrounding rocks and are therefore earlier or synchronous with the emplacement of the Allochthon.

Bell Island Granite

The name Bell Island Granite was proposed for the large granite body that comprises the eastern part of Bell Island by Kennedy et al. (1973). It is grey to pink, massive and equigranular, and locally coarse grained to pegmatitic, especially at the northeast tip of the island. Microcline is the most common mineral (40%), oligoclase-andesine and quartz each make up about 20 per cent of the rock, and muscovite and biotite occur in roughly equal amounts (5%). A zone of intrusion breccia up to 1.5 km wide is developed along its northwest margin. Large angular inclusions, screens, and possible roof pendants of Fleur de Lys schist up to 30 m long are surrounded by the granite. The schistosities in these inclusions are disoriented with respect to schistosities in the country rocks and nearby inclusions, indicating that the intrusion of the granite is posttectonic.

Metamorphism at the granite contact is restricted to the local development of calc-silicate skarn in marble inclusions and possible retrogression of garnet to chlorite in nearby schist. Common alteration of albite to microcline close to the granite contact or in inclusions is attributed to potash metasomatism (Kennedy et al., 1973).

Muscovite from a coarse-grained phase of the granite at the northeast tip of Bell Island yielded a K-Ar age 368 ± 16 Ma, Devonian (Wanless et al., 1973, p. 110).

Dykes

A thin purplish weathering lamprophyre dyke with phlogopite phenocrysts cuts the Goose Cove Schist at Deep Water Point and lamprophyre dykes that cut the Goose Cove Schist elsewhere have been described by Cooper (1937). Biotite from one dyke near the village of Goose Cove was dated by K-Ar at 408 ± 17 Ma (Wanless et al., 1968, p. 138 - 139). A similar lamprophyre with biotite phenocrysts cuts the Fleur de Lys Supergroup at the northwest end of Groais Island. The phlogopite yielded a K-Ar age of 353 ± 16 Ma (Wanless et al., 1973, p. 109 - 110). Pebbles of lamprophyre occur in nearby Carboniferous conglomerate of the Crouse Harbour Formation.

Northeast-trending, buff-weathering diabase dykes cut the Goose Cove Schist east of Goose Cove. The dykes are steeply dipping and range from a metre to several metres in width. One example that is about 2 m wide continues along strike for nearly a kilometre (Cooper, 1937). Similarly thin and persistent northeast-trending, southeast-dipping dykes cut the Table Head Formation at Little Spring Inlet and Big Spring Inlet. Other postemplacement dykes cut the wide mélange zone along the western side of Fishot Islands (Fig. 94).

The dykes at Little Spring and Big Spring inlets and Fishot Islands clearly postdate the latest steep cleavage in the bordering rocks. If the Goose Cove lamprophyres are also younger than this latest steep cleavage, then the local isotopic data of 408 Ma set an upper time limit to this postemplacement structural event.

CARBONIFEROUS COVER ROCKS

Gently folded Carboniferous rocks form the Cape Rouge and Conche peninsulas and the same rocks are exposed 6 km offshore at Rouge Island and at the northwest tip of Groais Island. The rocks comprise conglomerate, sandstone, and shale that are faulted against and unconformably overlie the Maiden Point Formation toward the west and they unconformably overlie the Fleur de Lys Supergroup at Groais Island. These Carboniferous rocks are referred to the Crouse Harbour and Cape Rouge formations (Baird, (1966) and the local exposures form the northern end of a belt of Carboniferous strata that extends for several hundred kilometres to the southwest.

Crouse Harbour Formation

The name Crouse Harbour Formation was proposed for the basal Carboniferous conglomerate that occurs at the isthmuses of the Conche and Cape Rouge peninsulas and at the northwest tip of Groais Island. A maximum thickness of 275 m is exposed at Southwest Crouse (Baird, 1966) and about 150 m of conglomerate with top unexposed occurs at Groais Island. The conglomerate is of local derivation. At the Conche and Cape Rouge peninsulas the conglomerate is composed almost entirely of subrounded boulders, cobbles, and pebbles of Maiden Point greywacke with only local vein quartz and quartzite clasts. Those at Groais Island are composed mainly of micaceous schists like the nearby Fleur de Lys Supergroup with local clasts of vein quartz, quartzite, and lamprophyre.

The matrix of the conglomerate is grey to brownish sandstone in the western occurrences but at Groais Island it is reddish sandstone with hematite-stained fragments. At the small island immediately north of Groais Island, the red conglomerate has thick interbeds of red arkosic sandstone near the base of the section. These are followed higher in the section, along the western side of the island, by grey conglomerate that has a large proportion of limestone and dolomite fragments and local pink granite, rhyolite porphyry, and red chert clasts.

Cape Rouge Formation

The Cape Rouge Formation consists of fine-grained sandstone, siltstone and shale that underlie the main parts of both the Conche and Cape Rouge peninsulas. The most continuous section, between Cape Rouge and Northeast Crouse, is approximately 1220 m thick (Baird, 1966). The shale and siltstone weather buff to brown and they are thinly laminated with interbeds of sandstone up to a metre or more thick and with local limestone interbeds about 30 cm thick. Desiccation cracks are abundant throughout the section so that bedding plane surfaces exhibit polygonal patterns of various scale. Ripple marked bedding surfaces are also evident locally.

Fossil plants and plant fragments occur in many places throughout the Cape Rouge Formation at the Conche and Cape Rouge peninsulas and at Rouge Island. Murray (in Murray and Howley, 1881, p. 44) also noted plant remains at the small island, north of Groais Island.

Structure

The Crouse Harbour conglomerate at Conche and Cape Rouge peninsulas dips moderately southeast and occupies the west limb of a northeast-trending syncline. The contact with the overlying Cape Rouge Formation is faulted or unexposed, except at the east side of Biche Arm where it is conformable. The absence of Crouse Harbour conglomerate in easterly exposures across the synclinal axis suggests that they may thin eastward. The Crouse Harbour Formation is cleaved locally at Biche Arm and Southwest Crouse and it has upright open folds about gently southeast-plunging axes. Cleavage strikes north to northwesterly and dips are steep to moderate both to the northeast and southwest.

The Cape Rouge Formation occupies a gently north-plunging open syncline at the southern end of the Conche Peninsula and the structure is truncated northward where the synclinal axis is offset right-laterally by a steep northeast-trending fault between Latin Point and Frauderesse Point. A similar syncline at Cape Rouge Peninsula is complicated by northeast-trending faults at Pilier Bay.

The east-dipping and east-facing beds at Rouge Island imply an anticlinal axis between Cape Rouge and Rouge Island and a complimentary synclinal axis to the east between Rouge Island and Groais Island.

Age and correlation

Fossil plants collected by Baird (1966) indicate a late Tournaisian or early Viséan (early Mississippian) age for the Cape Rouge Formation and correlation with the Chevierie Formation of the Horton Group of Nova Scotia. Spore identifications support this conclusion and indicate a Tournaisian age (Baird, 1966).

STRUCTURAL RESUME

The map area encompasses rocks of wide age span so that its structural history is diverse. The earliest structures are Precambrian and the latest affected Carboniferous cover rocks.

Deformation in basement gneiss

Gneisses of the Basement Gneiss Complex were first deformed and metamorphosed prior to the deposition of Lower Cambrian Bradore sandstone. This deformation is correlated with the Grenville orogenic cycle of the nearby Canadian Shield because of similar structural style and metamorphic grade. In addition, posttectonic granite in the Great Northern Highlands was dated isotopically by K-Ar at 945 and 960 Ma (Lowdon et al., 1963).

Deformation in the Fleur de Lys Supergroup

Rocks of the Fleur de Lys Supergroup, exposed on the Grey Islands, underwent polyphase deformation prior to deposition of the Carboniferous Crouse Harbour Formation. Two major and two minor deformative phases were recognized (Kennedy et al., 1973). The first produced a penetrative fine-grained muscovite-chlorite schistosity that is in most places destroyed by later metamorphic mineral growth. The second produced large scale, west to northwest-facing recumbent folds with an associated muscovite-biotite schistosity.

Later deformations were minor and produced upright, close to open folds and strain-slip or fracture cleavages. Locally, as at the north end of Groais Island, the second phase recumbent folds are refolded into steep positions.

The prograde metamorphism of the Fleur de Lys rocks was both synkinematic and postkinematic with respect to the early deformations and it preceded the third deformation (Kennedy et al., 1973).

Deformation in allochthonous and autochthonous rocks

The structural history of both the allochthonous and autochthonous rocks is complex and major contrasts in structural style are evident throughout the area. Within the Hare Bay Allochthon, polydeformed garnetiferous amphibolite of the St. Anthony Complex is sharply contrasted with undeformed pillow basalts of the nearby Cape Onion Formation. Likewise, the intensity of deformation varies within the autochthonous rocks where at Canada Bay the Table Head Formation consists of polydeformed marble, while at western Hare Bay the same formation is virtually flat lying.

