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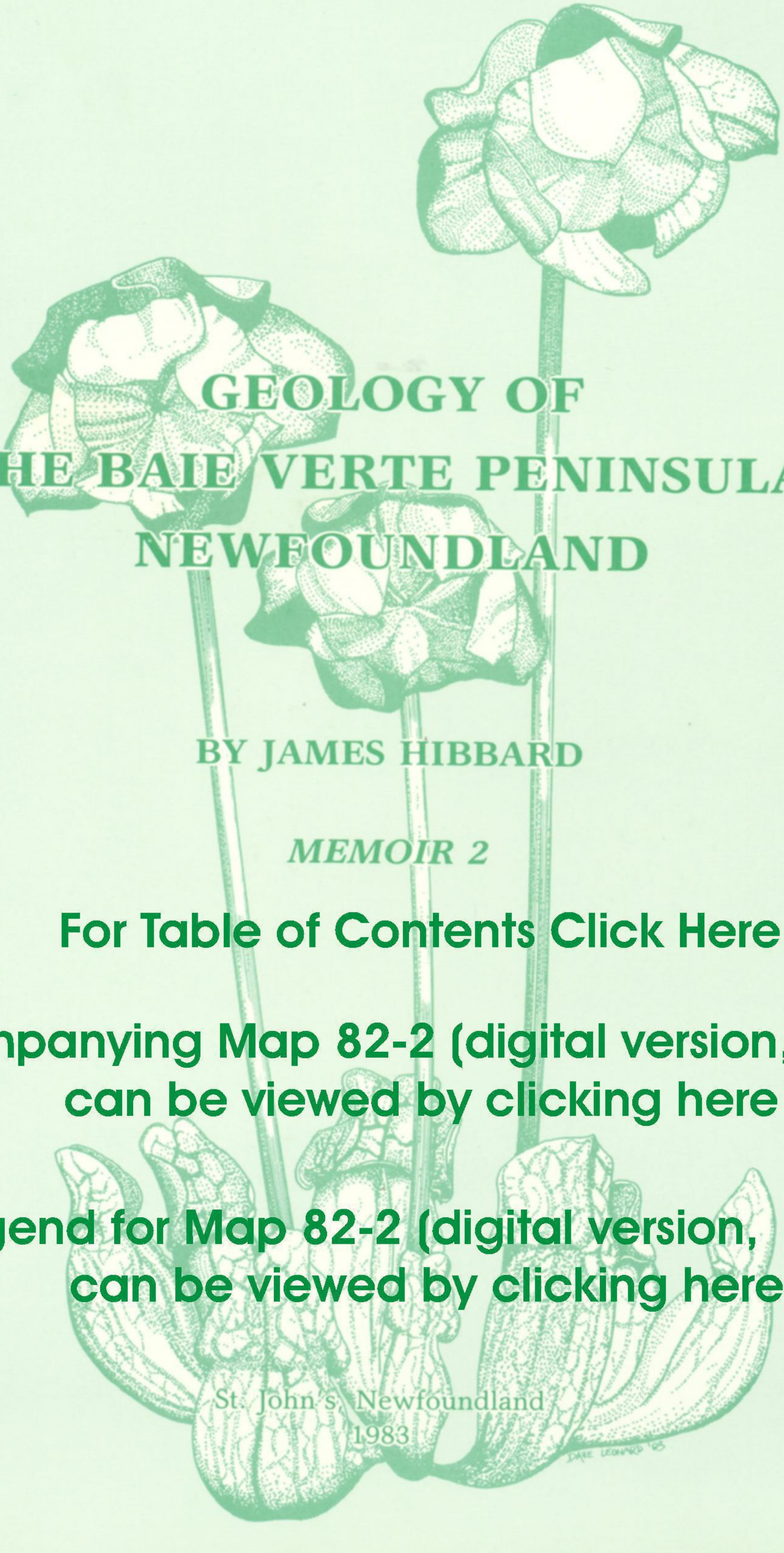

GEOLOGY OF THE BAIE VERTE PENINSULA, NEWFOUNDLAND

BY JAMES HIBBARD

MEMOIR 2

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Accompanying Map 82-2 (digital version, 17.8 mb)
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Legend for Map 82-2 (digital version, 3.5mb)
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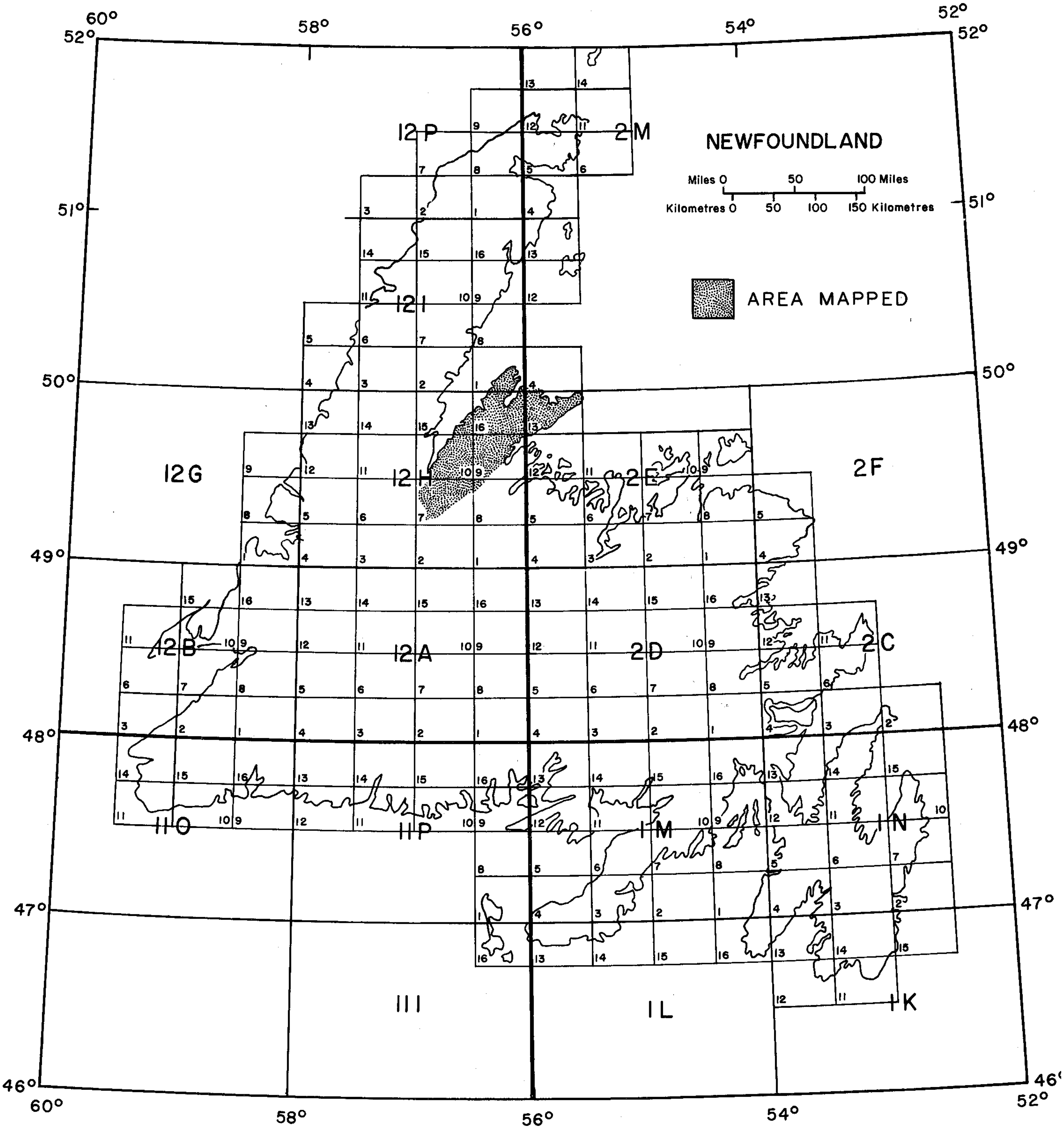
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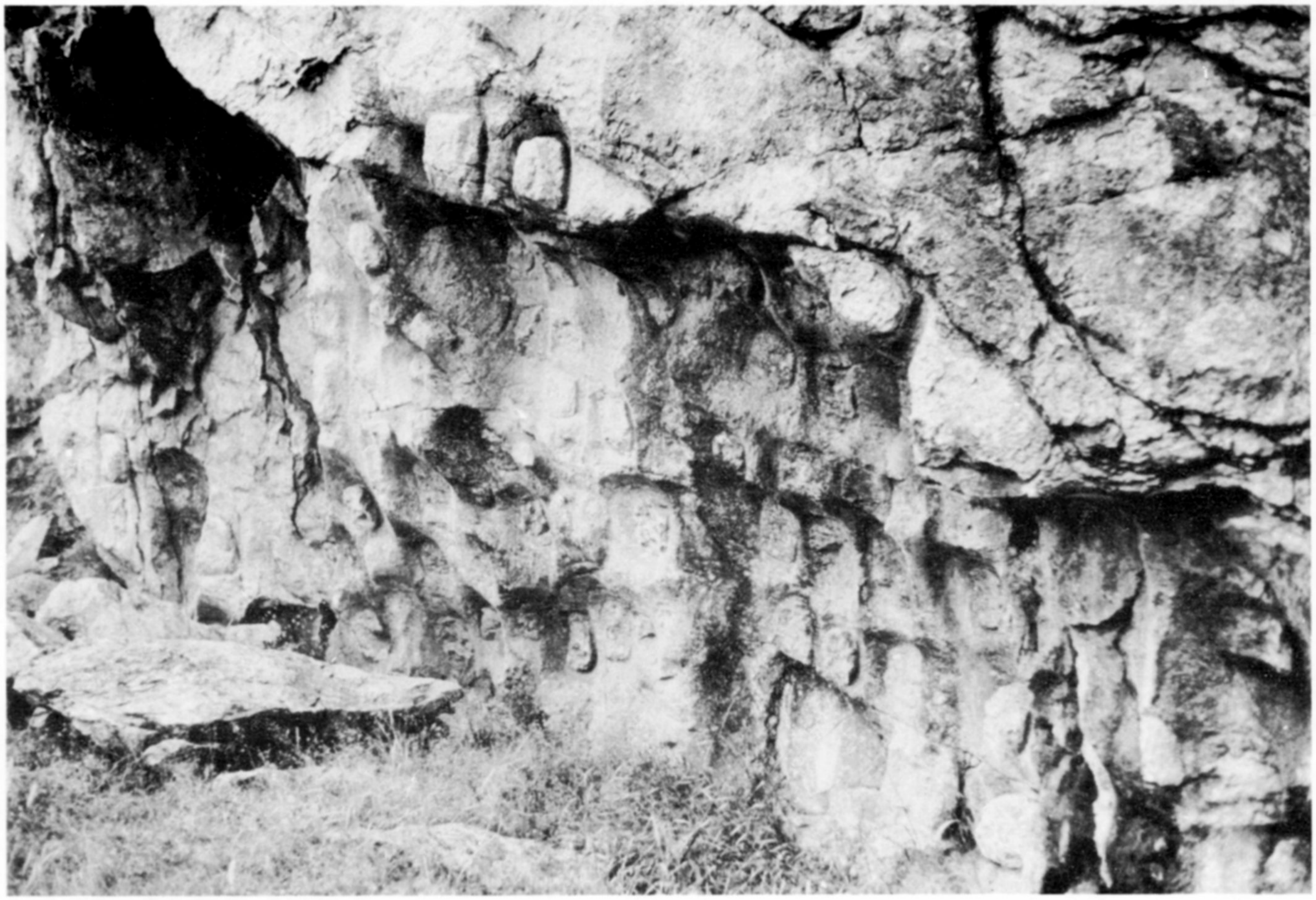

Frontispiece: *Aboriginal soapstone quarry in the town of Fleur de Lys - the first evidence of mineral exploitation on the Baie Verte Peninsula. The quarry is attributed to the Dorset people, who produced more than 1500 soapstone vessels from the pit (Nagle, 1982). Removal scars are approximately 50 cm square.*

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Memoir 2

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James Hibbard

St. John's, Newfoundland
1983

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The cover illustration is of a Pitcher Plant, the floral emblem of Newfoundland and Labrador. It was drawn by Dave Leonard.

PREFACE

The Baie Verte Peninsula lies at the northern end of the Appalachian Orogen, along the north coast of Newfoundland. The peninsula has been a center of mineral exploration and production since 1864. In more recent years it has received a great deal of academic attention as well, since it offers a well exposed section through the contact between the Paleozoic margin of North America and the ancient oceanic domain of Iapetus, a testing ground for the application of plate tectonic theories to ancient orogens.

The Mineral Development Division began regional metallogenic investigations on the peninsula in 1974, with the objective of determining the geological setting of the numerous mineral occurrences there. The initial work indicated that updated regional geological mapping was required before meaningful metallogenic assessment could be made, since much of the previous work had concentrated in specific local areas and no coherent regional geological picture was apparent. The present study was begun in 1977 with the objective of presenting an updated, comprehensive account of the geology of the peninsula and of relating the mineral occurrences to that geological framework. The memoir is based on six years of geological mapping, detailed studies of mineral occurrences, geochemical and geochronological studies, and compilation of all previous work.

This memoir presents, for the first time, an integrated account of the geology and mineral potential of an area which is perhaps the most studied in Newfoundland. The long history of stratigraphic nomenclature has been reviewed and revised, conflicting geological interpretations have been resolved, and new targets for mineral exploration have been identified. The definitive regional geological picture presented forms a basis from which more detailed studies, both academic and economic, can proceed.

B.A. Greene
Director
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ERRATA

Mineral Development Division
Newfoundland Department of Mines and Energy
Memoir 2

Geology of
the Baie Verte Peninsula,
Newfoundland
by
James Hibbard

Page 27 - the following paragraph should be added to de Wit's description:

The clasts are of a varied nature, and the following were observed: vein quartz-quartzite, banded quartzite, psammite, pelite, schist, epidosite, granite, gneiss, granitic gneiss, augen gneiss, migmatite and amphibolite.

Page 42 - the following paragraph should follow the first paragraph on the page:

In addition, the White Bay Group is linked to platformal sequences to the west by lithic correlation of carbonate breccias. The marble breccias of the group are identical in form to carbonate breccias of the Cow Head Group (Neale and Nash, 1963; de Wit, 1972; Bursnall and de Wit, 1975). This correlation suggests an upper age limit for the White Bay breccias of Llanvirnian (de Wit, 1972; Bursnall and de Wit, 1975).

Page 70 - line 6 from top of second column should read:

two-mica plutons, yield K/Ar <u>mica</u> ages of 368 ± 16 Ma

Page 189 - Table 7-6 should have the description of the Grenvillian Orogeny (D_B) under the heading "Western Orthotectonic Block" instead of under "Transition Blocks".

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ABSTRACT

The Baie Verte Peninsula is situated at the northern terminus of the Appalachian Orogen, along the rugged north coast of Newfoundland. This report is a compendium of significant geological work on the peninsula; it incorporates six years of original work and compiles all pre-existing geological data on the peninsula in an effort to clarify the geological setting of its many significant mineral occurrences.

The peninsula is bisected into contrasting lithic terranes by a north-northeast to east trending steep structural zone termed the Baie Verte Line. To the west and north of the line lies an arcuate belt of Helikian(?) to Lower Paleozoic, continentally derived, polydeformed schists and gneisses and granitoid intrusions termed the Fleur de Lys Belt. To the north, the belt is largely submerged beneath the Atlantic Ocean. The southeastern portion of the peninsula comprises Lower Paleozoic ophiolite suites, volcanic cover sequences, and various intrusions that collectively form the Baie Verte Belt.

The Fleur de Lys Belt represents the eastern margin of the Late Hadrynian to Early Paleozoic North American continent. It comprises three major lithic elements, including an infrastructure of schists and gneisses, a dominantly metaclastic schist cover sequence, and postkinematic granitoids; a fourth minor division of clastic rocks is confined to Granby Island in White Bay. The infrastructure, termed the East Pond Metamorphic Suite, comprises intensely deformed metaclastic rocks that include eclogitic amphibolite pods and small anatectic zones and that surround small windows of gneiss and migmatite. Locally, a metaconglomerate unit, containing clasts of predeformed gneiss, surrounds one of the windows of migmatite. The gneisses and migmatites are interpreted as windows of reworked Grenvillian(?) basement within intensely deformed Hadrynian to Lower Paleozoic supracrustal rocks. The suite is separated from less intensely deformed cover rocks by a steep tectonic zone of coarse mica schists.