The various structures can be shown in most places to have developed either before, during, or after the formation of mélange and the emplacement of the Allochthon. Pre-emplacement structures are restricted to the St. Anthony slice assemblage and show discordant relations with structures in underlying mélange. Locally, predeformed blocks of Goose Cove Schist are incorporated within mélange. Synemplacement structures are developed in the structurally lower allochthonous slices, mélange zones, and locally in the upper parts of the autochthon. These are characterized by shallow-dipping cleavages, recumbent folds, and thrust faults. Postemplacement structures affect all the rocks and in places fold the various tectonic contacts into steep attitudes.

Pre-emplacement deformation

Structures related to pre-emplacement deformation can only be demonstrated in the St. Anthony slice assemblage. In the Green Ridge Amphibolite and Goose Cove Schist, at least two penetrative deformations preceded final emplacement. The first produced a strong schistosity and local tectonic slides. No folds were observed related to this event. The second produced recumbent folds with related schistosity.

The intensity of development of the first schistosity decreases downwards and away from the contact with the White Hills Peridotite. In the Green Ridge Amphibolite, S_1 is defined by aligned hornblende, biotite, and rods of opaque minerals. It is strongly overprinted by the second fabric and generally only recognizable in the hinges of second phase folds. In the Goose Cove Schist, S_1 is defined by a preferred orientation of chlorite, trémolite, and actinolite. A local schistosity in the Ireland Point Volcanics is attributed to this same deformative event.

The second deformation event in the aureole rocks was the most intense and produced inclined to recumbent, tight to isoclinal folds with an associated axial plane crenulation cleavage in the Goose Cove Schist and a schistosity in the Green Ridge Amphibolite. F_2 folds are common in the Goose Cove Schist and are displayed clearly in sea cliffs south of Goose Cove (Fig. 95). In the Green Ridge Amphibolite, the F_2 folds are difficult to recognize except where interbedded contrasting lithologies are present, for example, marble beds near the western entrance to Fishot Harbour. The folds are recumbent, trend from northeast to north, and generally plunge gently northerly. At Fishot Islands, they face west.

The S_2 cleavage in the Goose Cove Schist consists of a closely spaced crenulation cleavage with little associated metamorphic growth. In the Green Ridge Amphibolite, S_2 is the dominant fabric element with complete transposition of S_1 into the S_2 planes and accompanying growth of aligned biotite and hornblende.

A strong tectonic fabric and associated recumbent folds in the White Hills Peridotite are parallel to structures in the underlying aureole. Recumbent folds trend northeast and the tectonic fabric is defined by flattened and aligned orthopyroxene. This fabric diminishes upwards and away from the aureole contact. At the contact the serpentinized peridotite contains hard, thin amphibolite bands that represent metamorphosed and mylonitized peridotite. Recrystallization and tectonic banding in the serpentinized peridotite at the contact are interpreted therefore to be of the same generation as that in the juxtaposed aureole. If so, the regional pyroxene fabric in the peridotite above is earlier and may therefore reflect deformation or plastic flow within the mantle. The mantle tectonite interpretation of the peridotite fabric is supported by the fact that it is cut posttectonically by thin pyroxenite dykes that are indigenous to the White Hills Peridotite.

Synemplacement deformation

Emplacement deformation produced the tectonic surfaces that define the various slices of the Allochthon, internal deformation concomitant with transport, mélange and its deformation, and local folding and thrusting in autochthonous rocks. These features are treated under this same heading as they cannot be separated or placed in chronologic order, and they relate to the same major episode.

The intensity of internal deformation in the allochthonous rocks is variable. The Northwest Arm Formation exhibits a brittle deformation with discontinuous recumbent folds and a chaotic, rubbly internal character. The Maiden Point Formation has a weak to strong slaty cleavage with associated recumbent folds. These are clearest along the north shore of St. Anthony Bight where they face slightly downwards to the southwest on the associated cleavage. South of Hare Bay at Crouque Harbour, the folds face upward to the northwest. Major recumbent folds south of Canada Head are inferred from cleavage-bedding relationships.

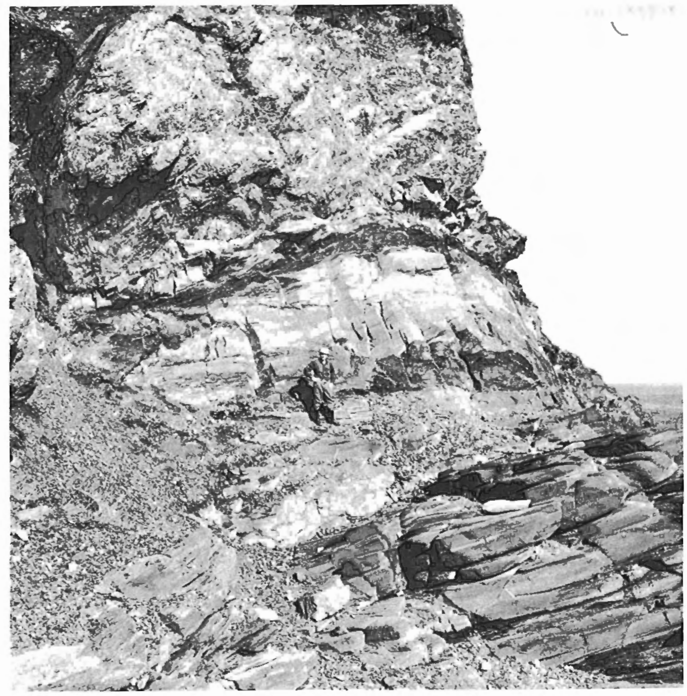


Figure 95. Recumbent pre-emplacement fold in greenschist of the Goose Cove Schist, Greenwood Cove, Hare Bay. (GSC 202609-I).

No emplacement folds were recognized in the Grandois Group or the Cape Onion Formation but a slaty cleavage is locally present.

A penetrative subhorizontal cleavage with local associated minor folds is developed in the mélanges on the south side of Hare Bay, especially at Fishot Islands and southwest of Great Islets Harbour. This cleavage is folded about postemplacement upright folds.

In most places the autochthonous rocks are unaffected by emplacement deformation. However, in Canada Bay a series of east-dipping imbricate thrust faults affects the autochthonous succession and involves basement gneiss as well. The thrusts are exposed across the south shore of Canada Bay where the autochthonous succession is repeated five times. A strong slaty cleavage and minor recumbent folds are associated with the thrusting.

The Canada Bay thrusts were related to emplacement of the Allochthon as they parallel the basal thrust of the Maiden Point slice assemblage, associated recumbent folds indicate that tectonic transport was westward, and the thrusts and associated structures in both allochthonous and autochthonous rocks at Canada Bay were affected by postemplacement folds (Smyth, 1973). These thrusts in the autochthonous rocks appear to die out rapidly northward away from outcrops of the basement gneiss.

At Whites Arm Window, a penetrative slaty cleavage is developed in the Goose Tickle Formation and the upper parts of the Table Head Formation and it decreases in intensity down the succession. West of the Allochthon contact in Hare Bay, the autochthonous rocks are unaffected by emplacement deformation.

Postemplacement deformation

Stratigraphic evidence indicates that the Hare Bay Allochthon was emplaced by Middle Ordovician time. After emplacement, the rocks were affected by another major deformation that increases in intensity from west to east across the area. Folds related to this event are steep and trend northeast. They control the regional structure south of Hare Bay. North of Hare Bay, they are generally restricted and only intense in eastern most exposures around Quirpon and St. Carols.

Postemplacement structures fold and cleave the Allochthon-Autochthon contact, and the earlier tectonic elements related to pre-emplacement and synemplacement deformations. The effect of postemplacement deformation is most evident in the map pattern in the Whites Arm Window where the Table Head and Goose Tickle formations are exposed in a broad anticlinal structure through the Maiden Point slice assemblage.

Postemplacement structures in autochthonous rocks west of the Allochthon are gentle to open, westerly inclined, northeast-plunging flexural folds. Eastwards across the Allochthon the folds steepen and tighten with an associated penetrative slaty or crenulation cleavage. Recumbent emplacment folds and associated cleavage are refolded west of St. Carols, and folded cleavage in the Maiden Point Formation is well displayed on the peninsula near Croque.

Emplacement thrusts and recumbent folds in autochthonous rocks at Canada Bay are refolded about upright axial surfaces and refolded folds in the Table Head Formation are exposed on the headland west of Burnt Point on the south side of Hare Bay. At Fishot Islands, the second phase pre-emplacement folds in th St. Anthony Complex are refolded about open northeast-trending, postemplacement folds.

Deformation in Carboniferous cover rocks

The Carboniferous rocks at Conche and Cape Rouge peninsulas are folded into large open upright anticlines and synclines about northeast-trending axes. Slaty cleavage is developed only locally. These folds had no apparent effect on nearby older rocks.

Discussion

The Grenville orogenic cycle affected eastern North America prior to initiation of the Appalachian System. The deposition and evolution of the allochthonous and autochthonous successions has been related to the opening of an ocean and formation of its western continental margin. Likewise, the successive orogenic events that deformed and metamorphosed these rocks are related to closure of this ocean and the resultant destruction of the continental margin (Smyth, 1973; Williams and Stevens, 1974).

The earliest pre-emplacement deformations affected the St. Anthony slice assemblage which formed farthest from the morphological shelf edge. The deformation and metamorphism of the aureole rocks were interpreted as the result of westward obduction of hot ophiolite in early Middle Ordovician or late Early Ordovician time (Williams and Smyth, 1973).

The Fleur de Lys rocks on the Grey Islands show similar structural style to the metamorphic rocks of the St. Anthony slice assemblage. They were probably also deformed in response to ophiolite obduction but at a deeper level and under more uniform conditions.

The emplacement deformation occurred synchronously with mélange production and westerly transport of the Allochthon. It began during earliest assembly of the slices and continued through the entire depositional period of the Goose Tickle Formation. This time period may have extended over several graptolite zones from late Lower to mid-Middle Ordovician (Williams and Stevens, 1974). The event is correlated with Taconic orogeny elsewhere in the Appalachians. In Newfoundland, it is characterized above all by subhorizontal structures.

In contrast, postemplacement deformation is characterized by generally upright folds and steep cleavage. It represents a regional compression and shortening across the area and the same structural style is evident in central Newfoundland where it involves Silurian and Devonian rocks (Williams et al., 1972). The postemplacement event in Hare Bay is therefore correlated with Acadian orogeny. It is viewed as representing late compression and further tightening across an already foreshortened and deformed terrane.

Post-Carboniferous deformation like that at Conche and Cape Rouge peninsulas is confined to a narrow zone in western Newfoundland that is also characterized by major high-angle faults.

ECONOMIC GEOLOGY

There are at present no commercial mining operations in the Hare Bay area. An occurrence of copper was actively mined at Goose Cove shortly after the turn of the century and two marble quarries were opened at Canada Harbour at about the same time. More complete descriptions of these mineral occurrences are given by Cooper (1937), Betz (1939), and Douglas et al. (1940).

The Goose Cove deposit consists of lenses of massive pyrite with associated pyrrhotite and chalcopyrite. These are structurally controlled and occur at the noses of moderately northwest-plunging folds in the Goose Cove Schist. An estimated 1600 tonnes of ore was mined, averaging from 2 to 12 per cent copper. The mining operations were abandoned after working to depths in the order of 30 m. This suggests a cessation of ore at the structural base of the St. Anthony slice assemblage at about that level.

Fine-grained white marble at Canada Harbour was quarried but no commercial shipments were made. Most of the material is cleaved so that the product is not well suited for solid building stone.

Chromite, asbestos, and nickel were reported from the White Hills Peridotite (Cooper, 1937) and copper from the Maiden Point Formation at St. Julien's (Douglas et al., 1940), but all are of minor extent.

SELECTED BIBLIOGRAPHY

Arai, M.N. and McGugan, A.
1968: A problematical coelenterate (?) from the Lower Cambrian, near Moraine Lake, Banff area, Alberta; Journal of Paleontology, v. 42, no. 1, p. 205-209.

Baird, D.M., compiler
1954: Geological map of Newfoundland; Geological Survey of Newfoundland, (scale 1 inch to 12 miles).
1956: Geological map coverage of Newfoundland; Department of Mines and Resources, Geological Survey of Newfoundland, Report 6A, 4 p.
1960a: Sandy Lake (West Half) map area; Geological Survey of Canada, Map 47- 1959.
1960b: Observations on the nature and origin of the Cow Head breccias of Newfoundland; Geological Survey of Canada, Paper 60-3, 26 p.
1966: Carboniferous rocks of the Conche-Groais Island area, Newfoundland; Canadian Journal of Earth Sciences, v. 3, p. 247-257.

Balsam, W.L.
1973: Ecological interactions in an early Cambrian archaeocyathid reef community; unpublished Ph.D. thesis, Brown University, 131 p.

Barnes, C.R. and Tuke, M.F.
1970: Conodonts from the St. George Formation (Ordovician), northern Newfoundland; Geological Survey of Canada, Bulletin 187, p. 79-97.

Bayfield, N.W.
1845: On the junction of the transition and primary rocks of Canada and Labrador; Geological Society of London, Quarterly Journal, v. 1, p. 450-459.

Beaglehole, J.C.
1974: The life of Captain James Cook; A. & C. Black, 771 p.

Beales, F.W., Carracedo, J.C., and Strangway, D.W.
1974: Paleomagnetism and the origin of Mississippi Valley-type ore deposits; Canadian Journal of Earth Sciences, v. 11, p. 211-223.

Bell, R.
1885: Observations on the mineralogy, zoology and botany of the Labrador coast, Hudsons Strait and Bay; Geological Survey of Canada, Reports of Progress 1882-83-84, pt. DD, 62 p.

Bergström, S.M., Riva, J., and Kay, M.
1974: Significance of conodonts, graptolites, and shelly faunas from the Ordovician of western and north-central Newfoundland; Canadian Journal of Earth Sciences, v. 11, p. 1625-1660.

Berry, W.B.N.
1960: Graptolite faunas of the Marathon Region, West Texas; University of Texas Press, Publication no. 6005, p. 1-179.
1972: Early Ordovician bathyurid province lithofacies, biofacies and correlations---their relationships to a proto-Atlantic Ocean; Lethaia, v. 5, p. 69-83.

Betz, F., Jr.
1939: Geology and mineral deposits of the Canada Bay area, northern Newfoundland; Geological Survey of Newfoundland, Bulletin 16, 53 p.
1943: Late Paleozoic faulting in western Newfoundland; Geological Society of America, Bulletin, v. 54, p. 687-706.

Bidgood, D.E.T.
1974: Final report on the bottom and sub-bottom surveys in the Strait of Belle Isle, Newfoundland; Nova Scotia Research Foundation, 54 p., 8 fig. and 16 maps, unpublished.

Billings, E.
1865: Paleozoic fossils: volume 1, Geological Survey of Canada.

Bird, J.M. and Dewey, J.F.
1970: Lithosphere plate-continental margin tectonics and the evolution of the Appalachian Orogen; Geological Society of America, Bulletin, v. 81, p. 1031-1060.

Black, W.A.
1965: Forest utilization, Great Northern Peninsula, Newfoundland; ARDA study 1043, Agriculture Rehabilitation and Development Administration, Department of Forestry, Ottawa, 66 p.

Bostock, H.H. and Cumming, L.M.
1973: Some notes on the Precambrian rocks of the Gros Morne National Park, western Newfoundland; in Report of Activities, Part B, Geological Survey of Canada, Paper 73-1B, p. 109-119.

Bostock, H.S.
1970: Physiographic subdivisions of Canada; in Geology and Economic Minerals of Canada, editor R.J.W. Douglas; Geological Survey of Canada, Economic Geology Report 1, p. 11-30.

Brückner, W.D.
1966: Stratigraphy and structure of West-Central Newfoundland; in Guidebook---Geology of Parts of the Atlantic Provinces, editor W.H. Poole; Geological Association of Canada/Mineralogical Association of Canada, Annual Meeting, Halifax, 1966, p. 137-155.

Brunton, S.
1922: The west coast of Newfoundland; Mining Magazine, v. 27, p. 265-273.

Bulman, O.M.B.
1950: Graptolites from the Dictyonema Shales of Quebec; Geological Society of London, Quarterly Journal, v. 106, p. 63-99.

1971: Garaptolite faunal distribution; in Faunal Provinces in Space and Time, editor F.A. Middlemiss et al., Geological Journal Special Issue 4, p. 47-60.

Butler, J. and Bartlett, G.
1969: Bibliography of the geology of Newfoundland and Labrador 1813 through 1968; Newfoundland Department of Mines, Agriculture and Resources, Mineral Resources Division, Bulletin 38, 273 p.

Cajka, C.J.
1969: Great Northern Peninsula concession, Newfoundland; Cominco Ltd., unpublished report no. 2, 12 p., 10 maps, in files of Mineral Development Division, Newfoundland and Labrador Department of Mines and Energy.

Callahan, W.H.
1974: Syngenesis versus epigenesis of Mississippi Valley-Appalachian type base metal mineral deposits; Economic Geology, v. 69, p. 123,124.

Chamberlain, T.C.
1895: Notes on the glaciation of Newfoundland (Abstract); Geological Society of America, Bulletin, v. 6, p. 467.

Chayes, Felix
1956: Petrographic Modal Analysis, an Elementary Statistical Appraisal; John Wiley & Sons, Inc., New York, 113 p.

Choquette, P.W.
1955: A petrographic study of the "State College" siliceous oolite; Journal of Geology, v. 63, p. 337 - 347.

Christie, A.M.
1951: Geology of the southern coast of Labrador from Forteau Bay to Cape Porcupine, Newfoundland; Geological Survey of Canada, Paper 51-13, 19 p.

Church, W.R.
1969: Metamorphic rocks of the Burlington Peninsula and adjoining areas of Newfoundland, and their bearing on continental drift in North Atlantic; in North Atlantic-Geology and Continental Drift, editor M. Kay; American Association of Petroleum Geologists, Memoir 12, p. 212 - 233.

Church, W.R. and Stevens, R.K.
1971: Early Paleozoic ophiolite complexes of the Newfoundland Appalachians as mantle-oceanic crust sequences; Journal of Geophysical Research, v. 76, p. 1460 - 1466.