The cover sequence, named the Fleur de Lys Supergroup, consists of metaclastic schists, marble, amphibolite and greenschist that are interpreted as tectonized submarine slope and basin deposits. Based on geochemistry and distribution, amphibolite pods and layers throughout the sequence are interpreted as being related to a rift episode responsible for the formation of the slope-basin environment. Regional correlation with rocks outside the map area, and a single fossil occurrence in marble of the supergroup, indicate the cover sequence to be Late Hadrynian to Early Ordovician in age. The cover sequence was deposited on continental basement in the west and central parts of the belt; however, to the east it interfingers with ophiolitic rocks included in the supergroup, indicating that the cover spans the junction of continental and oceanic crust. Both the infrastructure and cover sequence are intruded by postkinematic granitoids.

The Baie Verte Belt represents the westernmost vestiges of the Early Paleozoic Iapetus Ocean. It encompasses three major components, including ophiolitic suites, volcanic cover sequences, and a variety of intrusions; a minor patch of Carboniferous sedimentary rock is exposed at the southern end of the belt. On the basis of geographic position and structural state, four ophiolitic units are distinguished on the peninsula; they are considered to be mutually correlative and to represent remnants of a single slab of oceanic crust. The tightest age constraint on the ophiolites is on the easternmost suite which has been radiometrically dated as Early Ordovician and is immediately overlain by fossiliferous Arenig strata. The westernmost ophiolite is apparently stripped of its pillow lava member and overlain unconformably by a volcanic cover sequence; this ophiolite also has a dynamothermal aureole at its base. The pillow lava members of the three easterly suites contain unusually high-magnesian lavas that are progressively more magnesian from the westerly to the easterly suites. These pillow lava members are conformably overlain by volcanic cover sequences.

The volcanic cover sequences consist of two major divisions that are separated by an unconformity at two locales. The lower division is dominated by mafic submarine volcanic products and directly overlies the ophiolites. These cover rocks are probably Middle Ordovician and older in age, and at one locale contain Arenig graptolites. The upper division is characterized by mainly subaerial felsic volcanic and associated rocks that unconformably overlie the lower division. The upper division is considered to be Siluro-Devonian in age based on radiometric dates and regional correlation with units outside the map area.

Intrusive rocks of the belt are readily separable into two suites of different ages. The earlier suite is Early Ordovician in age and includes mainly granodioritic and granitic rocks whereas the later plutons are Siluro-Devonian in age and exhibit a wide range of composition.

All pre-Carboniferous strata and structures on the peninsula, including the Baie Verte Line, are folded around a major structure, the Baie Verte Flexure. It is defined by the change in structural trends from north-northeasterly on the southern portion of the peninsula to easterly on the northern portion. The flexure appears to be a primordial structure that predates the tectonism of rocks on the peninsula and is considered to reflect the original shape of

the ancient North American margin; younger structures have apparently been molded to its form. The tectonic history of the peninsula is largely controlled by the juxtaposition and interaction of the two lithostratigraphic belts along this irregular margin.

The entire Fleur de Ly Belt and the northern portion of the Baie Verte Belt display three main phases of deformation and exhibit upper greenschist to middle amphibolite facies regional metamorphism. The remainder of the Baie Verte Belt was affected by a single penetrative fabric and up to lower greenschist facies metamorphism. Radiometric cooling dates on metamorphic minerals indicate that Fleur de Lys rocks on the west limb of the Baie Verte Flexure were subjected to a Taconic event, whereas rocks of both belts along the eastern limb of the flexure record Acadian tectonism. The Taconic event is attributed to the westward regional obduction of ophiolites over the Fleur de Lys Belt; it appears to have affected the whole belt; however, most evidence of its affecting Fleur de Lys rocks on the east limb of the flexure has been obliterated by the later Acadian event. The Acadian event may have been related to a reversal of the Taconic structural polarity along the Baie Verte Line during uplift of the metamorphic Fleur de Lys Belt.

The structural juxtaposition of the Fleur de Lys and Baie Verte Belts has concentrated many environments favorable to mineralization in the small area of the peninsula. The peninsula has supported nine mines since 1864, eight for base and precious metals and one for asbestos. All of the mines and many major showings are associated with ophiolitic rocks of the peninsula. Two broad categories of sulfide mineralization are associated with the ophiolites and are informally termed the Betts Cove and Rambler types. The Betts Cove deposits are mineralogically simple (pyrite-chalcopyrite-pyrrhotite) and occur as stockwork and massive strata-bound zones in the sheeted dike and pillow lava members; in contrast, the Rambler types are mineralogically more complex and are situated either near the top or immediately above the ophiolite in mafic and felsic tuffs. Both types appear to be spatially related to the high magnesian basalts of the ophiolites. Other notable deposits associated with these rocks include gold mineralization in an iron formation above one of the ophiolites and asbestos fiber in the ultramafic member of the westernmost ophiolite. Significantly the highest grade asbestos occurs in an ultramafic body in the nose of the Baie Verte Flexure; it appears that the fracture pattern vital for asbestos deposits is best developed in this structurally complex zone. Mineral occurrences are found in most of the major lithostratigraphic divisions of the area; however, attention is now focussed on nontraditional targets in the metamorphic cover rocks of the Fleur de Lys Belt. In particular, marble in the belt is host to significant copper mineralization, and metaclastic cover rocks have potential for massive sulfide and barite deposits, based on the correlation of these rocks with other mineralized units in the Appalachian-Caledonian Orogen.

Chapter I
INTRODUCTION

Preamble

The Baie Verte Peninsula, formerly called the Burlington Peninsula, is located in northwest-central Newfoundland between 49°15′ and 50°10′ north latitude and 55°20′ and 57°00′ east longitude (Figure 1-1). It forms a roughly triangular projection of land that is bounded to the east by Green Bay and Notre Dame Bay, to the west by White Bay, and to the north by the Atlantic Ocean. The southern boundary of the peninsula is here taken to be the broad valley that links Birchy Lake with the Southwest Arm of Green Bay, and its southwestern boundary is marked by a less incised valley between Sandy Lake and White Bay. The Baie Verte Peninsula embodies three smaller peninsulas, including the Fleur de Lys Peninsula between White Bay and Baie Verte, the Point Rousse Peninsula between Baie Verte and Ming's Bight, and the Cape St. John Peninsula, encompassing the area east of the longitude of Burlington. The Horse Islands, immediately north of the peninsula, and Granby Island in White Bay are included in the present study area.

Geological accounts of the peninsula date back to 1867, and in recent years have flooded the geological literature of Newfoundland. Interest in the geology of the area has ultimately stemmed from its rich mineral potential. The peninsula has long been a center of mineral exploration and production, and rocks of the area are host to two mines active at the time of this writing (1980-1981). In the past decade, the region has also been in the academic forefront, as it has served as a testing ground for the application of the plate tectonics hypothesis to ancient orogens.

It is timely, now, to compile and integrate this accumulated body of geological data. Previous workers outlined the geology of many separate areas on the peninsula with little regard for the intervening terrain. This led to contradictions in regional tectonic-stratigraphic syntheses and, hence, ambiguous interpretation of the geological setting of many mineral occurrences. It is the intention of the present study to provide a consistent geological overview to which mineral deposits of the peninsula can be more successfully related.

Scope and Format

This report is intended to be a compendium of significant geological work on the Baie Verte Peninsula. It incorporates six years of work and compiles all pre-existing geological data on the peninsula. The study originated as a regional metallogenic investigation of the Baie Verte Peninsula initiated by J. DeGrace, in 1974, on the Nippers Harbour map sheet. During the early phase of that study it was found that updated regional scale mapping of the peninsula was necessary before an accurate economic assessment could be made. The project was suspended in 1976, and resumed under my supervision in 1977.

It has resulted in the publication of five 1:50,000 scale geological maps, including the following:

(i) Nippers Harbour/Cape St. John/Horse Islands south (2E/13, 2E/14, 2L/4S) (DeGrace et al., 1976)

(ii) Baie Verte/Jackson's Arm east (12H/16, 12H/15E) (Hibbard and Gagnon, 1980)

(iii) Hampden east/King's Point west (12H/11E, 12H/10W) (Hibbard et al., 1980a)

(iv) Sheffield Lake north (12H/7N) (Hibbard et al., 1980b)

(v) Fleur de Lys/Horse Islands (12I/1, 2L/4) (Bursnall and Hibbard, 1980)

These map sheets have been reduced and included on the 1:100,000 scale colored geological compilation map of the peninsula accompanying this report (Figure 1-1).

Detailed economic geology, geochemical, and geochronological studies were undertaken in conjunction with the mapping program. J. Tuach compiled and described the economic geology of the Baie Verte and Fleur de Lys sheets (Tuach, 1978b); his work served as the basis for the economic geology section herein. Petrochemical investigation of the peninsula was initiated in this project by J. DeGrace (DeGrace et al., 1976) and continued with geochemical data compilation by D. Wilton in 1978, and further analyses and compilation by myself in more recent years. Geochemical interpretation of data from the Wild Cove Pond Igneous Suite was provided by D.F. Strong of Memorial University of Newfoundland. In addition, a geochronological study was undertaken under the supervision of R.D. Dallmeyer, University of Georgia, U.S.A.. All geochemical and geochronological data collected during the project that are previously unpublished appear in the appendices of this report; where pertinent, analytical methods are included. Rock, mineral and element abbreviations used herein are given in Appendix I. Photographs throughout this text without credits were taken by me.

This study also stems from a substantial body of previous work (Figure 1-2), and in many places relies heavily on the earlier studies, some of which were very detailed. I visited some portions of the peninsula only briefly; hence, I rely almost totally on the maps and descriptions by earlier workers for these areas (Figure 1-3). In places throughout the report, I quote descriptions by previous workers. The reason for this is two-fold, including the presentation of previous work in the context of present ideas and the handiness afforded by a complete volume of Baie Verte geology.

The maps produced by Neale et al. (1960), Upadhyay (1973), Norman (1973), Kidd (1974), Bursnall (1975), Dean and Strong (1975a,b), and Tuach (1976) were field-checked, but otherwise largely incorporated into the study with only local modifications. De Wit's (1972) work was an invaluable guide to the geologically complex area west of the Baie Verte

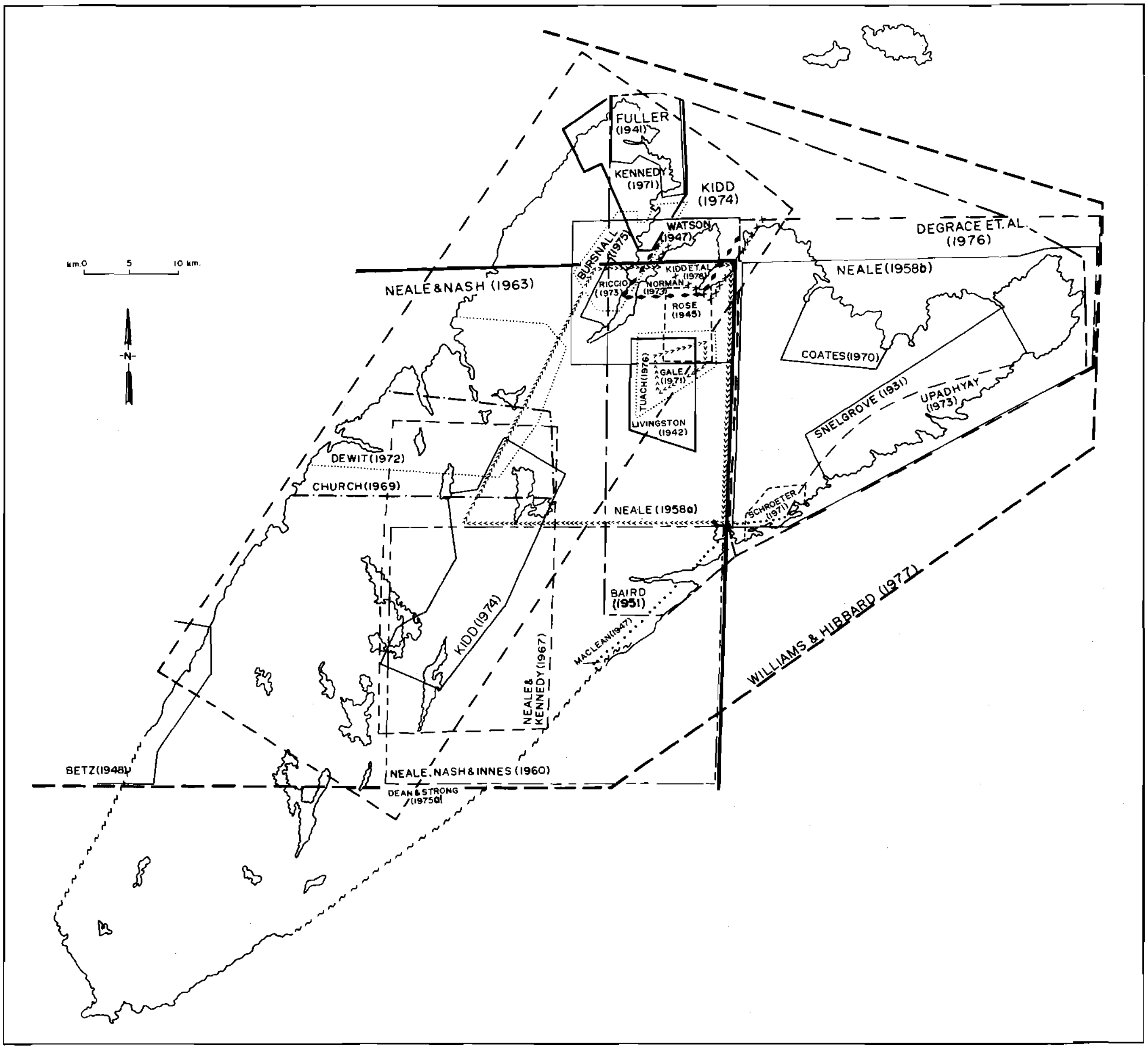

Figure 1-2: *Previous geological maps of the Baie Verte Peninsula.*

highway, though remapping of the best exposed sections and field checks of the inland area at 3 km traverse intervals led to modifications of the map pattern and reinterpretation in the area.

I have incorporated rock descriptions by previous workers either for areas which I visited only briefly or for areas previously mapped in detail. These are as follows: for the Cape St. John Peninsula, DeGrace et al. (1976) and Upadhyay (1973); for the area of Rambler Mines, Tuach (1976); for portions of the Point Rousse Peninsula, Norman (1973); for portions of the Advocate and Point Rousse Complexes, the Flat Water Pond Group, and all of the Micmac Lake Group, Kidd (1974) and Kidd et al. (1978); for the southeastern portion of the peninsula, Neale (1962) and Neale and Nash (1963); for the Fleur de Lys peninsula, Bursnall (1975, 1979, and personal communication); and for the area between Flat Water Pond and Western Arm, de Wit (1972, 1974, 1980).

During the compilation and writing of this report, I made no attempt to strike upon an equal level of coverage for the geology of the peninsula. Instead I tried to include all pertinent data available for distinguishing lithostratigraphic units, for determining their environments of deposition and interrelationships, and for determining their subsequent tectonic history. Some highly specialized studies undertaken by previous workers are abstracted; details are left for the reader to pursue.

The amount of data included in this volume compels the reader to treat it as a reference book, in which he will likely have interest in selected parts. It is recommended that Chapter III, "Regional Setting and General Geology," be read as a guide and general overview of the peninsula before reading other parts of the volume.

Geographic Setting

The community of Baie Verte, located at the head of Baie Verte, is the major population center in the area. It is accessible by paved road from the Trans Canada Highway, 65 km to the south, which forms the approximate southern boundary of the peninsula. Immediately south of Baie Verte, a paved branch of the Baie Verte highway leads eastward to La Scie. At the time of this writing, the roads to Fleur de Lys and Seal Cove, White Bay, are being upgraded and paved. All of the smaller communities on the peninsula are serviced from these main thoroughfares by upgraded dirt roads. The old, unpaved Trans Canada Highway transects the southern part of the peninsula and provides access to the area immediately north of Birchy Lake. Numerous logging roads facilitate access to many other parts of the area, though these are dry weather roads normally traversable only by 4-wheel drive vehicles. The most isolated areas, east of Upper Indian Pond, east of Wild Cove Pond, and southwest of Middle Arm, Green Bay, are accessible only by aircraft or overnight trips on foot. All of the coastline of the peninsula is accessible by small boat during the summer. Formal accommodations are available at Baie Verte, and overnight campgrounds with public facilities are available at Flatwater Provincial Park during the summer months.

Mining and woods operations constitute the most important factors in the local economy. At present, two mines are operational on the peninsula; Advocate Mines produces asbestos, and Consolidated Rambler Mines extracts copper, gold, and silver from a massive sulfide deposit. The gross value of asbestos produced at Advocate in 1977 was $34,500,000 and 541 people were employed full-time. During the same year, the production value at Consolidated Rambler Mines was approximately $13,000,000 with 190 people working full-time. An employment multiplication factor of 1.4, as recommended by provincial statisticians, suggests that at least 15% of the total population of the Baie Verte Peninsula obtain employment in mining and associated jobs.