Clifford, P.M.
1965: Paleozoic flood basalts in northern Newfoundland and Labrador; Canadian Journal of Earth Sciences, v. 2, p. 183 - 187.

Clifford, P.M. and Baird, D.M.
1962: Great Northern Peninsula of Newfoundland-Grenville inlier; Canadian Mining and Metallurgical Bulletin, no. 599, p. 150 - 157 and 276, 277.

Coleman, A.P.
1926: The Pleistocene of Newfoundland; Journal of Geology, v. 34, p. 193 - 223.

Collins, J.A. and Smith, L.
1972: Sphalerite as related to tectonic movements, deposition, diagenesis and karstification of a carbonate platform; 24th International Geological Congress, Montreal, (Abstract) p. 182,183 and section 6, p. 208 - 218.
1973: Lithostratigraphic controls of some Ordovician sphalerite; in Ores, Mn Sediments, G.C. Amstutz and A.J. Bernard, editors; International Union of Geological Series A, no. 3, p. 79 - 91.
1975: Zinc deposits related to diagenesis and intrakarstic sedimentation in the Lower Ordovician St. George Formation, western Newfoundland; Canadian Petroleum Geology, Bulletin, v. 23, p. 393 - 427.

Comeau, R.L.
1972: Transported slices of the Coastal Complex, Bay of Islands, western Newfoundland; unpublished M.Sc. thesis, Memorial University of Newfoundland, 105 p.

Cook, James
1763: A sketch of the Island of Newfoundland (scale 20 leagues to a degree); National Archives of Canada, VI/100.
1767: An exact trigonometrical survey of the west coast of the Island of Newfoundland taken by order of Commodore Palliser, Governor of Newfoundland-Labrador etc. scale 6000 ft. to an inch; National Archives of Canada---English charts, Ser. 1, No. 1.
1768: Directions for navigating the west coast of Newfoundland, London.
1965: James Cook, surveyor of Newfoundland. Being a collection of charts of the coast of Newfoundland and Labrador, etc., drawn from original surveys taken by James Cook and Michael Lane's London, Jefferys, 1769 - 1770. Facsimile with introductory essay by R.A. Skeleton, San Francisco. David Magee.

Cook James and Bullock, Lieut. F.
1834: Chart of the northern part of the Island of Newfoundland and the Strait of Belle Isle; Hydrographical Office of the Admiralty.

Cook James, Lane, M. and Gilbert, J.
1775: The North American pilot for Newfoundland, Labrador and the Gulf of St. Lawrence; engraved by T. Jeffrey, London.

Copper, G.A.
1956: Chazy and related brachiopods; Smithsonian Miscellaneous Collection, v. 127, 1245 p.(See pages 246, 341, 532, 560, 699, 712, 1108, 1116).

Cooper, J.R.
1937: Geology and mineral deposits of the Hare Bay area; Newfoundland Department of Natural Resources, Geology Section, Bulletin 9, 36 p.

Cowie, J.W.
1971: Lower Cambrian faunal provinces; in Faunal Provinces in Space and Time, editor F.A. Middlemiss et al.; Geological Journal, Special Issue 4, p. 31 - 46.

Cumming, L.M.
1965a: Biostratigraphic studies, Port au Port area; in Report of Activities, field, 1964, Geological Survey of Canada, Paper 65-1, p. 133 - 135.
1965b: Stratigraphy of the west coast of Newfoundland; in Report of Activities, Part II, Geological Survey of Canada, Paper 65-2, Part II, p. 66 - 67.
1966: Tectonics and stratigraphy of the west coast of Newfoundland (Abstract); in Abstracts of Publications in Scientific Journals by Officers of the Geological Survey of Canada, 1965, Geological Survey of Canada, Paper 66-4. p. 16,17.
1967a: Platform and klippe tectonics of western Newfoundland: a review; in Appalachian Tectonics, editor T.H. Clark; Royal Society of Canada, Special Publication 10, p. 10 - 17.
1967b: Table Head and St. George formations, western Newfoundland; in Report of Activities, Part A: May to October, 1966, S.E. Jenness, editor; Geological Survey of Canada, Paper 67-1, Part A, p. 178,179.

Cumming, L.M. (cont.)

1967c: Clonograptus from the St. George Formation Newfoundland; in Report of Activities, Part B, November, 1966 to April, 1967; Geological Survey of Canada, Paper 67-1, Part B, p. 61 - 63.

1968: St. George-Table Head disconformity and zinc mineralization, western Newfoundland; Canadian Mining and Metallurgical Bulletin 674, v. 61, p. 721 - 725.

1970: Operation Strait of Belle Isle, Quebec and Newfoundland-Labrador; (parts of 2M, 3D, 12I, 12P, 13A); Geological Survey of Canada, Paper 70-1A, p. 3 - 8.

1971: Operation Strait of Belle Isle, Newfoundland and Labrador; Geological Survey of Canada, Paper 71-1A, p. 2 - 6.

1972: Operation Strait of Belle Isle, Quebec and Newfoundland-Labrador (parts of 2M, 12I, 12P); Geological Survey of Canada, Paper 72-1A, p. 3 - 7.

1975: Geology of the L'Anse aux Meadows National Historic Park, northern Newfoundland (2M/11, 12); in Report of Activities, Part A, Geological Survey of Canada, Paper 75-1A, p. 3 - 7.

Cumming, L.M. and Grant, D.R.

1974: Crustal delevelling in northern Newfoundland; in Report of Activities, Part B, Geological Survey of Canada, Paper 74-1B, p. 7 - 10.

Cumming, L.M. and Tella, S.

1971: Stratigraphy and structure of Belle Isle region, Quebec and Newfoundland-Labrador (Abstract); Geological Association of Canada, Program Annual Meeting, Sudbury, p. 17.

Daly, R.A.

1903: Variolitic pillow-lava from Newfoundland; American Geologist, v. 32, p. 65 - 78.

Davies, G.R.

1970: Algal-laminated sediments, Gladstone embayment, Shark Bay, Western Australia; American Association Petroleum Geologists, Memoir 13, p. 169 - 205.

Davies, R.

1965: Geology of the Cook-D'Audhebourg area; Quebec Department of Natural Resources, Preliminary Report 537, p. 1 - 15.

Dawson, J.W.

1866: Comparison of icebergs of Belle Isle with the glaciers of Mont Blanc, with reference to the boulder-clay of Canada; Canadian Naturalist and Geologist, v. 3, p. 33 - 44.

DeGrace, J.R.

1974: Limestone resources of Newfoundland and Labrador; Newfoundland and Labrador Department of Mines and Energy, Report 74-2, 117 p.

Dewey, J.F.

1976: Ophiolite obduction; Tectonophysics, v. 31, p. 93 - 120.

de Wit, M.J.

1974: On the origin and deformation of the Fleur de Lys metaconglomerate, Appalachian Fold Belt, Northeast Newfoundland; Canadian Journal of Earth Sciences, v. 11, p. 1168 - 1180.

Dietz, R.S., Emery, K.O. and Shepard, F.P.

1942: Phosphorite deposits of the sea-floor off southern California; Geological Society of America Bulletin, v. 53, p. 815 - 848.

Douglas, G.V.

1953: Notes on localities visited on the Labrador coast in 1946 and 1947; Geological Survey of Canada Paper 53-1, 67 p.

Douglas, G.V., Williams, D. and Rove, O.N.

1940: Copper Deposits of Newfoundland; Newfoundland Geological Survey, Bulletin 20, 176 p.

Drapeau, G.

1969: Sedimentology of the narrowest portion of the Strait of Belle Isle; unpublished report, Bedford Institute of Oceanography, 10 p., 12 fig. and Appendix A - C.

Dunbar, C.O.

1933: Mid-Ordovician thrust faulting in western Newfoundland; Geological Society of America, Proceedings, p. 76 (plus comments by Twenhofel on p. 76,77).

Eade, K.E.

1962: Battle Harbour-Cartwright; Geological Survey of Canada, Map 22-1962.

Emslie, R.F.

1970: The geology of the Michikamau intrusion, Labrador (13L, 23I); Geological Survey of Canada, Paper 68-57.

1973: Some chemical characteristics of anorthosite suites and their significance; Canadian Journal of Earth Sciences, v. 10, p. 54 - 71.

Emslie, R.F., Irving E. and Park, J.K.

1976: Further paleomagnetic results from the Michikamau Intrusion, Labrador; Canadian Journal of Earth Sciences, v. 13, p. 1052 - 1057.

Erdtmann, B.

1971a: Ordovician graptolite zones of western Newfoundland in relation to the paleogeography of the North Atlantic; Geological Society of America, Bulletin, v. 82, p. 1509 - 1527.

1971b: *Tetragraptus otagoensis* and *Janograptus terranovensis* (n. sp.) Ordovician graptolites from western and northern Newfoundland; Journal of Paleontology, v. 45, p. 258 - 264.

Fåhraeus, L.E.

1974: Conodont-based correlations of Lower and Middle Ordovician strata in western Newfoundland; Geological Society of America, Bulletin, v. 81, p. 2061 - 2076.