Woods operations related to the pulp and paper industry support the second largest work force on the peninsula, though statistics are not readily available. Most of these operations are on a contract basis with Bowaters Limited, which operates a paper mill at Corner Brook. The inshore fishery, which is also a major contributor to the local economy, has suffered a decline over the past few decades, but locally appears to be undergoing revitalization; of particular note is the growing fishery in the communities of Wild Cove, White Bay, Round Harbour, and La Scie. The remainder of the economy is mainly encompassed by public works, subsistence farming, and small independent businesses, e.g. boat building at Nippers Harbour.

Topography

The Baie Verte Peninsula coastline is very rugged and highly irregular; only along White Bay, where the coastline runs parallel to the strike of the bedrock, is a smooth shore encountered, though even here towering cliffs and steep slopes prevail along the water's edge (Figure 1-1). Gentle slopes to the sea occur locally within sheltered arms, e.g. Northwest Arm at Burlington (Figure 1-1). Inland, the area is part of the Atlantic uplands division of the Canadian Appalachian geomorphic region (King, 1972). These uplands are gently southeasterly dipping and deeply dissected. On the peninsula, two erosional cycles are recognized within the uplands, the High Valley peneplain (Twenhofel and MacClintock, 1940) and the Lawrence peneplain (Heyl, 1936). In the western and southeastern portions of the peninsula, most of the hills culminate at 300 to 350 m above sea level, corresponding to the High Valley peneplain. These two highland areas are separated from each other by a narrow lowland corridor through the south-central portion of the peninsula. This lowland, ranging between 175 and 200 m above sea level, represents an arm of the Lawrence peneplain, which broadens to encompass the Cape St. John Peninsula to the east. These peneplains have been related to changes in subaerial erosion resulting from rifting and formation of the modern Atlantic Ocean (King, 1972). Subsequent erosion and glaciation of the peneplains has accentuated contrasts and structural features within the underlying bedrock. As a result, the two highland areas and the lowland area each have a distinct topographic expression.

The topography of the western highland area varies from north to south, and is ultimately controlled by the bedrock. The northern and central portions of this region are underlain predominantly by variably weathering, steeply dipping schists that produce a pronounced northeasterly trending topographic grain. It is marked by broad, locally terraced ridges averaging 325 m high, and boggy valleys, in many

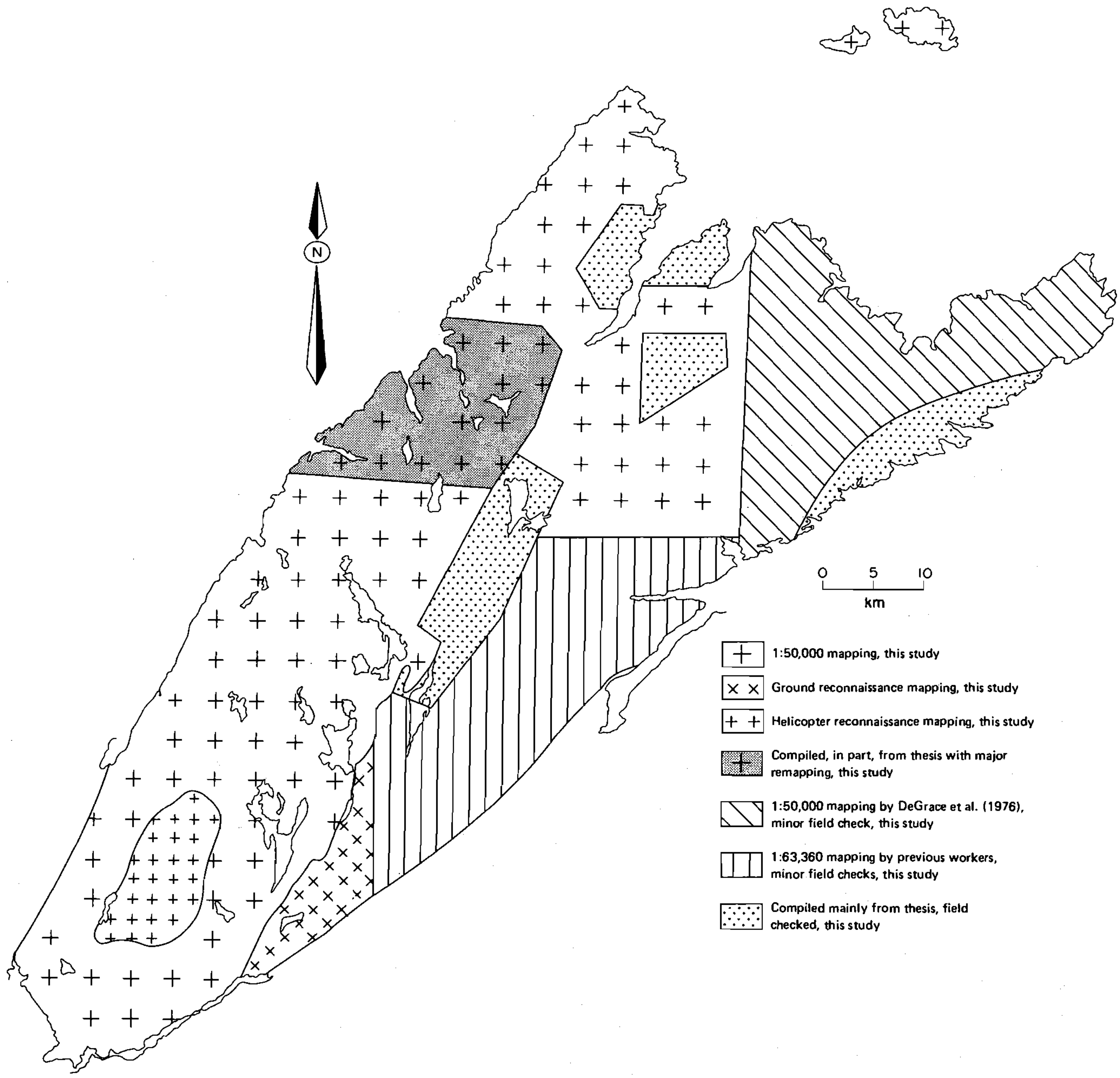

Figure 1-3: *Reliability diagram for Figure 1-1.*

places containing small linear ponds. On the exposed northern tip of the Fleur de Lys Peninsula, the altitude of the highlands diminishes and the topography is characterized by an irregular pattern of steep valleys. This pattern most likely reflects the complex fault system in the underlying bedrock. The bold Fleur de Lys Hill, which stands out in this area, is underlain by resistant amphibolite. The southern portion of the western highland area is underlain by more homogeneous weathering granitic rocks. The topography here is dominated by vast rolling boglands, a few large ponds, and broad, gently sloping ridges. The granitic terrane appears to have been passive in determining the local geomorphology; thus, the pronounced northerly trend of many of the ridges and larger ponds reflects glacial effects on the area. Locally, though, some of the prominent hills at the southern end of this area are underlain by resistant xenoliths within the granitic terrane. The southwestern portion of this region gently slopes down to a lowland about 100 m in altitude surrounding Sandy Lake, whereas the southeastern boundary is sharply delineated by a high ridge rising more than 100 m above the surrounding surface; the ridge drops off abruptly to the south into Birchy Lake. Throughout the western highlands, local topographic linears of pronounced relief trend northwesterly, transverse to the grain of the region, e.g. the Black Lake - Purbeck's Brook drainage system, the Wild Cove Pond drainage system, Middle Arm Brook, White Bay, Southern Arm, and the valley from Little Lobster Harbour to Lower Duck Island Cove Brook. These major valleys are apparently glacially scoured, and only locally, such as at Black Lake or Little Lobster Harbour where major faults trend athwart the regional grain, do they reflect bedrock structures.

The southeastern highland area is underlain mainly by felsic volcanic and intrusive rocks that impart very little character to the grain of this region. Neale (1962) reported that only local, low ridges reflect trends in the underlying pyroclastic and flow rocks. He also noted that the numerous linear valleys and depressions in the region correspond to a complex fault system within the bedrock. Otherwise, this highland is generally a rolling surface with a few irregularly distributed knobs rising above it. One of these knobs, at the headwaters of Rattling Brook, marks the highest point on the Baie Verte Peninsula.

The lower erosional surface is somewhat more homogeneous than the highlands. Throughout the peninsula, it is marked by a hummocky terrain with a more random dispersal of bogs and small ponds than that found in the highlands. Near the border with the western highland and in the valley between the highlands, ponds within the lowland are distinctly north-northeast trending, reflecting both underlying bedrock trends and the path of glacial erosion. In the area of Black Brook, the southerly extending arm of the lowland is partly blocked by two northeasterly trending ridges underlain by resistant felsic extrusives and granitoid rocks. South of there, the lowland descends approximately 150 m into the fault-controlled Birchy Lake - Green Bay Valley. In the central portion of the peninsula, above a large granodiorite body, the topography is more gentle and rolling than the lowlands to the east, and extensive bogs are developed in this area. A very rugged terrain is developed over the mafic-ultramafic bedrock found immediately south of Ming's Bight and in the Betts Cove - Tilt Cove area. The topography in these areas reflects the complex bedrock structure. A similar topography developed in granitic porphyry east of Nippers Harbour has been attributed to glacial erosion (Baird, 1951) though, again, bedrock complexities may be a contributing factor.

A few remnants of the High Valley peneplain stand out in bold relief within the lowland. These areas, including the Dunamagon Highlands west of Pacquet Harbour, the high semicircular ridge north of the Burlington road, and the Confusion Highlands, are all underlain by either granitoid or felsic extrusive rocks.

The present drainage patterns throughout the peninsula are young, being largely controlled by glacial features; however, major fault-controlled watersheds such as the Birchy Lake - Green Bay valley and the Sandy Lake - White Bay valley may be remnants of older drainage patterns.

Glacial Geology

Observations on the glacial history of the peninsula were presented by MacClintock and Twenhofel (1940), Neale and Nash (1963), Lundqvist (1965), Grant (1973), Neale (1962), and R. Ricketts (personal communication, 1981); as well, notes on glacially derived features of local areas on the peninsula were reported in numerous regional survey reports. During the present study, only casual observations were made with regard to glacial features. The following summary is largely drawn from the sources cited above, though previously unreported glacial features observed during this study are included.

Pleistocene glaciation affected the whole of the Baie Verte Peninsula; due to the freshness of till, erratics, and most ice-carved features, the glaciation is considered to be Wisconsinan by most workers. It has been suggested that, during an early stage of the Wisconsinan, Newfoundland was glaciated by ice from a Labrador center, and only later in the Wisconsinan did it support its own, independent ice cap (Flint, 1939; MacClintock and Twenhofel, 1940; Lundqvist, 1965). Meager evidence across the island supports the idea of an early Labrador ice sheet over Newfoundland. On the Baie Verte Peninsula, Lundqvist (1965) reported an old set of northwest-southeast glacial striae near Kidney Pond, and suggested that these represent this early incursion of Labrador ice; from evidence cited, it is difficult to draw any conclusions about this hypothesis. Glacial features discussed below are all presumed to be related to the last Wisconsinan glaciation.

Glacial features reported by previous workers and observed during the present study are compiled on Figure 1-4. The direction of glacial movement on the peninsula has been deduced from striae, grooves, gouges, roches moutonées, smaller scale stoss and lee forms, crag and tail features, and drumlinoid forms. They generally indicate that ice travelled northward over the peninsula, and radiated out from the central area. Neale (1962) noted that the lowland between Micmac Lake and Flat Water Pond acted as an ice divide; to the west of this corridor, ice flowed toward White Bay, whereas to the east it flowed toward Green Bay. Lundqvist (1965) noted numerous sets of intersecting striae on the peninsula; these are here considered to be the result of local irregularities in ice movement.

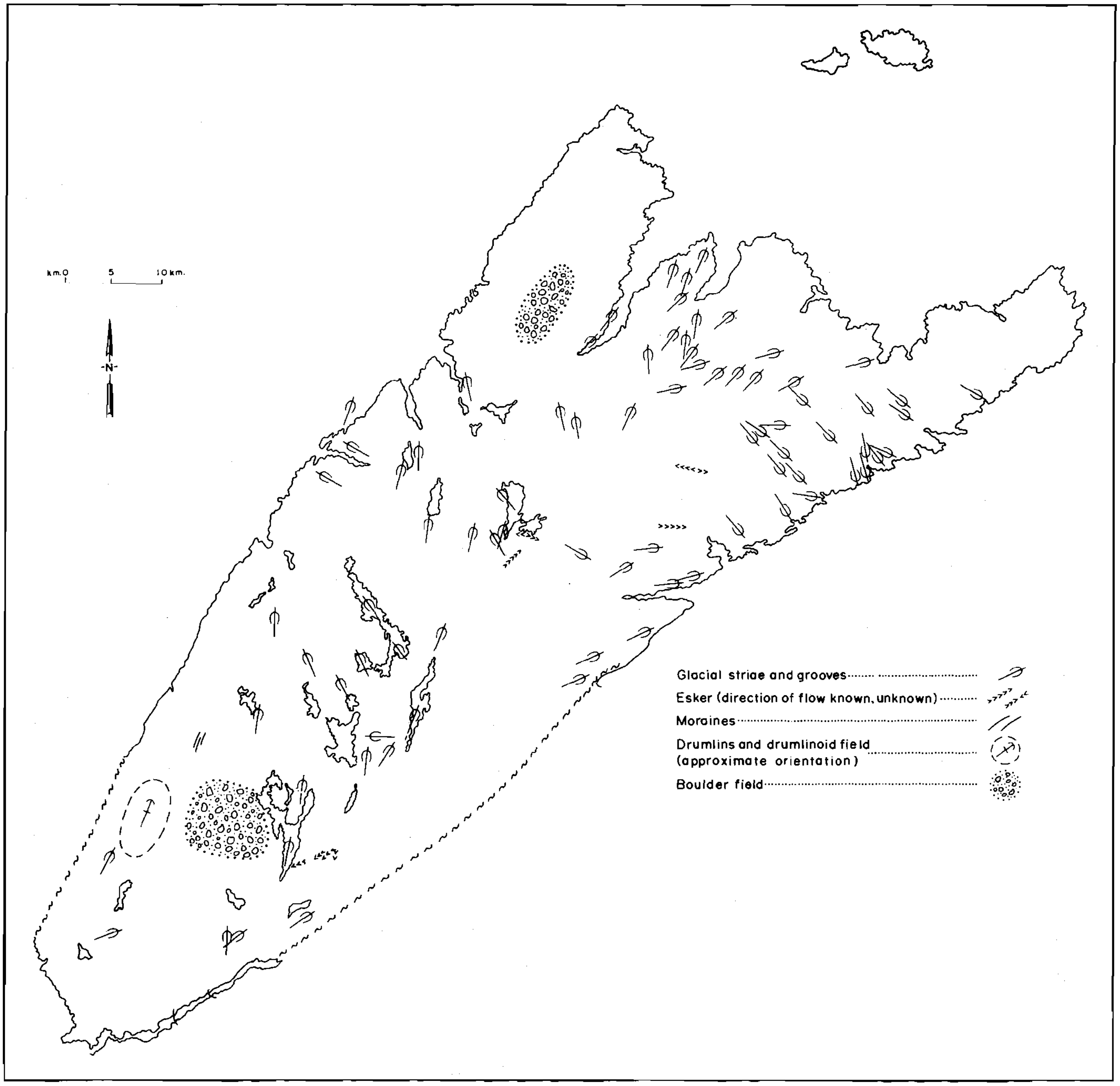

Figure 1-4: *Glacial features of the Baie Verte Peninsula (compiled from sources described in text).*

During the late Wisconsinan, receding ice deposited a veneer of till on the bedrock throughout the peninsula. The depth of this cover is difficult to assess, though it is much thicker in the lowlands than in the high areas, where in many places bedrock is exposed. Lundqvist (1965) noted the compactness of the till and suggested that it is largely lodgement till, though in the area between Sandy Lake and Flat Water Pond, he reported a hummocky covering of ablation till. Till overlying the larger granitic bodies does not appear to have travelled far, as more than 85% of observed boulders were derived from the immediately underlying plutons. In contrast, Neale (1962) and Neale and Nash (1963) reported far travelled ultramafic erratics which could only have been derived from the ultramafic belt that bisects the peninsula. The erratics occur 12 km west of the ultramafic belt in the Black Lake area and at least 25 km to the east of the belt, on Rattling Brook, Green Bay. A coarse grained anorthosite erratic near Shoal Point, Western Arm, White Bay has no immediate source terrain on the peninsula, and its origin is equivocal. Either it could have been derived by ice rafting from the Long Range Mountains on the western side of White Bay or, alternatively, it could have originated in Labrador, thus indicating the former presence of a Labrador ice sheet in Newfoundland.

Two large boulder fields were noted on the peninsula. One, just west of Baie Verte, is largely overgrown by vegetation, whereas the other, just west of Upper Indian Pond, is mostly barren. Boulders in the latter field attain house-size dimensions. Small eskers occur locally in the southern and central portions of the peninsula.

The coast of the Baie Verte Peninsula is a submerged feature, with many fiordlike indentations such as Middle Arm, White Bay, Southern Arm, Baie Verte, and Middle Arm, Green Bay. However, the occurrence of raised beaches and wave-cut benches in many places around the coast attests to some postglacial uplift of the region. The highest reported emergence feature on the peninsula is a gravel terrace at 65 m on the north shore of Seal Cove (Henderson, *in* Neale and Nash, 1963). Nearly all of the major valleys leading to the sea contain glacial outwash deposits that locally also indicate emergence of the present coast. Baird (1951) reported fossil marine pelecypods, gastropods, barnacles, and bryozoa from a clay-rich layer within terraced glacial outwash just west of the community of Middle Arm, Green Bay. Similarly, at the mouth of the Baie Verte River, approximately 7 m of till, fossiliferous marine clay and outwash occur 20 m above sea level (Murray and Howley, 1881; MacClintock and Twenhofel, 1940). Henderson (*reported in* Neale, 1962) retrieved marine fossils from other strata at this locality, at approximately 54 m above sea level.

Deep bowl-shaped depressions are common along the coastline of the peninsula, particularly on the shoreline of White Bay. Cirquelike depressions such as Downey's Cove, Lobster Harbour, Wild Cove, and Little Lobster Harbour all suggest a glacial origin, or at least glacial modifications of earlier depressions. Baird (1951) described other such depressions around the Cape St. John Peninsula and noted that their origin was ambiguous, though glaciation definitely influenced their final form. The overdeepened harbor with a shallow threshold at Lobster Harbour, White Bay, supports a cirque origin for at least some of these features.

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Chapter II

HISTORY OF GEOLOGICAL THOUGHT

The history of geological thought on the Baie Verte Peninsula has been nearly as complex as the geology itself. Throughout this history, there has been both a mainstream of regional geological interest and a myriad of accessory studies more concerned with local or specialized aspects of the geology of the peninsula. Thus, the previous investigations are presented here in two sections. The first is an overview of the growth of regional concepts and the second is a table of the accessory studies. The details of previous investigations are incorporated into succeeding chapters, where relevant.

REGIONAL CONCEPTS

The evolution of regional geological concepts on the peninsula can be viewed in two broad time periods, (1) the foundation work from 1864 to about 1965, and (2) the modern work from about 1965 to the present. The foundation work included mostly mapping projects concerned with outlining the distribution of rock types and their inter-relationships, and documenting the setting of mineral occurrences on the peninsula. The modern work has mirrored the worldwide revolution in geology introduced by the theory of plate tectonics. The detailed evolution of stratigraphic nomenclature and tectonic history for both time periods is summarized in Appendix II (Table AII-1).

FOUNDATION WORK (1864 to about 1965)

Regional investigations undertaken during this phase were directly concerned with the mineral deposits of the peninsula and their geological settings. Almost all of the early regional work was undertaken by government surveys.

The earliest recorded regional geological survey was that of Sir Alexander Murray in 1864 (Murray and Howley, 1881); Murray was lured to the peninsula by the Terra Nova Mine near Baie Verte. Undoubtedly, there were earlier unrecorded investigations, since the mine was active before that time and it is known that a local surveyor, Smith McKay, recognized copper mineralization at Tilt Cove in 1857 (see Chapter 8). During his trip, Murray made a coastal reconnaissance survey of the geology in the area and many of his regional observations are still adhered to by modern workers. In particular, he noted the antiformal distribution of regionally metamorphosed rocks now recognized as the Fleur de Lys Supergroup. He proposed a Laurentian age for rocks in the core of this structure and a Lower Paleozoic age for the flanking rocks. Murray returned to the peninsula in 1865, 1867 and 1875 to investigate the geology and workings of the Tilt Cove and Betts Cove Mines (Murray and Howley, 1881).

Regional geological mapping of the peninsula did not proceed until almost fifty years after Murray's last recorded visit. Four workers were particularly significant in the regional mapping, A.K. Snelgrove, K. de P. Watson, D.M. Baird, and E.R.W. Neale.

The first systematic mapping was undertaken on the southern part of the Cape St. John Peninsula by Snelgrove (1931) as part of a doctoral study at Princeton University. His major interest was the geological setting of copper deposits at Betts Cove and Tilt Cove.

Watson was another Princeton University student who was involved in a regional study of Newfoundland copper deposits (Douglas et al., 1940). He continued mapping in the Baie Verte - Ming's Bight area (Watson, 1942, 1943, 1947) where he was attracted by the Terra Nova and Goldenville Mines as well as the Rambler, Mud Pond, and Barry-Cunningham prospects.

The deposits at Betts Cove and Tilt Cove were later the subject of a doctoral dissertation by Baird (1947) at McGill University. Baird also mapped the northern part of the peninsula and compiled all of its known geology, in an effort to characterize the geological setting of the mineralization (Baird, 1947, 1948, 1951). Baird was largely responsible for much of the present stratigraphic nomenclature used on the peninsula.

Baird also introduced E.R.W. Neale to the geology of the peninsula. Neale methodically produced three and one-half 1:63,360 scale geological maps, including the east half of the Baie Verte sheet (Neale, 1958a), the Nippers Harbour sheet (Neale, 1958b), the Fleur de Lys sheet (Neale, 1959a), and the King's Point sheet (Neale et al., 1960; Neale, 1962). He reported on key relationships between units in the Tilt Cove area (Neale, 1957) and on the contact of Fleur de Lys rocks with easterly volcanic units of the Baie Verte Group (Neale, 1959b). In addition, Neale and Nash (1963) included most of the western half of the peninsula in a 1:250,000 scale regional mapping project. Thus, Neale was the first worker to map and compile all of the geology of the peninsula. His maps are notably accurate and have been the standard base maps for all modern workers.

By the close of the foundation phase, most major stratigraphic units were outlined and named (see Appendix II). Besides this major task, the most notable contributions to the regional geology during this period that had particular bearing on modern ideas included the following:

(i) discovery of Arenig graptolites in rocks now called the Snooks Arm Group (Snelgrove, 1931) and their correlation with greenschist on the Point Rousse Peninsula (Snelgrove, 1935);

(ii) recognition of a major fault between the Fleur de Lys metaclastic rocks and the "Baie Verte" volcanic rocks (Watson, 1947);

(iii) recognition of a major unconformity between the Lower Ordovician Snooks Arm Group and the overlying Cape St. John Group (Neale, 1957); Neale interpreted the latter to be Devonian;

(iv) recognition that the Cape St. John Group contains ultramafic detritus, yet is apparently intruded by ultramafic rocks of the present Betts Cove Complex (Neale, 1957); Neale interpreted the latter contact as a cold remobilization and injection of an old (Ordovician?) serpentinite during a post-Cape St. John period of tectonism;

(v) recognition of a nonconformity between Micmac Lake Group rocks of probable Devonian age and the Burlington Granodiorite (Neale and Nash, 1963); and

(vi) recognition of marble and marble breccia in the Fleur de Lys terrane and their correlation to carbonates in western Newfoundland, thereby indicating a Paleozoic age for at least some of the Fleur de Lys schists (Murray and Howley, 1881; Neale and Nash, 1963).

Two major uncertainties existed at the close of this period, namely, the age of the ultramafic plutonic rocks on the peninsula and the age of deformation and metamorphism of the crystalline rocks. Most modern regional concepts on the peninsula are an outgrowth of the resolution of these problems.

MODERN WORK (about 1965 to present)

Most modern regional workers on the Baie Verte Peninsula have been oriented more toward pure research than were the foundation workers. The modern approach has been tempered largely by the exciting hypothesis of an expanding and contracting Atlantic Ocean (Wilson, 1966) and the ensuing worldwide geological revolution brought about by the plate tectonic theory. Modern regional workers have mostly been academicians supported by government- and industry-subsidized research grants. These workers have addressed the two major problems inherited from the foundation workers concerning the age and nature of both tectonism and ultramafic rocks on the peninsula.

Two structural geologists, W.R. Church and M.J. Kennedy, significantly influenced the direction of geological thought on the peninsula in the late 1960's. Both were introduced to the geology of the peninsula by E.R.W. Neale. Detailed structural studies were first undertaken by Church (1965a,b; 1966; 1969). Church had worked in northwestern Ireland and, with the advent of the hypothesis of an expanding and contracting Atlantic Ocean, wished to compare crystalline rocks of the Baie Verte Peninsula with those he had studied in Ireland. Church (1969) introduced a major transformation of ideas on the local stratigraphy that was to influence all geological thought concerning the peninsula for the following decade. He proposed the "... designation of the metamorphic rocks of the Burlington Peninsula which are considered to be coeval with those in the Fleur de Lys region as the 'Fleur de Lys Supergroup'..." This entailed the recognition of western and eastern divisions of the supergroup. The western division consisted mainly of metaclastic rocks that had traditionally been designated as the Fleur de Lys Group. The eastern division was an apparently conformable sequence of deformed rocks, including, from bottom to top, metaclastic rocks of the Ming's Bight Group identical to the rocks of the western division, and metavolcanic rocks of the Pacquet Harbour and Grand Cove Groups. The latter groups were previously considered to be the deformed and metamorphosed northerly portions of the Baie Verte and Cape St. John Groups, respectively. Church (1969) recognized clasts of metamorphic detritus in the Snooks Arm Group, and cobbles of granodiorite similar to the Burlington Granodiorite (which intruded the supergroup) in the Baie Verte Group. Thus, he concluded that the polydeformation and metamorphism of the Fleur de Lys Supergroup had to be pre-Early Ordovician. This age of tectonism also supported his trans-Atlantic correlation of the Fleur de Lys rocks with similar metamorphic rocks of the British Isles.

M.J. Kennedy had also worked previously in Ireland. In Newfoundland, he carried out detailed mapping in the Fleur de Lys area. Kennedy (1969, 1971, 1973, 1975a,b) supported Church's (1969) ideas and eventually proposed the term "Burlingtonian" for the pre-Early Ordovician tectonic event proposed by Church. In addition, Kennedy and Neale rescinded Neale's (1957) earlier notion of the cold injection of serpentinite into the Cape St. John Group. Instead, they considered the ultramafic rock to be a primary intrusion into the group. Thus, the group contained ultramafic detritus and was intruded by ultramafic rocks; therefore, they concluded that there were two generations of ultramafic rock on the peninsula. This idea of two generations of ultramafic rock, though later modified, was to have a strong influence on future workers. In contrast to Church, they still considered the whole of the Cape St. John Group to be Siluro-Devonian in age.

The significance of Church's and Kennedy's ideas was not fully realized until Dewey (1969b) incorporated them into an holistic model for the evolution of the Appalachian-Caledonide System. Dewey added no new local data, but depicted the evolution of the orogen by integrating ideas of orogenesis with the regional geology of Great Britain and Newfoundland. This model had resounding impact on the geological world and became the template for many succeeding models by other workers on the peninsula. In the model, Dewey modified some of the ideas of Church and Kennedy for the Baie Verte area. In contrast to Church (1969), Dewey considered the Grand Cove and Cape St. John Groups as one unit that represented a subaerial Cambrian volcanic pile. He incorporated Neale and Kennedy's (1967) idea of two generations of ultramafic rock; however, he indicated that the first generation intruded the Cape St. John - Grand Cove package. The ideas of Church and Kennedy became firmly established because of the strong impact of Dewey's model on the geological world; on a more subtle level, Dewey's liberties with these ideas were almost unquestionably accepted.

One unusual and unexplained feature of Dewey's (1969b) model was the localized intrusion of ultramafic rocks on the peninsula. However, almost as soon as Dewey's model was published, it became evident that these cryptic intrusions were ophiolites representing oceanic crust and mantle (Church and Stevens, 1971; Dewey and Bird, 1971; Bird et al., 1971; Kennedy and Phillips, 1971; Upadhyay et al., 1971). Church and Stevens (1971) were the first to recognize the significance of the ophiolites. They speculated that the Betts Cove - Baie Verte ophiolites and those of western Newfoundland "...may have formed a single continuous sheet..." and that this sheet may have been thrust out of a small ocean basin over the Fleur de Lys Supergroup. Unfortunately, they did not develop this idea further, possibly because most field evidence appeared to negate this model. Ironically, Church's (1969) own model of pre-Early Ordovician tectonism for the Fleur de Lys rocks directly contrasted with the Early to Middle Ordovician emplacement of ophiolites in western Newfoundland; how could the western Newfoundland ophiolites be transported over a predeformed, uplifted terrane?

Thus, workers turned to a different explanation of the origin of ophiolites on the peninsula, that of marginal basins separated from each other by the predeformed Fleur de Lys terrane (Dewey and Bird, 1971; Kennedy et al., 1972; Kennedy, 1975a; Kidd, 1974, 1977). The root zone for western Newfoundland ophiolites was now considered to be a marginal basin west of the Fleur de Lys terrane in White Bay. This model accounted for a pre-Early Ordovician Fleur de Lys deformation as well as two generations of ultramafic rocks on the peninsula. However, the evidence for the two ultramafic generations changed again. The eastern Fleur de Lys Supergroup was now believed to be underlain by old oceanic crust, because the Cape Brulé porphyry, a granite body included by Dewey and Bird (1971) in the supergroup, intruded ophiolitic rocks at Nippers Harbour and also contained xenoliths of ophiolitic rocks. Thus, the pre-Early Ordovician Nippers Harbour ophiolite represented the first generation of ophiolitic rocks (Dewey and Bird, 1971; Bird et al., 1971). Their second generation was represented by the Lower Ordovician Baie Verte and Betts Cove ophiolites, which were interpreted to represent closed marginal basins. Because Dewey and Bird (1971) considered all of the Cape St. John Group to be pre-Early Ordovician in age, they had to reassess the significance of the unconformity previously recognized between the Cape St. John and Snooks Arm Groups; after all, how could pre-Lower Ordovician rocks overlie Arenig rocks? Therefore, they reassigned rocks directly beneath the unconformity, which were identical to the Snooks Arm Group, to the pre-Lower Ordovician Beaver Cove Group, which they considered equivalent to the Nippers Harbour ophiolite.

This interpretation of the regional geology was further supported by the work of W.S.F. Kidd (1974, 1977). He found a block of deformed Fleur de Lys metaclastic rock within conglomerates of the Baie Verte Group near Micmac Lake. Since the Baie Verte Group was considered to be correlative with the Lower Ordovician Snooks Arm Group, the Fleur de Lys rocks had to have been deformed before the Early Ordovician. Kidd (1974) also confirmed the presence of small clasts of schistose rock in the Snooks Arm Group sedimentary rocks.

M.J. Kennedy supported many of these ideas (Kennedy, 1971, 1973, 1975a,b; Kennedy and Phillips, 1971; Kennedy et al., 1972); however, his conception of the two generations of ultramafic rocks was different from that of other workers. Based on structural data, he thought that deformed ultramafic rocks in the Fleur de Lys and Baie Verte areas, as well as some on the Point Rousse Peninsula and in the Nippers Harbour - Tilt Cove area, were pretectonic layered intrusions of pre-Early Ordovician age (Kennedy and Phillips, 1971). He considered the second generation of ultramafic rocks to consist of the Lower Ordovician Baie Verte and Betts Cove ophiolites. Later, Kennedy (1973) conceded that both generations were ophiolitic.

During the early 1970's, while most seasoned workers were refining the regional concepts of the pre-Early Ordovician tectonism and two generations of ultramafic rocks on the peninsula, a silent insurrection was slowly building against the accepted ideology. Two Cambridge University students, M.J. de Wit and J.T. Bursnall, initiated doctoral studies on the peninsula under the supervision of J.F. Dewey. De Wit (1972, 1974, 1980) mapped a substantial transection of the Fleur de Lys terrane and recognized continental basement and cover in the belt. Bursnall (1975) mapped the structurally complex area north of Baie Verte. Together (Bursnall and de Wit, 1975), they proposed the existence of only one generation of ophiolites in the area, contemporaneous with Fleur de Lys deposition, based on structural relationships between the Fleur de Lys Supergroup and ultramafic rocks of the adjacent Advocate Complex to the east. Furthermore, they concluded that the Fleur de Lys deformation was Taconic, and could be as young as Llanvirnian, based on a comparison of the structure and stratigraphy of transported rocks in western Newfoundland with rocks in the Baie Verte area. In his thesis, Bursnall (1975) recognized mélange zones that predated the main Fleur de Lys deformation and related them to the passage of ultramafic rocks through the terrane; he considered the transport of these ultramafic rocks as responsible for the main Fleur de Lys deformation. Here, in the work of Bursnall and de Wit, was the first support of the original speculative ideas of Church and Stevens (1971).