Fish, R.

1974: Newfoundland zinc readying for production; Canadian Mining Journal, v. 95, no. 12, p. 20 - 24.

Fleming, J.M.

1970: Petroleum exploration in Newfoundland and Labrador; Newfoundland Department of Mines and Energy, Report 3, 118 p.

1971: "Once upon a billion years ago..." The geological history of the Great Northern Peninsula, Newfoundland; The Great Northern Peninsula: An Archeological and Geological History; Park Interpretation Publication 3, Newfoundland and Labrador Provincial Parks, p. 19 - 27.

Flint, R.F.
1940: Late Quaternary changes of level in western and southern Newfoundland; Geological Society of America, Bulletin, v. 51, p. 1757 - 1780.

Flower, R.H.
1975: American Lituitidae (Cephalopoda); Bulletins of American Paleontology, v. 67, no. 287, p. 139 - 173.
1976: Part I---New American Wutinoceratidae with review of Actinoceroid occurrences in Eastern Hemisphere, and Part II---Some Whiterock and Chazy Endoceroids; New Mexico Bureau of Mines and Mineral Resources, Memoir 28, 76 p.

Fogwill, W.D.
1965: Mines and mineral occurrences map of the Island of Newfoundland; Newfoundland Department of Mines and Agricultural Resources, Information Circular 11.
1966: Petroleum and natural gas in west Newfoundland with emphasis on Parson's Pond area; Newfoundland and Labrador Corporation, St. John's, unpublished 275 p.
1970: Mineral deposits and prospecting environments in Newfoundland; Newfoundland and Labrador Department of Mines and Agricultural Resources, Information Circular 14, 45 p.

Fogwill, W.D. and Hawkins, V., compilers
1976: Bibliography of the geology of Newfoundland and Labrador, 1969 to 1974; Newfoundland Department of Mines and Energy, Open File 887, 355 p.

Foley, F.C.
1937: Geology and mineral deposits of Hawke Bay-Great Harbour Deep area, northern Newfoundland; Geological Survey of Newfoundland, Bulletin 10, 22 p.

Fong, C.C.K.
1967: Paleontology of the Lower Cambrian archaeocyathus-bearing Forteau Formation in southern Labrador; unpublished M.Sc. thesis, Memorial University of Newfoundland, 227 p.
1972: Newfoundland and Labrador mining industry; Georama 1972, 24th International Geological Congress, Montreal, 18 p. and map.

Fritts, C.E.
1953: Geological reconnaissance across the Great Northern Peninsula of Newfoundland; Geological Survey of Newfoundland, Report 4, 22 p.

Fritz, W.H.
1973: Cambrian biostratigraphy MacKenzie Mountains and Labrador; in Report of Activities, Part A, Geological Survey of Canada, Paper 73-1A, p. 254 - 259.

Geological Survey of Canada
1969: Blanc-Sablon, Quebec, Aeromagnetic series, Geological Survey of Canada Map 7367G (scale 1 inch to 4 miles).
1970: Port Saunders, Newfoundland, Aeromagnetic series, Geological Survey of Canada Map 7357G (scale 1 inch to 4 miles).

Gillis, J.W.
1966: Great Northern Peninsula; in Report of Activities, May to October 1965, Geological Survey of Canada, Paper 66-1, p. 178 - 181.

Goodacre, A.K., Brule, B.G. and Copper, R.V.
1969: Results of regional underwater gravity surveys in the Gulf of St. Lawrence; Dominion Observatory, Ottawa, Gravity Map Series 86.

Gosling, W.G.
1910: Labrador, its discovery, exploration and development; London.

Grabau, R.W.
1936: Paleozoic formations in the light of the pulsation theory; v. II, University Press, Peiping, 751 p.

Grant, D.R.
1969a: Surficial deposits, geomorphic features, and Late Quaternary history of the terminus of the Great Northern Peninsula of Newfoundland and adjacent Quebec - Labrador; Maritime Sediments, v. 5, p. 123 - 125.
1969b: Late Pleistocene re-advance of piedmont glaciers in western Newfoundland; Maritime Sediments, v. 5, p. 126 - 128.
1970: Quaternary geology, Great Northern Peninsula, Island of Newfoundland; in Report of Activities, Part A: April to October, 1969, Geological Survey of Canada, Paper 70-1, Part A, p. 172 - 174.
1971a: Pleistocene and Recent history; in The Great Northern Peninsula, an Archeological and Geological History, compiler G.F. Draskoy, Newfoundland Department of Tourism, Park Interpretation Publication 3, p. 27 - 31.
1971b: Surficial geology, western Newfoundland; in Report of Activities, Part A: April to October, 1971, Geological Survey of Canada, Paper 72-1, Part A, p. 157 - 160.
1972: Postglacial emergence of northern Newfoundland; in Report of Activities, Part B: November 1971 to March 1972, Geological Survey of Canada, Paper 72-1, Part B, p. 100 - 102.

Grasty, R.L., Rucklidge, J.C. and Elders, W.A.
1969: New K-Ar age determination on rocks from the east coast of Labrador; Canadian Journal of Earth Sciences, v. 6, p. 340 - 344.

Greene, B.A., compiler
1972: Geological map of Labrador, scale 1:1 000 000; Newfoundland Department of Mines and Energy.
1974: An outline of the geology of Labrador; Newfoundland Department of Mines and Energy, Information Circular 15, 64 p.

Grenfell, Sir W.T.
1913: Labrador: the country and the people; Macmillan, New York, 350 p.
1925: The story of a Labrador doctor: Hodder and Stoughton, London, 158 p.

Hale, W.E.
1961: Geological reconnaissance, St. Augustin area, north shore, Gulf of St. Lawrence; Canadian Mining and Metallurgical, Bulletin, v. 55, p. 326, 331.

Hamilton, W. and Myers, W.B.
1967: The nature of batholiths; U.S. Geological Survey, Professional Paper 554-C, 30 p.

Harp, E., Jr.
1964: The cultural affinities of the Newfoundland Dorset Eskimos; National Museum of Canada, Bulletin 200, p. 183.

Harris, I.M.
1962: Newfoundland limestone; Newfoundland Department of Mines, Agriculture and Resources, unpublished report, 81 p.

Harris, W.J. and Thomas D.E.
1935: Victorian graptolites; Royal Society of Victoria, Proceedings, New Series, Part 3, v. 47, p. 288-313.

Hart, S.
1962: Excess radiogenic argon in pyroxenes; Journal of Geophysical Research, v. 67, p. 2998, 2999.

Haworth, R.T. and Sanford, B.V.
1976: Paleozoic geology of northeast Gulf of St. Lawrence; in Report of Activities, Part A, Geological Survey of Canada, Professional Paper 76-1A, p. 1-6.

Higgins, M.W.
1971: Cataclastic rocks; U.S. Geological Survey, Professional Paper 687, p. 1-97.

Hinde, G.J.
1889: On *Archaeocyathus* Billings, and on other genera, allied to or associated with it, from the Cambrian strata of North America, Spain, Sardinia, and Scotland; Geological Society of London, Quarterly Journal, v. 45, p. 125-148.

Hoagland, A.D.
1971: Appalachian strata-bound deposits, their essential features, genesis and the exploration problem; Economic Geology, v. 66, p. 805-810.

Hoagland, A.D., Hill, W.T. and Fullweiller, R.E.
1965: Genesis of Ordovician zinc deposits in East Tennessee; Economic Geology, v. 60, p. 693-714.

Hoegbom, A.G.
1915: Zur Deutung der Scolithus-Sandsteine und "Pipe-Rocks"; Bulletin of the Geological Institute of Uppsala, v. 13, p. 45-60.

Howell, B.F.
1943: Faunas of the Cambrian Cloud Rapids and Treytown Pond formations of northern Newfoundland; Journal of Paleontology, v. 17, p. 236-247.

Hutchison, W.W.
1969: Metamorphic framework and plutonic styles in the Prince Rupert region of the central Coast Mountains, British Columbia; Canadian Journal of Earth Sciences, v. 7, p. 376-405.

Ingstad, H.
1962: Discovery of Vinland; The Arctic Circular, v. 15, no. 1, p. 2-13.

Jackson, D.E.
1969: Ordovician graptolite faunas in lands bordering north Atlantic and Arctic Oceans; in North Atlantic-Geology and Continental Drift, editor M. Kay; American Association of Petroleum Geologists, Memoir 12, p. 504-512.

James, J.F.
1892: Studies in problematic organisms---the genus *Scolithus*; Geological Society of America Bulletin, v. 3, p. 32-44.

James, N.P. and Fong, C.C.K.
1976: Archeocyathid reefs: northern maritime Appalachians (Abstract); Geological Association of Canada, Program with Abstracts, v. 1, p. 83.

Jamieson, R.A.
1977: A suite of alkali basalts and gabbros associated with the Hare Bay Allochthon of western Newfoundland; Canadian Journal of Earth Sciences, v. 14, p. 346-356.

Johannsen, A.
1932: A descriptive petrography of the igneous rocks, Volume 2 (The quartz-bearing rocks); University of Chicago Press.