Accepted ideas on the regional geology were further assaulted by work on the eastern part of the peninsula. Regional mapping sponsored by the Newfoundland government, under the supervision of J.R. DeGrace, indicated that Church's (1969) Grand Cove Group was deformed Cape St. John Group and that the Nippers Harbour ophiolite was equivalent to the Betts Cove ophiolite (DeGrace et al., 1975, 1976). Neale et al. (1975) as well as DeGrace et al. (1975, 1976) reconfirmed the unconformable relationship between the Cape St. John Group and the underlying Lower Ordovician Snooks Arm Group. These workers therefore considered the whole of the Cape St. John Group to be Siluro-Devonian in age. The rocks thus could not represent a pre-Ordovician island arc (Dewey and Bird, 1971; Kennedy, 1973) and their deformation had to be much younger than the Early Ordovician age originally proposed by Church (1969). Now, the validity of the Fleur de Lys Supergroup, the pre-Early Ordovician Fleur de Lys deformation, and the concept of two generations of ultramafic rocks were all being seriously questioned.

In an effort to reconcile contrasting interpretations of the regional geology of the peninsula, Williams et al. (1977) locally revised the stratigraphy and added new evidence concerning the timing of the Fleur de Lys deformation. Along with Williams (1977), they agreed with Bursnall's (1975) interpretation that mélanges in the Fleur de Lys Supergroup marked the passage of ophiolites over the terrane; however, they now linked this transport of ophiolites directly with the Taconic emplacement of allochthonous ophiolites in western Newfoundland. They considered the Baie Verte Line, that is, the discontinuous chain of ophiolitic ultramafic rocks that bisects the peninsula, to be the root zone for the allochthonous ophiolites. In the past, the major obstacle to this interpretation had been that the Baie Verte Group ophiolites along the line contained predeformed Fleur de Lys detritus; thus, the ophiolites of the group could not have been involved in the Fleur de Lys deformation. Williams et al. (1977) noted that there was also deformed ophiolite detritus in the group. They inferred that the volcanic and clastic rocks along the Baie Verte Line were younger than the ophiolitic parts of the group; thus, the ophiolites could have been involved in the Fleur de Lys deformation. They proposed the name Flat Water Group for the younger rocks, abandoned the term Baie Verte Group, and assigned remaining ophiolitic portions of the group to the Advocate and Point Rousse Complexes. They correlated these ophiolites of the peninsula with those of western Newfoundland. Since Williams et al. (1977) considered the Baie Verte Line to be a major structural zone, they questioned any lithic and structural correlations of the Fleur de Lys terrane with rocks eastward across the line. Thus, they supported the abandonment of the eastern division of the Fleur de Lys Supergroup (Church, 1969), as suggested by DeGrace et al. (1975, 1976).

From the outset of this project, the ideas of a single generation of ophiolites and an Early Ordovician age of deformation for the Fleur de Lys terrane have appeared to fit most reasonably with the field relationships observed on the peninsula and with the regional geology of Newfoundland. However, three problems that were integral parts of Church's (1969) concept of the Fleur de Lys Supergroup remain unsolved. These include (i) the significance of predeformed detritus in the Snooks Arm Group, (ii) the nature of the metaclastic Ming's Bight Group, which is identical to parts of the Fleur de Lys terrane, yet appears to form part of a sequence conformable with the Siluro-Devonian Cape St. John Group (Neale et al., 1975; DeGrace et al., 1976), and (iii) the age and nature of deformation and metamorphism on the eastern portion of the peninsula.

Since most of the available evidence supports an Early Ordovician age of deformation for the Fleur de Lys terrane, the first problem appears insignificant at this time. It is possible that the fabric in the small, predeformed schistose fragments in the Snooks Arm Group resulted from local, rather than regional, events that are common in volcanic-plutonic terranes; for example, the fabric may have formed either in the aureole rocks around a pluton or along an active fault. The relationship of the Ming's Bight Group to surrounding units and the age and mechanism of polytectonism on the eastern part of the peninsula appear to be more significant and complex dilemmas; these are the major regional problems addressed in the present study.

ACCESSORY STUDIES

Many investigations undertaken were not directly concerned with the evolution of regional geological concepts for the peninsula, but were significant in their individual ways. These accessory studies include government studies, theses, specialized and review papers and unpublished manuscripts, reports and notes. They vary in nature from mineral exploration to geological investigations. The accessory studies are summarized in Table 2-1.

TABLE 2-1: *Accessory studies.*

WORKER	NATURE OF INVESTIGATION
Wadsworth (1884)	mineral deposits of Cape St. John peninsula mentioned in general survey of Notre Dame Bay
Garland (1888)	described Tilt Cove deposits
deLauney (1894)	compared Tilt Cove and Norwegian deposits
Howley (1918)	described metamorphic rocks on east side of White Bay and Granby Island
Sampson and Agar (1916) (Summarized in Betz, 1948)	unpublished notes on White Bay geology and the Tilt Cove deposits
Fuller (1941)	mapped metamorphic rocks of Fleur de Lys area; doctoral study and government report
Livingstone (1942)	related mineralization of Rambler prospect to Acadian fracturing and vein filling; doctoral thesis
Quinn (1945)	reviewed geology and history of Rambler area
Rose (1945)	mapped Baie Verte greenstones between Rambler area and Ming's Bight; M.Sc. thesis
Maclean (1947)	mapped felsic and plutonic rocks in a small portion of southeastern corner of peninsula; government report

WORKER	NATURE OF INVESTIGATION
Betz (1948)	mapped a small area along southeastern White Bay; government report
Czamanske (1956)	report to Canadian Johns-Mansville Co. Ltd. on Birchy Lake ultramafic body; part of a doctoral thesis
Donaghue et al. (1959)	reported on revived operations at Tilt Cove
Purcell (1961)	study of the Terra Nova Mine; M.Sc. thesis
Williams (1963)	included eastern tip of Cape St. John Peninsula in 1:250,000 map of the Botwood (2E) sheet
Williams (1964, 1969)	reviewed previous work on Fleur de Lys Group
Papezik (1964)	described Tilt Cove nickel deposit
Craig (1967)	investigated the East Mine in Tilt Cove; M.Sc. thesis
Neale (1967)	government report of activities with notes on Fleur de Lys metaconglomerate and Horse Islands geology
Harland (1969)	brief report on Fleur de Lys metaconglomerate
Phillips et al. (1969)	compared Fleur de Lys rocks with those of northwestern Ireland
Williams and Stevens (1969); Stevens (1970)	suggested correlation of parts of Fleur de Lys Group with basal clastic sequences of western Newfoundland
Coates (1970)	structural-metamorphic study of area south and east of Pacquet Harbour; M.Sc. thesis
Cockburn (1971)	described intrusive rocks in La Scie area
Frew (1971)	petrography and some geochemistry of the Goldenville deposits - B.Sc. thesis
Schroeter (1971)	mapped Nippers Harbour - Rogues Harbour area; M.Sc. thesis
Gale (1971)	mapped and investigated geochemistry of Rambler area; Ph.D. thesis
Williams et al. (1972, 1974)	included the area in regional reviews of the geology of Newfoundland
Riccio (1972)	studied petrography of ophiolitic rocks at Betts Cove
de Wit (1970)	studied origin of sodic feldspar porphyroblasts in Fleur de Lys Supergroup
de Wit (1976a, b)	investigated origin of rotated garnet porphyroblasts in White Bay Group
de Wit and Strong (1975)	reported on occurrence and geochemistry of eclogites in Fleur de Lys terrane
Papezik and de Wit (1973)	described rutile occurrences along eastern White Bay
Upadhyay (1973)	outlined stratigraphy and geochemistry of the Snooks Arm Group, including Betts Cove ophiolite; Ph.D. thesis
Kennedy et al. (1973)	correlated rocks on Grey Islands with those of Fleur de Lys terrane

WORKER	NATURE OF INVESTIGATION
Norman (1973); Norman and Strong (1975)	described stratigraphy, structure and geochemistry of ophiolitic rocks on the Point Rousse Peninsula; M.Sc. thesis
Church and Riccio (1974)	discussed Betts Cove sheeted dikes
Blagdon (1974)	reported on Advocate Mine
Tuach (1974)	reported on Rambler Mines
Dean and Strong (1975a, b)	included southeastern portion of peninsula in regional compilation of Notre Dame Bay; 1:63,360 map sheets
Sangster and Thorpe (1975)	reported lead isotope age for Ming ore body
Tuach (1976); Tuach and Kennedy (1978)	detailed structural and stratigraphic assessment of the Rambler area; M.Sc. thesis
Miller and Deutsch (1976)	conducted gravity survey across peninsula and recognized contrasting basement on either side of peninsula
Upadhyay and Neale (1976)	reported on geochemistry of the Betts Cove ophiolite and Snooks Arm Group
Church (1977)	noted a west to east decrease in TiO_2 in ophiolites in western Newfoundland
Coish (1977a, b); Coish and Church (1979)	described geochemistry and metamorphism of Betts Cove ophiolite in Betts Cove area; Ph.D. thesis
Jenner (1977); Jenner and Fryer (1980)	undertook detailed geochemical study of Snooks Arm Group; M.Sc. thesis
Mattinson (1975, 1977)	U-Pb ages for rocks on Cape St. John Peninsula
Dallmeyer (1977)	$^{40}Ar/^{39}Ar$ age spectra studies of the Fleur de Lys terrane
Bell and Blenkinsop (1977, 1978a, b)	Rb/Sr studies on Cape St. John Peninsula
Pringle (1978)	Rb/Sr studies on selected units
Kidd et al. (1978)	described detailed aspects of ophiolitic rocks on the Point Rousse Peninsula
Dean (1978)	included the eastern part of the peninsula in a regional compilation of Notre Dame Bay
Church (1978a)	discussed the origin of the Fleur de Lys eclogites
St-Julien et al. (1976); Williams and St-Julien (1978)	defined the Baie Verte - Brompton Line and correlated rocks in eastern Quebec with those of the Baie Verte Peninsula
Upadhyay (1978a)	reported on high magnesia lavas from the Betts Cove ophiolite
Upadhyay (1978b)	discussed Church's data on the geochemical spectrum of Newfoundland ophiolites

WORKER	NATURE OF INVESTIGATION
Church (1978b)	reply to Upadhyay's discussion with details of correlations between Quebec and Newfoundland ophiolites
Young (1978)	reported on slope stability studies at Advocate Mines
Tuach (1978a, b)	compiled data on mineral deposits of the Baie Verte Peninsula
Fitzpatrick (1981)	outlined geological evolution of the Goldenville Mine; B.Sc. thesis
Squires (1981)	outlined the geology and ore history at Tilt Cove Mines; B.Sc. thesis

In addition to the workers cited above and in Table 2-1, many mineral exploration companies and past and present mining concerns in the area have made contributions to the understanding of the local geology. Their reports are too numerous to mention here, but of particular note is the work of Falconbridge Nickel Mines, Johns-Manville, Brinex, M.J. Boylen and Associates, Consolidated Rambler Mines and, more recently, the work of Phillips Management, Noranda, Newmont Exploration, and Ionex. Reports on this work are on file at the Newfoundland Department of Mines and Energy.

Chapter III

REGIONAL SETTING AND GENERAL GEOLOGY

Geologically, the Baie Verte Peninsula is situated at the northernmost tip of the Appalachian Orogen, along the rugged north coast of Newfoundland. It is from a study of this shoreline that Williams (1964) initiated the concept of tectonic divisions on the island; his original scheme has since evolved into a tectonostratigraphic breakdown of Newfoundland geology into four zones (Williams, 1976) (Figure 3-1). These tectonic zones, including the Humber,

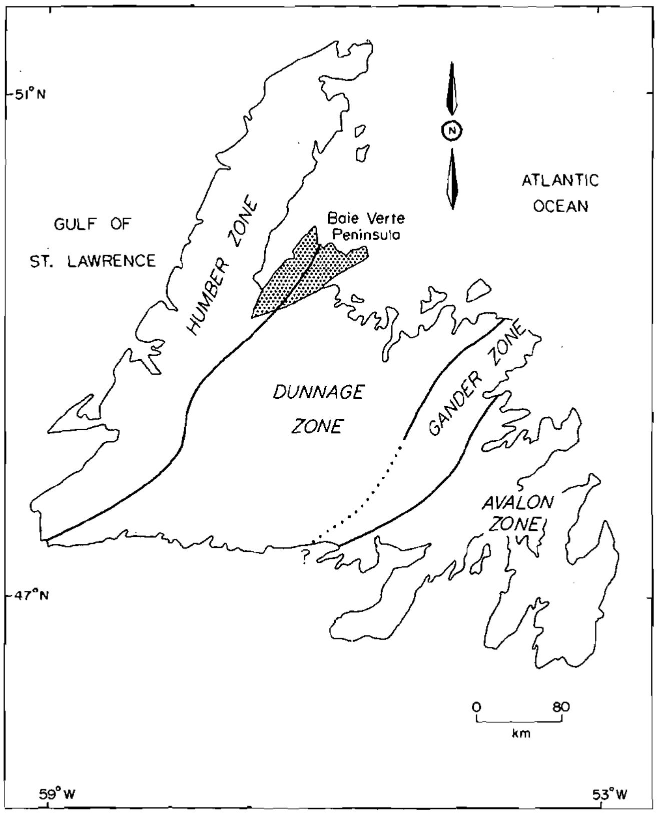


Figure 3-1: *Tectonostratigraphic zones of Newfoundland (after Williams, 1978a).*

Dunnage, Gander and Avalon Zones, are based on the contrasts of pre-Middle Ordovician rocks and are applicable throughout most of the Appalachian Orogen (Williams, 1978a). The Baie Verte Peninsula exposes rocks of both the Humber and Dunnage Zones; toward the west, rocks of the Humber Zone record the evolution and destruction of the ancient continental margin of eastern North America, whereas the centrally located Dunnage Zone contains ophiolite suites and volcanic complexes that collectively represent the vestiges of an Early Paleozoic Iapetus ocean. The more easterly two elements of the orogen, the Gander and Avalon Zones, represent the eastern margin of Iapetus and an easterly continent-based terrane, respectively.

The Baie Verte Peninsula is unique within the Appalachian Orogen, for only here, along the wavewashed North Atlantic coast, is the contact between the Paleozoic margin of North America and the ancient oceanic domain of Iapetus nearly continuously exposed. The contact between these fundamental tectonic environments appears razor-sharp from the reference frame of the whole mountain chain, though on the scale of the peninsula, the lithologic and structural complexities of this zone are obvious. In particular, the apparent dearth of fossils compounds the difficulties in understanding this highly tectonized area; only two fossil locales are known. Most geological problems on the peninsula are resolvable, though, by using both good exposures of key relationships along the coast and in stream beds, and the growing data bank of isotopic age dates on key rock units.

The following discussion outlines the major geological elements of the peninsula and the immediate area, and considers briefly the nomenclature of the rock units (Table 3-1) and use of radiometric age dates on the peninsula.