Johnson, H.
1941a: Paleozoic lowlands of northwestern Newfoundland; New York Academy of Science, Transactions Series II, v. 3, no. 6, p. 141-145.
1941b: Possible future oil provinces of eastern Canada; American Association of Petroleum Geologists, Bulletin, v. 25, p. 1558-1562.

Jones, P.J.
1971: Lower Ordovician conodonts from the Bonaparte Gulf Basin and the Daly River Basin, Northwestern Australia; Australian Bureau of Mineral Resources, Bulletin 117, p. 38-41.

Jukes, J.B.
1842: Excursions in and about Newfoundland during the years 1839 and 1840; London, 2 volumes.
1873: Geological map of Newfoundland; Newfoundland Geological Survey.

Karolyi, M.
1976: Karst of the Western Platform of Newfoundland, preliminary observations (Abstract); Program with Abstracts, Geological Association of Canada, v. 1. p. 61.

Kay, M.
1969: Continental Drift in North Atlantic Ocean; in North Atlantic, Geology and Continental Drift, editor M. Kay; American Association of Petroleum Geologists, Memoir 12, p. 965-973.

Kennedy, M.J.
1973: Pre-Ordovician polyphase structure in the Burlington Peninsula of the Newfoundland Appalachians; Nature, Physical Science, v. 241, p. 114-116.

Kennedy, M.J., Williams, H. and Smyth, W.R.
1973: Geology of the Grey Islands, Newfoundland---Northernmost extension of the Fleur de Lys Supergroup; Geological Association of Canada, v. 25, p. 79-90.

Kindle, C.H. and Whittington H.B.
1965: New Cambrian and Ordovician fossil localities in western Newfoundland; Geological Society of America, Bulletin, v. 76, p. 685-688.

Kluyver, H.M.
1975: Stratigraphy of the Ordovician St. George Group in the Port au Choix area, western Newfoundland; Canadian Journal of Earth Sciences, v. 12, p. 589-594.

Krank, E.H.
1939: Bedrock geology of the seaboard region of Newfoundland-Labrador; Newfoundland Geological Survey, Bulletin 19, 44 p.

Krogh, T.E. and Davis, G.L.
1973: The significance of inherited zircons on the age and origin of igneous rocks---an investigation of the ages of the Labrador adamellites; Carnegie Institution of Washington, Yearbook, v. 72, p. 610-613.

Lieber, O.M.
1860: Notes on the geology of the coast of Labrador; report of the Superintendent of the U.S. Coast Survey, 1860, p. 402-408.

Lilly, H.D.
1967: Some notes on stratigraphy and structural style in central-west Newfoundland; in Geology of the Atlantic Region, editors E.R.W. Neale and H. Williams; Geological Association of Canada, Special Paper 4, p. 201-211.

Lochman, C.
1938: Middle and Upper Cambrian faunas from western Newfoundland; Journal of Paleontology, v. 12, p. 461-477.

Logan, W.E.
1863: Geology of Canada; Geological Survey of Canada, Reports of Progress to 1863, Montreal. (Includes observations of James Richardson, 1860 to 1862, in western Newfoundland.

Lowden, J.A., Stockwell, C.H., Tipper, H.W. and Wanless, R.K.
1963: Age determination and geological studies (including isotopic ages---report 3); Geological Survey of Canada, Paper 62-17.

Lysaght, A.M.
1971: Joseph Banks in Newfoundland and Labrador, 1766---his diary, manuscripts and collections; University of California Press, 512 p.

MacClintock, P. and Twenhofel, W.H.
1940: Wisconsin glaciation of Newfoundland; Geological Society of America, Bulletin, v. 51, p. 1729-1756.

Mazzullo, S.J. and Friedman, G.M.
1975: Conceptual model of tidally influenced deposition on margin of epeiric seas: Lower Ordovician (Canadian) of eastern New York and southwestern Vermont; American Association of Petroleum Geologists, v. 59, p. 2123-2141.

McKillop, J.H.
1962: Limestone potential of Newfoundland; Canadian Mining Journal, v. 83, p. 80-83.

McMillan, N.J.
1973: Geology of Labrador Sea and Baffin Bay; Oilweek, v. 24, no. 13, p. 59-66.

Monahan, D.
1971: On the morphology of the northeast Gulf of St. Lawrence; Maritime Sediments, v. 7, no. 2, p. 72-75.

Morris, R.W. and Kay, M.
1966: Ordovician graptolites from the Middle Table Head Formation at Black Cove, near Port au Port, Newfoundland; Journal of Paleontology, v. 40, p. 1223-1229.

Mullins, W.J., Fogwill, W.D. and Greene, B.A.
1969: Map index, geological, geophysical and related maps; Newfoundland Department of Mines, Agriculture and Resources, Information Circular 13, 59 p.

Murray, A.
1877a: Geography and resources of Newfoundland; Journal of the Royal Geographic Society, v. 43, no. 13, p. 267-278.
1877b: Geological map of Newfoundland; Geological Survey of Newfoundland.

Murray, A. and Howley, J.P.
1881: Geological Survey of Newfoundland from 1864 to 1880; London, Edward Stanford, 536 p.
1918: Reports of Geological Survey of Newfoundland from 1881 to 1909; St. John's, Robinson & Company, Ltd., Press, 725 p.

Neale, E.R.W.
1972: A cross section through the Appalachian Orogen in Newfoundland; 24th International Geological Congress, Montreal, 1972, guidebook to excursion A62-C62, 84 p.

Neale, E.R.W. and Nash, W.A.
1963: Sandy Lake (East half), Newfoundland, 12H E1/2; Geological Survey of Canada, Paper 62-28, 40 p.

Nelson, S.J.
1955: Geology of Portland Creek, Port Saunders area, West Coast; Geological Survey of Newfoundland, Report 7, 58 p.

North, F.K.
1971: The Cambrian of Canada and Alaska; in Cambrian of the New World, editor C.K. Holland; Wiley Interscience, p. 219-324.

Nova Scotia Research Foundation
1973: Final Report on a bathymetric and bottom profiling survey in the Strait of Belle Isle, Newfoundland; unpublished report to Newfoundland and Labrador Power Commission, v. 1, sec. 3, 26 p., 29 fig., 4 maps.

Okulitch, V.J.
1946: Intervallum structure of *Cambrocyathus amourensis*; Journal of Palaenotology, v. 20, p. 275-278.

Oxley, P.
1953: Geology of Parsons Pond-St. Pauls area, west coast; Geological Survey of Newfoundland, Report 5, 53 p.

Packard, A.S.
1891: The Labrador Coast; N.D.C. Hodges, 47 Lafayette Place, New York.

Palmer, A.R.
1969: Cambrian trilobite distribution in North America and their bearing on Cambrian paleogeography of Newfoundland; in North Atlantic---Geology and Continental Drift, editor M. Kay; American Association of Petroleum Geologists, Memoir 12, p. 139-144.

Poliscuk, V.E. and Cumming, L.M.
1977: Geology of the Strait of Belle Isle tunnel project (Abstract); Canadian Mining and Metallurgical, Bulletin, v. 72, p. 82.

Pollett, F.
1968: Peat Resources of Newfoundland; Newfoundland and Labrador Department of Mines, Agriculture and Resources, Mineral Resources Report 2, p. 23 - 30 and 158 - 168.

Pontin, T.
1967: Unpublished reports on mining exploration in the Precambrian inlier, Great Northern Peninsula, for New Jersey Zinc Co. (Canada) Ltd.

Poole, W.H.
1967: Tectonic evolution of Appalachian region of Canada; in Geology of the Atlantic Region, editors E.R.W. Neale and H. Williams; Geological Association of Canada, Special Paper 4, p. 9 - 51.

Poole, W.H. and Rodgers, J.
1972: Appalachian geotectonic elements of the Atlantic Provinces and southern Quebec; 24th International Geological Congress, Montreal 1972, Guidebook to Excursion A63 and C63, 200 p.

Poole, W.H., Sanford, B.V., Williams, H. and Kelley, D.G.
1970: Geology of southeastern Canada: Chapter VI in Geology and Economic Minerals of Canada, editor R.J.W. Douglas; Geological Survey of Canada, Economic Geology Report 1, 5th ed., p. 228 - 304.

Pouba, Z.
1967: Several hypotheses on the origin of a common fault net of the earth's crust; Czech casopis pro mineralogié a geologii, v. 1, p. 83 - 90.

Pringle, I.R., Miller, J.A. and Warrell, D.M.
1971: Radiometric age determinations from the Long Range Mountains, Newfoundland; Canadian Journal of Earth Sciences, v. 8, p. 1325 - 1330.

Raymond, P.E.
1925: Some trilobites of the lower Middle Ordovician of eastern North America; Museum of Comparative Zoology, Harvard, Bulletin 67, 180 p., 10 plates.

Richards, H.G.
1937: Pleistocene fossils from Newfoundland collected by expeditions from Princeton University; American Midland Naturalist, v. 18, p. 457 - 459.
1940: Marine Pleistocene fossils from Newfoundland; Geological Society of America, Bulletin, v. 51, p. 1761 - 1788.
1962: Studies on the marine Pleistocene Pt. 1, The marine Pleistocene of the Americas and Europe; American Philosophical Society, Transactions, new series, v. 52, pt. 3, 141 p.

Rodgers, J.
1968: The eastern edge of the North American continent during the Cambrian and early Ordovician; in Studies of Appalachian Geology; Northern and Maritime editors E-an Zen, W.S. White, J.B. Hadley and J.B. Thompson Jr.; New York, Interscience Publications, p. 141 - 149.
1970: The tectonics of the Appalachians; Wiley-Interscience, p. 211 - 224.
1972: Latest Precambrian (post-Grenville) rocks of the Appalachian region; American Journal of Science, v. 272, p. 507 - 520.

Rodgers, J. and Neale, E.R.W.
1963: Possible "Taconic" Klippen in western Newfoundland; American Journal of Science, v. 61, p. 713 - 730.

Ross, R.J., Jr. and Ingham, J.K.
1970: Distribution of the Toquima-Table Head (Middle Ordovician Whiterock) faunal realm in the Northern Hemisphere; Geological Society of America, Bulletin, v. 81, p. 393 - 408.

Rozanov, A.Y. and Debrenne, F.
1974: Age of archaeocyathid assemblages; American Journal of Science, v. 274, p. 833 - 848.

Sanford, B.V., Grant, A.C., Wade, J.A. and Barss, M.S.
1979: Geology of eastern Canada and adjacent areas; Geological Survey of Canada, Map 1401A, scale 1: 2 000 000.

Sangster, D.F.
1968: Some chemical features of lead-zinc deposits in carbonate rocks; Geological Survey of Canada, Paper 68-39, 17 p.

Scary, E.R.
1959: Toponymy of the Island of Newfoundland - Checklist no. 2, names, The Northern Peninsula; Memorial University of Newfoundland.

Schenk, P.E.
1971: Southeastern Atlantic Canada, northeastern Africa, and continental drift; Canadian Journal of Earth Science, v. 8, p. 1218 - 1251.

Schuchert, C. and Dunbar, C.O.
1921: Stratigraphy and diastrophism of western Newfoundland (Abstract); Geological Society of America, Bulletin, v. 32, p. 38, 39.
1934: Stratigraphy of western Newfoundland; Geological Society of America, Memoir 1, 123 p.

Selwyn, A.R.C.
1890: Summary reports of the operations of the Geological Survey of the year 1889; Geological Survey of Canada, Annual Report (new series), v. 4.

Shearer, J.M.
1973: Bedrock and surficial geology of the northern Gulf of St. Lawrence as interpreted from continuous seismic reflection profiles; in Earth Science Symposium on Offshore Eastern Canada, editor P.J. Hood; Geological Survey of Canada, Paper 71-23, p. 285 - 303.

Silver, L.T. and Lumbers, S.T.
1965: Geochronological studies in the Bancroft-Madoc area of the Grenville Province, Ontario, Canada (Abstract); Abstracts of Program 1965, annual meeting, Geological Society of America, p. 153.

Skelton, R.A. and Tooley, R.V.
1967: The marine surveys of James Cook in North America, 1758 - 1768; Map Collectors Circle, v. 4, no. 37, 32 p., 22 plates.

Smitheringale, W.G.
1972: Low potash Lushs Bight tholeiites: ancient oceanic crust in Newfoundland?; Canadian Journal of Earth Sciences, v. 9, p. 574 - 588.

Smyth, W.R.
1971: Stratigraphy and structure of part of the Hare Bay Allochthon, Newfoundland; Geological Association of Canada, Proceedings, v. 24, p. 47 - 57.
1973: The stratigraphy and structure of the southern part of the Hare Bay Allochthon, Northwest Newfoundland; unpublished Ph.D. thesis, Memorial University Newfoundland, 172 p.

Stevens, R.K.
1967: Great Northern Peninsula (parts of 2L, 2M, 12I, 12P); in Report of Activities, Part A: May to October, 1966, Geological Survey of Canada, Paper 67-1, pt. A, p. 186 - 187.
1968: Taconic klippen of Western Newfoundland; in Report of Activities, Part A: May to October, 1967, Geological Survey of Canada, Paper 68-1, pt. A, p. 8 - 10
1969: Flysch sedimentation and Ordovician tectonics in West Newfoundland (Abstract); Geological Association of Canada, Annual Meeting, Montreal, p. 51.
1970: Cambro-Ordovician flysch sedimentation and tectonics in west Newfoundland and their possible bearing on a Proto-Atlantic Ocean; in Flysch Sedimentology in North America, editor J. Lajoie; Geological Association of Canada, Special Paper 7, p. 165 - 177.

Stevens, R.K. and Williams, H.
1973: The emplacement of the Humber Arm Allochthon, western Newfoundland (Abstract); Geological Society of America, Northeast Section, Annual Meeting, Program, Allentown, Pennsylvania, p. 22.

Strong, D.F.
1974: Plateau lavas and diabase dykes of northwestern Newfoundland; Geological Magazine, v. 3, p. 501 - 514.

Strong, D.F. and Williams, H.
1972: Early Paleozoic flood basalts of northwestern Newfoundland, their petrology and tectonic significance; Geological Association of Canada, v. 24, p. 43 - 54.

Stukas, V. and Reynolds, P.H.
1974: $^{40}Ar/^{39}Ar$ dating of the Long Range dikes, Newfoundland; Earth and Planetary Science Letters, v. 22, p. 256 - 266.

Swett, K. and Smit, D.E.
1972: Cambro-Ordovician shelf sedimentation of western Newfoundland, northwest Scotland and central East Greenland; 24th International Geological Congress, Montreal, 1972, Section 6, p. 33 - 41.

Tanner, V.
1947: Outlines of the geography, life and customs of Newfoundland and Labrador; Cambridge University Press, 2 volumes.

Tasch, P.
1973: Paleobiology of the invertebrates; Wiley and Sons, Toronto, p. 99 - 105.

Taylor, T.G.
1946: Newfoundland---a study of settlement with maps and illustrations; Canadian Institute of International Affairs, Special Serial, 34 p.

Thomas, D.E.
1960: The zonal distribution of Australian graptolites; Royal Society of New South Wales, Journal and Proceedings, v. 94, p. 1 - 58.

Tuck, J.A.
1971: Aboriginal inhabitants of Newfoundland's Great Northern Peninsula; in The Great Northern Peninsula, an Archeological and Geological History, compiler G.F. Draskoy, Newfoundland Department of Tourism, Park Interpretation Publication 3, p. 7 - 15.

Tuke, M.F.
1966: The Lower Paleozoic rocks and Klippen of the Pistolet Bay area, northern Newfoundland; unpublished Ph.D. thesis, University of Ottawa, 142 p.
1968: Autochthonous and allochthonous rocks in the Pistolet Bay area in northernmost Newfoundland; Canadian Journal of Earth Sciences, v. 5, p. 501 - 513.

Twenhofel, W.H.
1912: Physiography of Newfoundland; American Journal of Science, v. 33, p. 1 - 24.

Twenhofel, W.H. and MacClintock, P.
1940a: Surface of Newfoundland; Geological Society of America, Bulletin, v. 51, pt. 2, p. 1665 - 1728.
1940b: Wisconsin glaciation of Newfoundland; Geological Society of America, Bulletin, v. 51, pt. 2, p. 1729 - 1756.

Wade, J.A., Grant, A.C., Sanford, B.V. and Barss, M.S., compilers
1977: Basement structure, eastern Canada; Geological Survey of Canada, Map 1400A, scale 1:2 000 000.

Walcott, C.D.
1912: Cambrian brachipoda; U.S. Geological Survey, monograph, v. 51, pt. 1, 872 p.

Wanless, R.K., Stevens, R.D. Lachance, G.R. and Rimsaite, J.Y.H.
1965: Age determination and geological studies, K-Ar isotopic ages, report 5; Geological Survey of Canada, Paper 64-17, pt. 1.
1966: Age determinations and geological studies, K-Ar isotopic ages, report 6; Geological Survey of Canada, Paper 65-17.

Wanless, R.K., Stevens, R.D., Lachance, G.R. and Edmonds, C.M.
1968: Age determinations and geological studies, K-Ar isotopic ages, report 8; Geological Survey of Canada, Paper 67-2. pt. A.

Wanless, R.K., Stevens, R.D., Lachance, G.R. and Delabio, R.N.
1973: Age determinations and geological studies, K-Ar isotopic ages, report 11; Geological Survey of Canada, Paper 73-2.
1974: Age determinations and geological studies, K-Ar isotopic ages, report 12; Geological Survey of Canada, Paper 74-2.

Weaver, D.F.
1967: A geological interpretation of the Bouguer anomaly field of Newfoundland; Dominion Observatory Canada, v. 25, no. 5, p. 228 - 250.

Weeks, L.J., compiler
1955: Geological map of the Island of Newfoundland; Geological Survey of Canada, Map 1043A, scale 1 inch to 12 miles.

Weston, T.C.
1895: Notes on concretions found in Canadian rocks; Transactions of the Nova Scotia Institute of Science, v. 9, p. 1 - 9.