MAJOR GEOLOGIC ELEMENTS

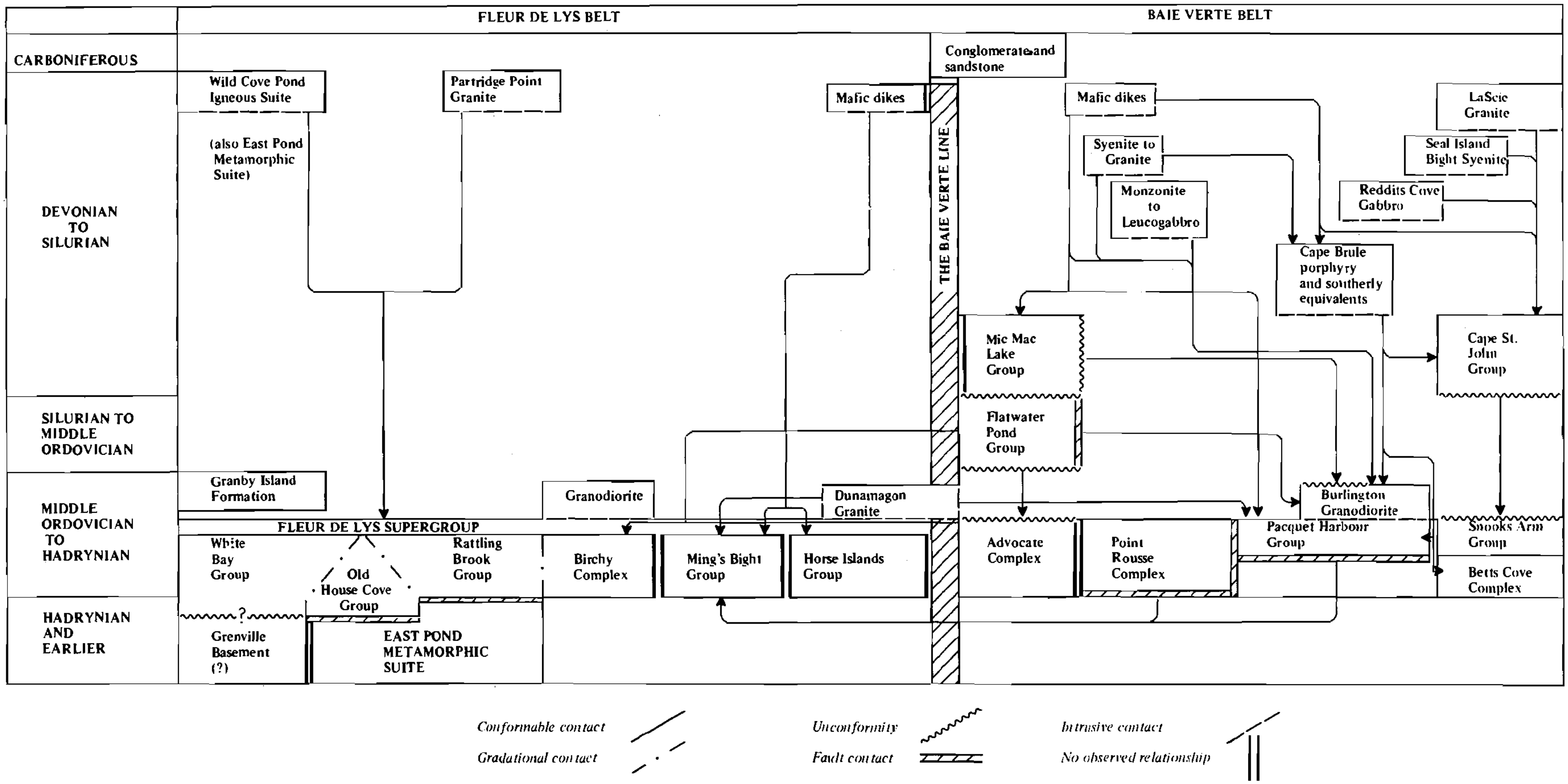
On the Baie Verte Peninsula, the boundary between the Humber and Dunnage Zones is a remarkably sharp structural zone, herein termed the Baie Verte Line, that bisects the peninsula into contrasting lithic terranes from Birchy Lake northward to Baie Verte; at Baie Verte it swings abruptly eastward and heads out to sea through Pacquet Harbour (Figure 1-1). Along its north trending section, the line is a major fault accentuated by a steeply dipping screen of dismembered ultramafic bodies; the east-west portion of the line has been polytectonized and intruded by granite thus forming a broad structural zone. To the west and north of the line lies an arcuate belt of polydeformed schists and gneisses and granitoid intrusions of the eastern margin of the Humber Zone, termed here the Fleur de Lys Belt. To the north the belt is largely submerged beneath the Atlantic Ocean. The southeastern portion of the peninsula comprises ophiolite suites, volcanic cover sequences, and various intrusive rocks of the Dunnage Zone, that are here informally referred to as the Baie Verte Belt.

FLEUR DE LYS BELT

The three major lithic elements recognized in the Fleur de Lys Belt on the peninsula are the local structural basement, a dominantly metaclastic cover sequence and postkinematic granitoids; a fourth, incidental, division of clastic rocks is confined to Granby Island in White Bay. The orthotectonic character of the basement and cover sequence distinguish the belt from other portions of the Humber Zone.

Migmatites, banded gneisses, metaconglomerate, and psammitic and semipelitic schists of the East Pond Metamorphic Suite form the local structural basement in the core of

Table 3-1: *Table of formations.*



the belt. They are separated from the surrounding cover rocks, in many places, by a steeply dipping tectonic zone of mica schists, although locally the contact appears to be gradational. The intense deformation of the suite compared with that of the cover rocks, as well as the presence of local zones of anatexis, eclogite pods, and a coarse polymictic conglomerate, all prompted de Wit (1972, 1980) to interpret the area underlain by the East Pond Metamorphic Suite as Grenvillian basement overlain by metaconglomerate. Alternatively, it is herein suggested that the suite may largely represent a deeper structural, and possibly stratigraphic, level of the cover with only local windows of predeformed regional Grenvillian (?) basement rocks.

The cover sequence overlying the East Pond Metamorphic Suite is termed the Fleur de Lys Supergroup. The main outcrop belt of the supergroup is composed of four apparently conformable and interfingering groups composed of psammitic, semipelitic and graphitic schist, marble, greenschist[1] and amphibolite. One of these units, the White Bay Group, appears to unconformably overlie Grenville basement at Hampden. Two mainly psammitic-semipelitic units, one at Ming's Bight and the other on the Horse Islands, are geographically isolated from the main belt, but are apparently rooted to it. The age of the supergroup is uncertain, though it has been considered to be the metamorphosed distal equivalent of uppermost Hadrynian to Lower Ordovician platformal deposits of western Newfoundland (Neale and Nash, 1963; Williams and Stevens, 1969; Bursnall and de Wit, 1975). A small fragment of a brachiopod shell recovered from a marble breccia in the supergroup (S. Stouge, personal communication, 1979) indicates a Paleozoic age for at least portions of the supergroup.

Both the East Pond Metamorphic Suite and the Fleur de Lys Supergroup were post-tectonically intruded by a largely granitoid batholith termed the Wild Cove Pond Igneous Suite. A muscovite-bearing granite that intrudes the Fleur de Lys Supergroup at Partridge Point may be a late phase satellite of the suite.

BAIE VERTE BELT

There are three major components of the Baie Verte Belt, including ophiolitic suites, volcanic cover sequences, and intrusive rocks. The Cambro-Ordovician stratigraphy of the belt largely distinguishes it from other portions of the Dunnage Zone. In particular, the geochemistry and the nature of strata overlying the ophiolitic units appear to be unlike those of other parts of the zone. The four ophiolitic units on the peninsula, the Advocate, Point Rousse, and Betts Cove Complexes and the Pacquet Harbour Group, are considered mutually correlative though they are distinguishable on the basis of their geographic position and their structural state. The Betts Cove Complex has been isotopically dated as Early Ordovician and is conformably overlain by the fossiliferous Arenigian Snooks Arm Group, a cover sequence of mafic volcanics, volcaniclastics and epiclastics. The Pacquet Harbour Group, a sequence of mafic and felsic volcanic rocks in the center of the peninsula, appears to represent only the uppermost portion of an ophiolite and its overlying cover. Pillow lavas of the group are geochemically identical to the Betts Cove lavas, though in the Pacquet Harbour Group, lower members of the ophiolite are unexposed. The group is Early Ordovician or older based on the isotopic age dates of rocks intrusive into it. The remaining ophiolites are probably also Early Ordovician or older. The remainder of the cover sequences range in age from Early Ordovician to Devonian, based on isotopic ages and regional relationships. Rocks above the Point Rousse ophiolite are considered equivalent to the Early Ordovician Snooks Arm Group, whereas the Flat Water Pond Group and equivalent rocks that overlie the Advocate Complex are probably younger than Early Ordovician based on the clast assemblage of their basal conglomerates (Neale and Kennedy, 1967; Williams et al., 1977). Younger portions of the volcanic sequences, the Cape St. John and Micmac Lake Groups, are largely felsic volcanic sequences that unconformably overlie the Snooks Arm and Flat Water Pond Groups, respectively. They are considered to be Siluro-Devonian in age because of their unconformable relations with underlying Ordovician units, isotopic age data, and their similarity to Siluro-Devonian sequences elsewhere in the Dunnage Zone.

A variety of intrusive rocks cut the Dunnage Zone ophiolitic and volcanic sequences. Isotopic age dates and field relationships indicate that two broad intrusive pulses occurred, one during the Ordovician, and the other spanning the Silurian and Devonian periods. The Ordovician intrusions include a biotite-hornblende granodiorite body, termed the Burlington Granodiorite, and the pink, biotite-bearing Dunamagon Granite. The Burlington pluton is unconformably overlain by the Micmac Lake Group. Younger intrusive rocks of the area include the Cape Brulé granite porphyry, the La Scie Granite, the Seal Island Bight Syenite, the Reddits Cove Gabbro, and associated granitoid to gabbroic rocks in the southeastern portion of the area. Locally, the younger intrusions have an alkaline to peralkaline character.

In addition to these main elements, a patch of probably Carboniferous sedimentary rocks is poorly exposed at the extreme southern portion of the belt; their contact with surrounding rocks is unexposed.

STRUCTURE AND METAMORPHISM

All pre-Carboniferous strata and structures on the peninsula, including the Baie Verte Line, wrap around a major structure termed the Baie Verte Flexure (Hibbard, 1982). It is defined by the change in structural trends from a north-northeasterly to an easterly orientation. The flexure appears to be a primordial feature that pre-dates the tectonism of rocks on the peninsula; younger structures have apparently been molded to its form. Based on radiometric dates, tectonism of Fleur de Lys Belt rocks on the west limb of the flexure appears to be broadly of Taconic age (Early Ordovician), whereas a period of younger (Devonian) tectonism is recorded

[1] Greenschist, in this report, refers to a green schistose rock that is rich in any combination of amphibole, chlorite and epidote. There is no intended implication of metamorphic grade. Where reference to greenschist grade metamorphism is made in this report, the word "greenschist" is followed by either of the terms grade or facies.

in rocks of the easterly Fleur de Lys Supergroup and all units of the Baie Verte Belt.

Fleur de Lys Belt rocks on the west limb of the flexure, with the exception of the Granby Island Formation, display three main phases of deformation, although up to seven phases have been recognized locally (Church, 1969; Bursnall, 1979); the metamorphic grade of these units ranges from upper greenschist to middle amphibolite facies, with eclogitic phases occurring locally in the East Pond Metamorphic Suite. The Granby Island Formation displays a maximum of two fabrics and greenschist facies metamorphism.

Tectonism of Fleur de Lys rocks on the east limb of the flexure is similar in aspect to that of the westerly Fleur de Lys, but is apparently related to the polydeformation of the Baie Verte Belt. In this belt all rocks north of the La Scie highway, with the exception of the Flat Water Pond Group and the Point Rousse Complex, are polydeformed and attain upper greenschist to lower amphibolite facies metamorphism. South of the highway, these features grade into greenschist facies rocks that contain only one penetrative fabric, such as the southern portions of the Cape St. John and Pacquet Harbour Groups, the southern portion of the Cape Brulé porphyry, and the Snooks Arm Group.

Deformation of the Advocate Complex ranges from polyphase near the boundary with the Fleur de Lys Supergroup to a single phase in the southeast; the whole of the complex has been metamorphosed to greenschist facies. The Flat Water Pond and Micmac Lake Groups as well as the Point Rousse Complex display a single penetrative fabric and lower greenschist facies metamorphism throughout their outcrop area.

SURROUNDING ROCKS

Rocks that bound the peninsula are grossly similar in character to those on the peninsula. To the east and southeast of the Green Bay Fault (Figure 1-1) lie rocks similar to those of the Baie Verte Belt. These include ophiolitic rocks of the Lushs Bight Group and its overlying Ordovician volcanic sequences, the Western Arm and Catchers Pond Groups (Dean, 1978). The Lushs Bight Group is intruded by both granodiorite similar to the Burlington batholith and granite porphyry like the Cape Brulé porphyry. South of the peninsula lie predominantly intrusive and extrusive rocks of the Upper Silurian Topsails Intrusive Complex (Taylor et al., 1980) that are similar to those of the Middle Arm Ridge area on the peninsula. The Carboniferous strata of the Deer Lake basin border the peninsula to the southwest and west; they are juxtaposed against the Fleur de Lys Supergroup and Wild Cove Pond Igneous Suite along the Cabot Fault, to the west, whereas to the southwest the contact is unexposed but presumed to be the Green Bay Fault. The Carboniferous rocks appear to extend northward beneath the waters of White Bay, since Carboniferous rocks outcrop to the north along strike at Conche and have been detected on the bottom of White Bay during oceanographic studies (Haworth et al., 1976). Rocks exposed on the west side of White Bay consist of Cambro-Ordovician platformal clastic and carbonate rocks of the Coney Arm Group that unconformably overlie Grenville basement, a transported granitoid complex named the Coney Head Complex, and the unconformably overlying Silurian volcanics and sediments of the Sops Arm Group (Williams, 1977b).

STRATIGRAPHIC NOMENCLATURE

During the past decade geological understanding of the peninsula has grown rapidly, so rapidly that the stratigraphic nomenclature of the region has been a makeshift system. Its evolution is complex; in many cases, it violates the code of stratigraphic nomenclature. In particular, the following major transgressions of the code are noted:

(i) Lithostratigraphic units have been defined on the basis of time equivalency and inferred structural-metamorphic histories; the original definition of the Fleur de Lys Supergroup (Church, 1969) is based on these premises.

(ii) Units have been subdivided without proper definition of the newly proposed division nor of the reduced original unit; this applies in particular to the separation of the Pacquet Harbour Group (Church, 1969) and the Advocate sequence (Kennedy, 1975a) from the Baie Verte Group (Baird, 1951).

(iii) Incidental names with vague definitions or names that transgress the rules of priority have been introduced in the literature; these include the Rambler Group (Bird and Dewey, 1971), the Baie Verte Ophiolite Complex (Bird et al., 1971), and the Ming's Bight Ophiolite Complex (Kidd et al., 1978).

(iv) Formations have been defined by type sections rather than by mappability; in particular, this applies to the Big Head Formation and Barry-Cunningham Formation of Kidd et al. (1978).

In addition, numerous revisions of the stratigraphic nomenclature of local areas, by many workers, have compounded the problem.

In an attempt to comply with the code of stratigraphic nomenclature, and yet preserve as many traditional names as possible, a revised formal stratigraphy for the Baie Verte Peninsula is herein proposed.

The present study introduces three new major units, the East Pond Metamorphic Suite for rocks in the area mapped as "the basement" by de Wit (1972, 1980), the Horse Islands Group for metamorphic rocks confined to the Horse Islands, and the Granby Island Formation for the clastic sedimentary rocks on Granby Island, White Bay. The Fleur de Lys Supergroup is redefined, and one of its components, the Seal Cove Group (de Wit, 1972), is renamed the Old House Cove Group, because the term Seal Cove has been reserved for a unit elsewhere in Newfoundland. The Rattling Brook Group (Watson, 1947) is redefined and the Birchy Schist (Fuller, 1941) is changed to the Birchy Complex following the recommendation of Williams et al. (1977). The term "Baie Verte" is dropped from stratigraphic usage (Williams et al., 1977), and it is replaced by the following units: the Advocate Complex (Bursnall, 1975; Kennedy, 1975a; Williams et al., 1977), the Point Rousse Complex (Williams et al., 1977), the Flat Water Pond Group (Williams et al., 1977) and the Pacquet Harbour Group (Church, 1969). The Snooks Arm Group is redefined to exclude the ophiolitic rocks at its base; the latter are assigned to the Betts Cove Complex, which is defined to include the former Nippers Harbour Group (Baird, 1951). The traditional Cape St. John Group (Baird, 1951) is used in much the same sense as DeGrace et al. (1976) with minor additions. The following names that have been either properly defined or traditionally used are also recognized: Micmac Lake Group (Kidd, 1974), Cape Brulé porphyry (Baird, 1951), Burlington Granodiorite (Baird, 1951; Neale and Nash, 1963), Dunamagon Granite (Baird, 1951), Partridge Point Granite (Fuller, 1941; Kennedy, 1971), Ming's Bight Group (Watson, 1947; Baird, 1951), and White Bay Group (Betz, 1948). The names Wild Cove Pond Igneous Suite (Kidd, 1974), Reddits Cove Gabbro, Seal Island Bight Syenite and La Scie Granite (DeGrace et al., 1976), which have all been informally proposed for various intrusive rocks on the peninsula, are herein recognized.

RADIOMETRIC AGE DATES

Radiometric age dates have proven useful in interpreting the geological history of the peninsula. These dates have been determined over the past twenty years by a number of methods, including U/Pb, Pb/Pb, Rb/Sr, K/Ar, and Ar/Ar (Figure 1-1). Each system represents a chronometer that is susceptible, to different degrees, to a variety of geochemical and physical parameters; hence, each type of chronometer records different events. The geological significance of each date generated on the peninsula must thus be assessed in the context of the method employed, the quality and reliability of the data, and the regional geological relationships. Generally, the data used in determining dates previously reported on the peninsula have not been available; hence, in the following chapters, these dates are discussed in the context of methods and regional relationships; where data have been available, it is discussed where relevant. All new geochronological data collected during the present study are compiled in Appendix IV. All radiometric ages quoted in this report are based on decay constants and isotopic abundance ratios listed in Steiger and Jager (1977).

Chapter IV

STRATIGRAPHY OF THE FLEUR DE LYS BELT

INTRODUCTION

The Fleur de Lys Belt forms the well defined easterly orthotectonic margin of the Appalachian Humber Zone in Newfoundland. It is discontinuously exposed from the Baie Verte Peninsula to the Grey Islands (Kennedy et al., 1973) more than 100 km to the north, and at least as far south as Grand Lake in west-central Newfoundland (Knapp et al., 1979) (Figure 4-1).

The three major elements of the belt, including basement, cover, and postkinematic granitoids, are exposed on the Baie Verte Peninsula, and at Grand Lake (Knapp et al., 1979) whereas only cover and granitoids outcrop on the Grey Islands (Kennedy et al., 1973).

The structural basement for most of the belt on the Baie Verte Peninsula is represented by the East Pond Metamorphic Suite; however, more easterly parts of the belt may be underlain, locally, by ophiolitic basement. The Fleur de Lys Supergroup encompasses almost all of the cover rocks, which are locally intruded by prekinematic and synkinematic igneous rocks. The contact between the suite and the supergroup is marked by a zone of highly strained schists. All of these rocks are intruded by postkinematic granitoids, the Wild Cove Pond Igneous Suite and the Partridge Point Granite, and mafic dikes. In addition to these major elements, an incidental unit of sedimentary rocks, the Granby Island Formation, occurs in the map area just offshore from the western edge of the Fleur de Lys Belt.

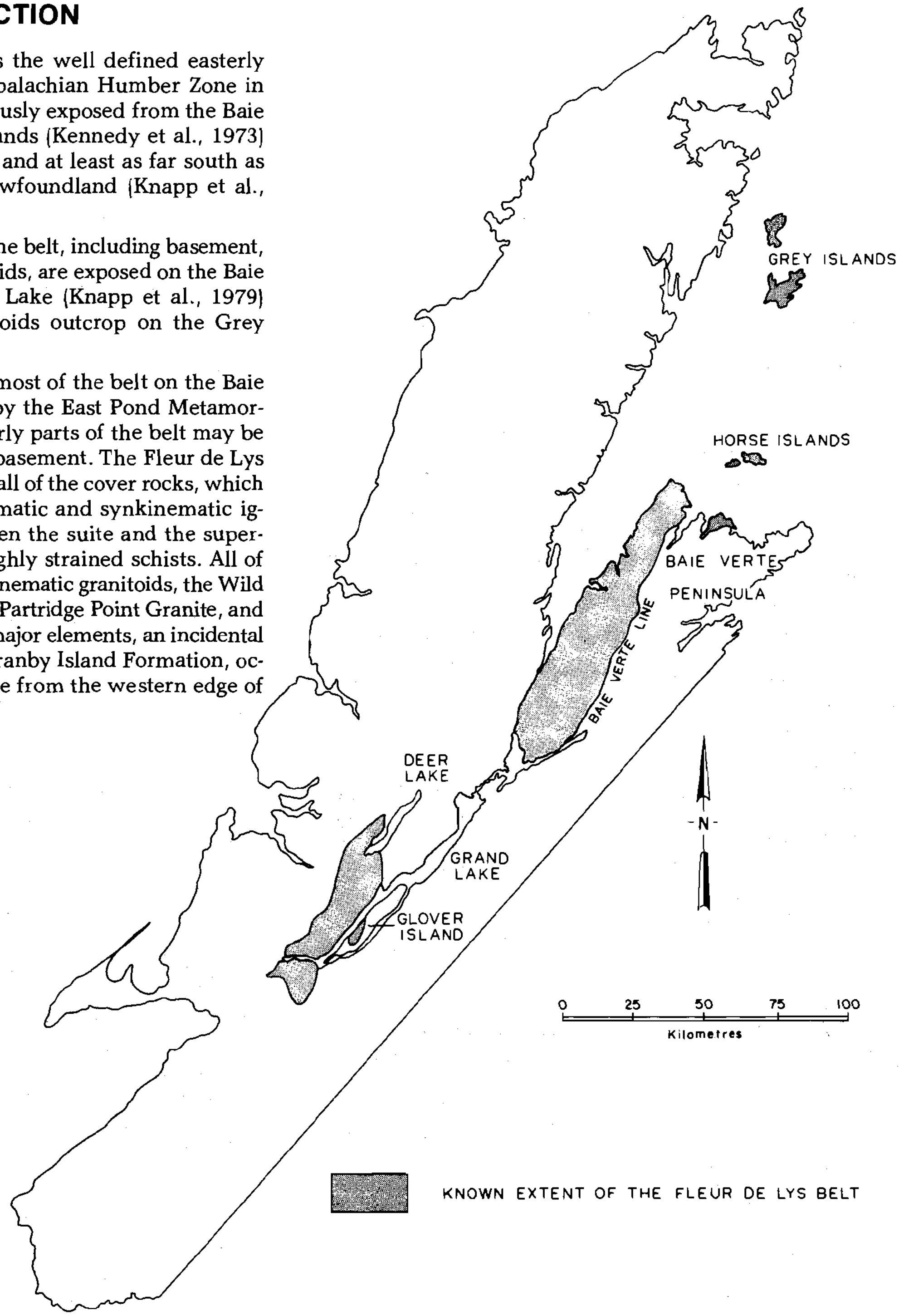


Figure 4-1: *The Fleur de Lys Belt in Newfoundland.*

EAST POND METAMORPHIC SUITE

The name East Pond Metamorphic Suite is herein proposed for the structurally complicated schists and gneisses that outcrop in two ellipsoidal areas in the center of the Fleur de Lys Belt, between Middle and Southern Arms and Wild Cove Pond (Figure 1-1). The suite forms the core of a major anticlinorium in the belt (see Chapter VII). These rocks were previously considered as part of the archaic Fleur de Lys Group (Neale and Nash, 1963) and were later mapped as part of the Rattling Brook Group (Church, 1969). De Wit (1972, 1974, 1980) informally distinguished most of these rocks as "the basement" to the Fleur de Lys Supergroup, and he rigorously described their field aspect, contact relationships, and petrography. Kidd (1974) briefly described rocks from the extreme southeastern portion of the unit and informally included them in his western sequence. The boundaries of the East Pond Metamorphic Suite roughly coincide with those of de Wit's (1972, 1974, 1980) basement and Kidd's (1974) western sequence, but modifications are made to de Wit's interpretation of the internal relationships of the unit.

The suite is named for exposures on East Pond (also locally known as Middle Arm Pond), where almost all of the constituent rock types occur. It is characterized by extremely heterogeneous structure, such that the layering observed in most outcrops is of uncertain origin. Only locally, such as in the migmatites and banded gneisses, can layering be shown to be tectonic or, as in outcrops of little deformed metaconglomerate, to represent bedding. Thus, the designation of a meaningful type section is impractical without more detailed work; however, exposures on East Pond are here considered as a representative section. Likewise, due to the complicated structure of the unit, its base and top cannot be reliably identified. The maximum outcrop width of the unit is approximately 12 km immediately northeast of Wild Cove Pond.

The East Pond Metamorphic Suite is composed of four major lithologic associations, including (i) migmatitic and banded gneisses, (ii) the Middle Arm metaconglomerate, (iii) psammitic and semipelitic schists, and (iv) granitic gneiss. The psammitic and semipelitic schists compose more than 80% of the suite, with most of the remainder represented by the metaconglomerate; migmatite and banded granitic gneisses are minor.

In Figure 1-1, the granitic gneiss is included with the migmatitic rocks because of its very limited outcrop area and close association with these rocks at East Pond. The metaclastic rocks of the suite also contain numerous amphibolite layers and podiform bodies that are similar in aspect to those in the Fleur de Lys Supergroup, and will be discussed with them as Fleur de Lys amphibolites (see Prekinematic Intrusive Rocks). Locally in the East Pond Metamorphic Suite, they are eclogitic. A pyroxene-quartz-plagioclase gneiss reported by de Wit (1972, 1980) was not located during the present study.

An intense steeply dipping, north-northeasterly trending, penetrative L-S fabric is developed throughout the suite; the migmatite and banded gneiss contain complex structures that predate this main fabric. The suite has been metamorphosed to at least lower amphibolite grade (see Chapter VII).

The original contact relationships between rock types in the suite are uncertain due to the intense deformation and poor exposure. In the field the contacts appear to be gradational; it is uncertain if these represent depositional, structural, or metamorphic gradation. On a structural basis, de Wit (1972, 1974, 1980) considered the migmatites as an anatectic front within deformed supracrustal rocks (the psammitic and semipelitic schists); he interpreted both of these elements as regional Grenville basement that is unconformably overlain by the metaconglomerate. This sequence was not recognized during the present study. An alternative interpretation of the original relationships between these units, based mainly on lithologic aspects, will be discussed following a description of the rock types in the suite.

MIGMATITE AND BANDED GNEISS

Migmatite and banded gneiss of this report correspond, in part, to the acidic rocks and mixed banded gneisses, respectively, of de Wit (1972, 1980). These rocks are confined to small, poorly defined, areas in the central regions of the suite. Migmatites occur in small patches at East Pond and Pine Pond, and in the area southwest of Gull Pond; collectively, these areas constitute less then 1 km². Banded gneiss of uncertain origin occurs on the Westport road approximately 6 km west of Flat Water Pond; it is similar to banded paragneiss elsewhere in the suite, but appears to have a more complex structural history. It appears to be as limited in areal extent as the migmatite.

Migmatite

Migmatite is best exposed on the west shore of East Pond, where a light gray, granoblastic, quartzofeldspathic neosome permeates a pelitic, quartzitic, and amphibolitic paleosome. Where preserved, layering within the migmatite is steeply dipping and northwesterly striking, askew to the typical northeasterly trend of rocks in the suite. Pelitic and quartzitic components occur as well defined schollen, or rafts, within the neosome and, locally, the sooty gray weathering quartzite is injected by the neosome, forming agmatite (Plate 4-1). Commonly, the amphibolite paleosome and the neosome are discontinuously interlayered (stromatic structure, Plate 4-2). Well defined amphibolite layers grade laterally into ghostly schlieren and, locally, in these vaguely defined zones of the migmatite, large clots of amphibole freckle the outcrop (Plate 4-3), resembling stictolithic structures (Mehnert, 1968). Similar features occur in the smaller exposures of migmatite elsewhere in the area, though on the west side of Pine Pond, the amphibolite paleosome forms well defined rafts. In the field these amphibolite rafts are indistinguishable from other amphibolites in the suite and in the Old House Cove Group. At Pine Pond, crosscutting relationships between amphibolite layers indicate that they are relict dikes (de Wit, 1972) (Plate 4-4). At the southern end of East Pond, elongate, rounded, cobble-sized clasts of quartzite, semipelite, augen gneiss, and granitic gneiss occur in a fine grained, sugary, quartzofeldspathic matrix that is similar to the migmatite neosome. Due to poor exposure, it is uncertain if this is a portion of the migmatite terrane. Ellipsoidal quartzite clasts occur in the neosome in the migmatite terrane southwest of Gull Pond. Clasts at both of these localities strongly resemble those in the Middle Arm metaconglomerate.

Plate 4-1: *Agmatite formed from quartz-feldspar neosome injecting sooty gray weathering quartzite; migmatite of East Pond Metamorphic Suite at East Pond.*

Plate 4-2: *Interlayered amphibolite paleosome and granitoid neosome; migmatite of East Pond Metamorphic Suite at East Pond.*

Plate 4-3: *Amphibolite schlieren and amphibole clots in East Pond Metamorphic Suite migmatite at East Pond.*

Plate 4-4: *Crosscutting relationship between amphibolite dikes in East Pond Metamorphic Suite migmatite at Pine Pond.*

The neosome is composed of fine grained quartz with varying amounts of plagioclase and microcline that form an equigranular (0.3 to 0.5 mm) groundmass to muscovite, biotite, yellow-green epidote, local hornblende and garnet, and accessory sphene. Typically, the texture of these rocks is dominated by irregular grain boundaries and complex vermicular intergrowths of quartz and feldspar, with unoriented mica. Where layering is present, it is generally defined by the concentration of biotite and epidote in centimetre scale bands. De Wit (1972) noted the variability in composition of these rocks; he recognized an increase in biotite and epidote toward the amphibolite paleosome and a decrease in biotite and lack of microcline within the amphibolites.

The mafic paleosome at Pine Pond contains both amphibolitic and eclogitic mineral assemblages. One foliated amphibolite pod is composed of skeletal amphiboles (<1 mm) with fine grained plagioclase intergrowths and minor epidote, biotite, carbonate and sphene. Later actinolite poikiloblasts (up to 2 mm) overgrow the skeletal amphibole, but are oriented in the plane of the fabric. A separate amphibolite pod contains an equigranular (<1 mm) granoblastic assemblage of garnet-omphacite-quartz with accessory sphene. The pale green omphacite is bordered by a fine grained symplectite dominantly of amphibole. Minor biotite occurs in association with the idioblastic garnet.

Development of the migmatitic layering of the rocks at East Pond predates the main fabric of the suite. On the east side of East Pond, schlieren in the migmatite are crenulated, with the axial planar cleavage of these folds parallel to the main fabric in the complex. The distinct and unusual vermicular metamorphic texture indicates that the migmatites were recrystallized after their formation (de Wit, 1972, 1980). Thus, these rocks were subjected to anatexis prior to the main deformation and metamorphism of the suite. The nature of the neosome and paleosome indicates that the protoliths for the migmatite may have been a sequence of interlayered clastic sedimentary and mafic igneous rocks, some of which were dikes (de Wit, 1972, 1980). Locally, as at the south end of East Pond, metaconglomerate appears to have been mobilized to form a small part of the migmatite.

Banded Gneiss

Banded quartzofeldspathic gneiss of uncertain origin is equivalent to part of de Wit's (1972, 1980) mixed banded gneisses division. It occurs in exposures along the Westport road, in the central part of the eastern outcrop area of the suite. The gneissic layering strikes north-northeast and is steeply dipping. The areal extent of the gneiss is uncertain due to poor exposure away from the road. The banded gneiss consists of extremely regular leucocratic and mesocratic bands (centimetre to millimetre scale) that are primarily defined by the concentration of biotite and muscovite. The thin, light and dark gray weathering bands can be traced for metres across an outcrop. Leucocratic bands are granitic in composition and generally thinner than the granodioritic mesocratic bands. Intrafolial folds are common in both banding types. In the more easterly part of the outcrop area, thin pink granitic veins and pods, less than 5 cm across, crosscut the banding and may indicate that these rocks once reached the physical conditions of incipient melting. The westerly outcrop area of banded gneiss is crosscut by two very fine grained, white weathering quartzofeldspathic dikes, one of which contains a xenolith of the banded gneiss (Plates 4-5, 4-6).

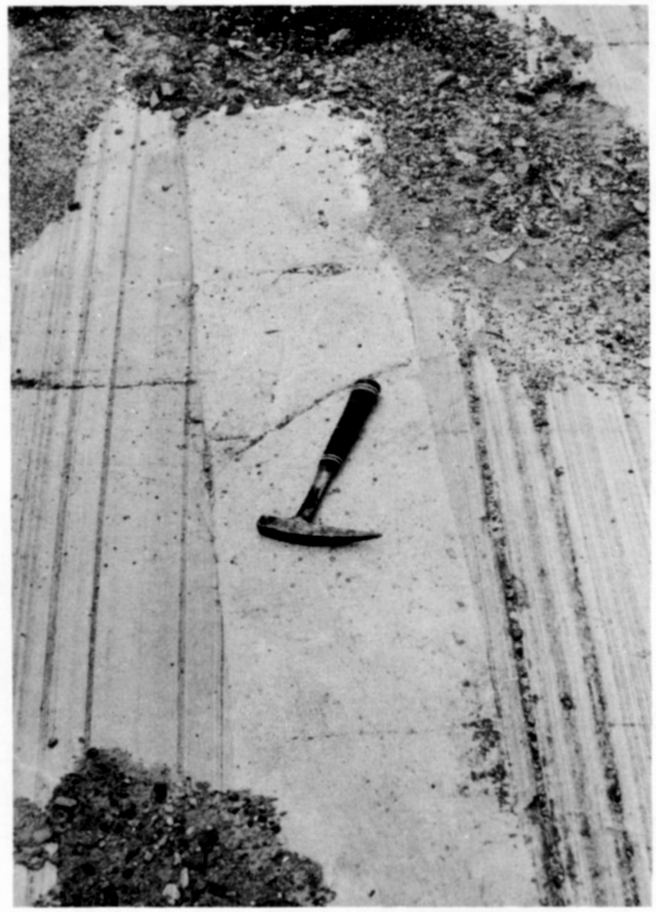


Plate 4-5: *Banded gneiss of the East Pond Metamorphic Suite crosscut by quartzofeldspathic dike, along the Westport Road.*

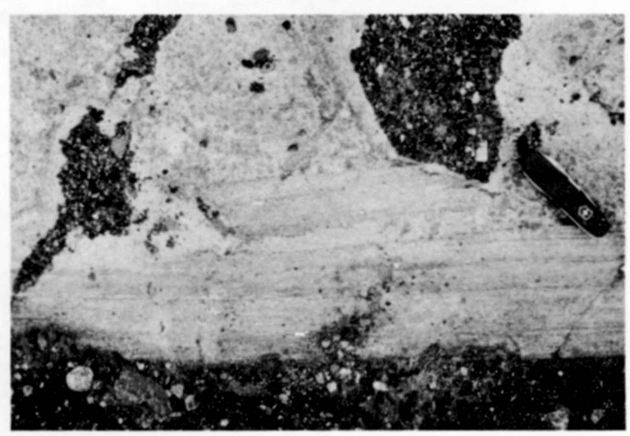


Plate 4-6: *Xenolith of banded gneiss within quartzofeldspathic dike, along the Westport Road; East Pond Metamorphic Suite.*

The gneiss is composed of fine grained (<0.5 mm), interlocking, equigranular quartz and feldspar, with dark green-gray biotite, muscovite, and pale yellow epidote. Locally, muscovite defines an intrafolial fabric, suggesting a complex history of development for the banding seen in outcrop.