White, A.S.
1914: Newfoundland---a study in regional geography; Magazine of Scotish Geographical Society, v. 30, p. 113 - 127.

Whittington, H.B.
1965: Trilobites of the Ordovician Table Head Formation, western Newfoundland; Harvard University, Museum of Comparative Zoology, Bulletin, v. 132, p. 275 - 441.
1968: Zonation and correlation of Canadian and Early Mohawkian Series; in Studies of Appalachian Geology, Northern and Maritime, editors E-an Zen, W.S. White, and J.B. Hadley; Interscience Publications, New York, p. 49 - 60.

Whittington, H.B. and Kindle, C.H.
1963: Middle Ordovician Table Head Formation, western Newfoundland; Geological Society of America, Bulletin, v. 74, p. 745 - 758.
1966: Middle Cambrian strata at the Strait of Belle Isle, Newfoundland, Canada (Abstract); Geological Society of America, Northeastern Section, Program 1966 Annual Meeting, p. 46.
1969: Cambrian and Ordovician stratigraphy of western Newfoundland; in North Atlantic---Geology and Continental Drift, editor M. Kay; American Association of Petroleum Geologists, Memoir 12, p. 655 - 664.

Williams, E.P.
1972: Quebec and Maritimes boast five potential petroleum areas; Oilweek, May 15, 1972, p. 20 - 22.
1974: Geology and petroleum possibilities in and around Gulf of St. Lawrence; American Association of Petroleum Geologists, Bulletin, v. 58, p. 1137 - 1155.

Williams, H., compiler
1967: Geological map, Island of Newfoundland; Geological Survey of Canada, Map 1231A, scale 1:1 000 000.
1971: Mafic-ultramafic complexes in western Newfoundland Appalachians and the evidence for their transportation: a review and interim report; Geological Association of Canada, Proceedings, v. 24, p. 9 - 25.
1973: Bay of Islands map area, Newfoundland; Geological Survey of Canada, Paper 72-34, 7 p.
1975: Structural succession, nomenclature, and interpretation of transported rocks in western Newfoundland; Canadian Journal of Earth Sciences, v. 12, p. 1874 - 1894.

Williams, H., Kennedy, M.J. and Neale, E.R.W., Co-ordinators
1972: The Appalachian structural province; in Variations in Tectonic Styles in Canada, editors R.A. Price and R.J.W. Douglas; Geological Associations of Canada, Special Paper 11, p. 181 - 261.

Williams, H. and Malpas, J.
1972: Sheeted dikes and brecciated dike rocks within transported igneous complexes, Bay of Island, western Newfoundland; Canadian Journal of Earth Sciences, v. 9, p. 1216 - 1229.

Williams, H. and Smyth, W.R.
1973: Metamorphic aureoles beneath ophiolite suites and alpine peridotites: Tectonic implications with West Newfoundland examples; American Journal of Science, v. 273, p. 594 - 621.
1974: The Hare Bay Allochthon, northern Newfoundland; in Report of Activities, Part A, Geological Survey of Canada, Paper 74-1A, p. 3 - 6.

Williams, H., Smyth, W.R. and Stevens, R.K.
1973: Hare Bay Allochthon, Northern Newfoundland; in Report of Activities, Part A, Geological Survey of Canada, Paper 73-1A, p. 8 - 14.

Williams, H. and Stevens, R.K.
1969: Geology of Belle Isle---northern extremity of the deformed Appalachian miogeosynclinal belt; Canadian Journal of Earth Sciences, v. 6, p. 1145 - 1157.
1974: The ancient continental margin of eastern North America: in The Geology of Continental Margins, editors C.A. Burk and C.L. Drake; Springer-Verlag, New York, p. 781 - 796.

Wilson, J. Tuzo
1962: Cabot fault, an Appalachian equivalent of the San Andreas and Great Glen faults and some implications for continental displacement; Nature, v. 195, no. 4837, p. 135 - 138.

Winkler, H.G.F.
1967: Petrogenesis of metamorphic rocks; Springer-Verlag, New York, 220 p.

Wolfe, P.E.
1952: Land exploration and land planning through geological methods in Newfoundland; International Geological Congress, Great Britain, 1948, Report 18th Session, pt. 13, Proceedings Section M, p. 256 - 263.

Woodard, H.H.
1957: Geology of Port Saunders-Castor River area, west coast; Geological Survey of Newfoundland, Report 10, 45 p.

Wynne-Edwards, H.R.
1969: Tectonic overprinting in the Grenville Province, southwestern Quebec; Geological Association of Canada, Special Paper 5, p. 163 - 182.
1971: Plutonites, gneisses and granulites of the granulite facies; Freiberger Forschungsheft, C 268 Geologie, p. 11-24.
1972: The Grenville Province; in Variations in Tectonic Styles in Canada, editors R.A. Price and R.J.W. Douglas; Geological Association of Canada, Special Paper 11, p. 263 - 334.

Zaslow, M.
1975: Reading the rocks; Macmillan Co. Toronto, 599 p.

Zeitz, I. and Zen, E-an
1973: Northern Appalachians; Geotimes, v. 18, no. 2, p. 24 - 28.

Zen, E-an
1972: The Taconide Zone and the Taconic Orogeny in the western part of the northern Appalachian Orogen; Geological Society of America, Special Paper 135, 72 p.

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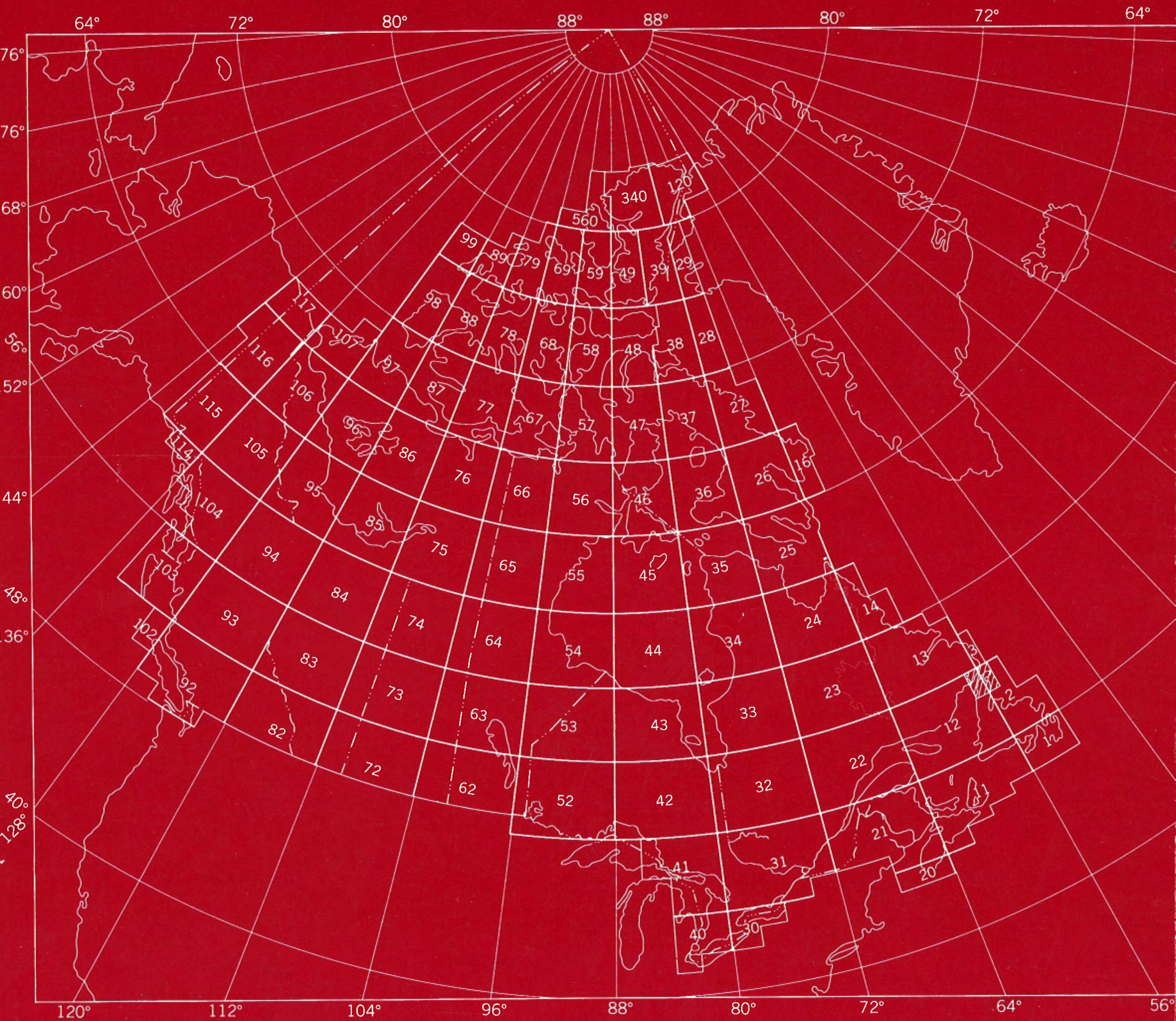

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