The banded gneiss displays the main deformational fabric prevalent throughout the suite; the fabric also appears to affect the thin granitic veins and aplite dikes. At one locality, however, complex intrafolial fold interference patterns are crosscut by the main fabric of the suite. Thus, the complex banding in the gneiss predates the main deformation of the suite (see also de Wit, 1972, 1980). Their fine grain size indicates that they may represent blastomylonitic rocks, though the protolith for these rocks is uncertain.

MIDDLE ARM METACONGLOMERATE

Neale and Nash (1963) first outlined a metaconglomerate unit that outcrops in a belt trending south-southwesterly from Middle Arm. Subsequently, Church (1969) informally referred to it as the Middle Arm tilloid, a member of his Middle Arm Brook Formation of the Rattling Brook Group. De Wit (1972, 1974) assigned it to the Middle Arm Pond Metaconglomerate Formation, a division of his Seal Cove Group. The present study informally reverts to a modified version of Church's nomenclature. The metaconglomerate, as considered here, encompasses rocks mapped as the Middle Arm Pond metaconglomerate and "basement" by de Wit (1972, 1974).

The thickness of the metaconglomerate is uncertain due to the structural complexity of the area; the base and the top of the unit have not been identified, though the base of the unit may be represented by outcrops near the migmatitic gneiss on East Pond (see below). The maximum outcrop width of the unit is approximately 2 km. De Wit (1972) described the metaconglomerate:

> The metaconglomerate occurs in well-bedded units (several cms to at least 10 m thick), separated by beds of fine to coarse grained, light yellowish-gray psammite of similar thickness. The psammite sometimes shows interlamination of thin (mm-cm) pelitic layers which are generally continuous, but in quantity very subordinate to the psammite. No extensive petrological study was made of the matrix of the clasts, but microcline, epidote, allanite and garnet are commonly abundant constituents of the quartz-plagioclase (An<25) psammites.
>
> Within the conglomerate beds, lamination is usually absent; the clasts are set in a uniform matrix. The pebble to matrix ratio varies considerably, and may change abruptly or gradually. When relatively little matrix is present (? indicative of high energy environments), the clasts sometimes display a pronounced imbrication which is thought to be at least partly original. In some of the thicker psammitic beds, rare clasts float in the psammitic matrix. The conglomerate unit thus shows a self-supporting framework of clasts in some beds and matrix-supported clasts in others.

De Wit (1972, 1974) also reported local channel structures from the unit.

The metaconglomerate was heterogeneously deformed during the main deformation of the suite. It ranges from a slightly deformed, recognizable metaconglomerate to a banded paragneiss (Plates 4-7 to 4-9). Deformation within the metaconglomerate is zonal, with abrupt transitions from relatively undeformed rock into strongly banded paragneiss; this transition occurs over a distance of less than 100 m in Middle Arm, and over approximately 30 m on the Bear Cove road. The paragneiss locally resembles the banded gneiss of

Plate 4-7: *Mildly deformed Middle Arm metaconglomerate of the East Pond Metamorphic Suite along the west shore of Middle Arm. Note predepositionally deformed clast in foreground.*

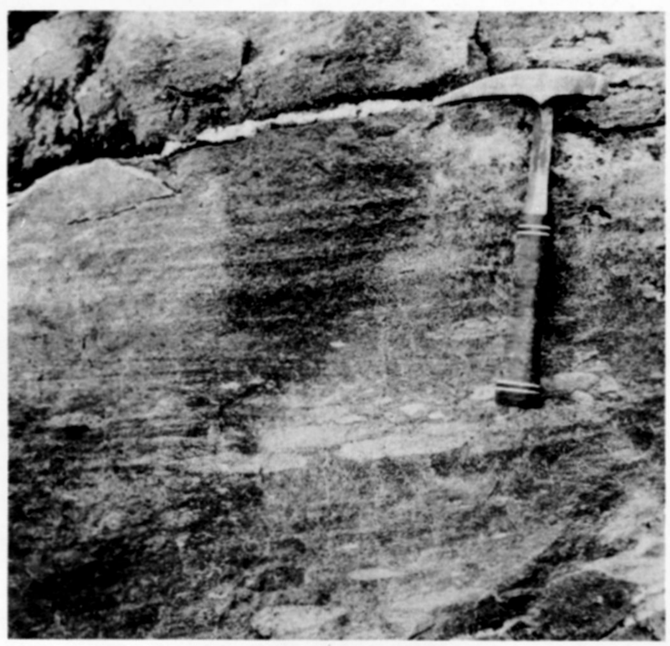


Plate 4-8: *Strongly deformed Middle Arm metaconglomerate at the bottom of Middle Arm; East Pond Metamorphic Suite.*

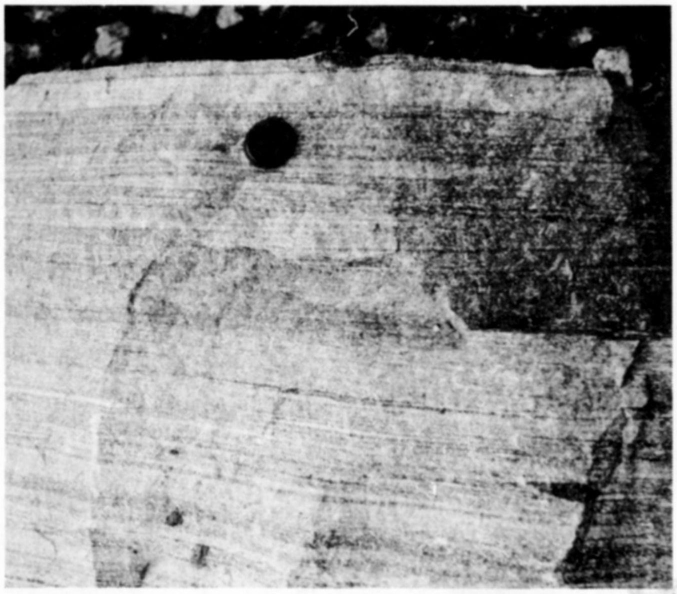


Plate 4-9: *Intensely deformed metaconglomerate along the Bear Cove Road. This outcrop grades into less deformed metaconglomerate on either side. Note highly attenuated clasts just to the left and below lens cap and at lower right of photograph. Middle Arm metaconglomerate, East Pond Metamorphic Suite.*

uncertain origin, though the former is generally coarser grained and its banding is not as well defined. Poorly banded, coarse grained paragneiss, probably derived from the psammitic interlayers of the unit, occurs at the north end of East Pond and along the Bear Cove road.

The interpretation of the environment of deposition of the metaconglomerate has been disputed. Church (1966, 1969) and Harland (1969) considered it to be a tilloid, and possibly correlative with other Eocambrian tilloids of the Caledonides. De Wit (1972, 1974) noted that no features within the metaconglomerate demand a glacial origin for the unit; he considered it to be fluviatile and/or shallow marine in origin. Cogently, Neale (1967) noted that "...it is unlikely that any evidence is preserved to support the recent proposal [referring to Church, 1966] that these rocks are tilloids." This likewise applies to de Wit's interpretation (1972, 1974), though correlation with units outside the map area appears to support a shallow marine origin for the unit (see final section of East Pond Metamorphic Suite).

PSAMMITIC AND SEMIPELITIC SCHISTS

Description

Psammitic and semipelitic schists comprise most of the outcrop area of the East Pond Metamorphic Suite. These rocks were incorporated in the Middle Arm Brook Formation and semipelite formation of Church (1969) and referred to as the gray gneisses by de Wit (1972); they also include a portion of de Wit's mixed banded gneiss division. Kidd (1974) referred to these rocks in the southeastern portion of the suite as the western sequence.

The thickness of these schists is uncertain due to structural complexities; the maximum outcrop width is approximately 10 km in the area just north of Wild Cove Pond.

The fine to medium grained, gray to buff weathering psammitic to semipelitic schists are interlayered with subordinate quartzite, biotite quartzite, pelite, calc-silicates and pebbly metaconglomerate; the assemblage is sedimentary in origin. The schists are variably layered, ranging from thin, millimetre to centimetre scale, gneiss-like banding, to massive, flaggy, centimetre to metre scale layering that may represent primary bedding (Plate 4-10). Layering is generally northeast striking and steeply dipping. The psammitic and semipelitic schists are exposed best along the Westport road, where the layering, defined by the concentration of biotite-muscovite-rich bands and quartz-rich bands, is very regular along strike. At Pine Pond, fine grained psammite and biotite quartzite are intensely strained. Quartzite is common at the eastern edge of the suite as well as just northeast of Gull Pond. At the latter locality, it is biotite-rich, in places containing biotite poikiloblasts up to 4 cm long. The prismatic outline of these crystals suggests that they are pseudomorphs of staurolite, though there is no staurolite present (de Wit, 1972). Calc-silicates and epidosite are most common on East Pond; de Wit (1972) reports that the epidosite locally grades into amphibolite. Pebbly psammite and minor metaconglomerate occur at Gull Pond and, locally, to the east and the south of East Pond.

In the area east of Wild Cove Pond and northeast of Kidney Pond, exposures of medium to coarse grained, feldspathic psammitic schist, similar to that of the Old House Cove Group, are interspersed among typical fine grained, thinly layered, gneiss-like psammites of the suite. Poor exposure in this area prevents the interpretation of the relationships between these rocks; hence, all are included here in the East Pond Metamorphic Suite.

Petrography

The psammite schist of the suite is composed dominantly of quartz, muscovite, biotite and feldspar, with accessory epidote, sphene and apatite; garnet occurs locally. Grain size rarely exceeds 1 mm. Granoblastic, strained quartz with serrated boundaries and small plates of muscovite and biotite define the fabric in this schist. The biotite is typically altered at its edges to chlorite. De Wit (1972) described a color zonation of biotite in the suite, in which dark green varieties occur to the east, whereas medium to dark brown biotite occur to the west. Plagioclase is rarely twinned; microcline was noted locally. Epidote, in many places, contains a ruddy brown allanite core. With an increase in mica content, the psammite grades into semipelitic schist. The calc-silicate rocks commonly consist of quartz, epidote, carbonate, biotite, locally altered to chlorite, and feldspar with accessory sphene and apatite.

Structural Relationships

The deformation of the psammitic and semipelitic rocks is apparently the same as that of the Middle Arm metaconglomerate. Locally, as on East Pond, psammitic paragneiss of the unit loses its defined layering concomitantly with an increase in potassic feldspar; in thin section, these rocks

Plate 4-10: *Well banded quartz-rich psammitic schist of the East Pond Metamorphic Suite, along the Westport Road near Gull Pond; M.J. Kennedy becomes reacquainted with winged friends of the peninsula.*

resemble the granitic gneiss. These features suggest incipient migmatization of the unit at East Pond. The intense deformation of the complex prevents any interpretation of the environment of deposition for these metasediments.

The layering in the psammitic and semipelitic schists has been interpreted as gneissic layering that predated the main deformation of the suite (de Wit, 1972, 1980). However, during the present study, no evidence of complex gneissic layering, such as intrafolial folds, was noted, and hence the layering is considered to represent original bedding that was probably transposed.

GRANITIC GNEISS

A small body of granitic gneiss, somewhat similar to the neosome of the migmatite, is exposed over an area of less than 100 m^2 on the west side of East Pond. The gneiss is composed of coarse equigranular quartz, feldspar, and muscovite, with ghost layers and folia of fine grained biotite; garnet occurs locally. Layering is absent only in very small areas. Pink granitic dikes up to 10 cm wide, which are similar in composition to this body, intrude the Middle Arm metaconglomerate on the east side of the pond. De Wit (1972) described the petrography of the granitic gneiss as follows:

...the granite contains 20-40% microcline of a total 65-75% feldspar content. Microcline occurs subidiomorphically (1-4 mm) together with an early idiomorphic to subidiomorphic plagioclase (albite/oligoclase, An < 20-25). Sometimes plagioclase overgrows a relict foliation. Patches of string perthite are common in the microcline, and true myrmekite is present in small amounts at the boundary between the two feldspars.

A late albite phase is prominent, because it is always extremely dirty (densely sericitized/saussuritized).

Muscovite occurs in large sheets (3 mm) or booklets, randomly oriented, and/or as small flakes, with biotite, chlorite, yellow granular epidote and sphene, which are ubiquitous accessories, sometimes up to 5%. Traces of garnet and apatite occur.

...Melanocratic folia consist mainly of biotite, epidote, muscovite, quartz and albite with some microcline (0.1 - 0.3 mm). They vary from equigranular, with randomly orientated idioblastic flakes to symplectic intergrowths, sometimes with dimensional orientation of the flakes, parallel to the foliation.

INTERPRETATION OF ORIGINAL STRATIGRAPHY

Most original contacts between constituent rock types of the East Pond Metamorphic Suite were reworked and completely masked by an intense deformational and metamorphic event; as a result, all contacts are gradational. Based on structural evidence and the nature of the metaconglomerate, de Wit (1972, 1974) interpreted the migmatite, gneiss and psammitic and semipelitic schists as remobilized, predeformed basement unconformably overlain by the metaconglomerate. De Wit maintained that the earlier units all display a banding that predated the main deformation evident in the metaconglomerate. Detailed structural analysis, like that of de Wit, was not undertaken during this study; however, clear evidence for the existence of a predeformed terrane could be recognized only in the migmatites at East Pond and in the banded gneisses, as noted above. Thus, I consider only these small patches of gneisses to be remobilized predeformed basement to the suite.

The psammitic and semipelitic schists are interlayered with the metaconglomerate; in addition, metaconglomerate pods and layers occur within the psammitic-semipelitic schist unit. The gradational contact between these units is thus interpreted to represent an original depositional transition. The original sequence of the metasediments is uncertain, though the metaconglomerate appears to be a basal deposit (see Age and Correlation). Since the metaconglomerate occurs in close association with the migmatite at East Pond and contains boulders of predeformed rocks, I interpret the contact between the migmatite and the metaconglomerate as a major unconformity (see also de Wit, 1972, 1974, 1980). Where the metaconglomerate is absent, as in the eastern outcrop area of the complex, the psammitic and semipelitic schists may directly overlie the unconformity.

The granitic gneiss on East Pond occurs along the contact of the migmatite and metaconglomerate and appears to intrude the metaconglomerate on the east side of the pond. Also, feldspar and garnet porphyroblast growth is more pronounced in both the psammitic schists and metaconglomerate near this contact. These features, along with the local mobilization of the metaconglomerate and incipient anatexis of the psammitic schist, may indicate remobilization of the predeformed rocks along the inferred unconformity during the deformation of all of these rocks. De Wit (personal communication, 1978) conjectured that a regolith may have developed on the predeformed rocks; such a plastic zone would help promote mobilization during deformation, as suggested above.

Migmatitic rocks at Pine Pond and in the area southwest of Gull Pond are of uncertain origin; they may represent either predeformed basement or the product of local migmatization of the cover rocks. The nature and eclogitic mineralogy of the amphibolite paleosome at Pine Pond hints at a remobilized cover origin.

This interpretation of the original stratigraphy of the East Pond Metamorphic Suite is attractive in that it agrees with previous interpretations of the structure of this portion of the peninsula. Murray (Murray and Howley, 1881), Baird (1951), and Neale and Nash (1963) indicated that the rocks of the western Baie Verte Peninsula form a major, north-northeasterly trending anticlinorium. In the present study, the two outcrop areas of the suite may represent a modification of the anticlinorium (see cross-sections, Figure 1-1), with older predeformed migmatites and gneisses in the core of each area, surrounded by younger metasediments. De Wit's (1972, 1974, 1980) interpretation of the stratigraphy demands a more complex structure to account for the metaconglomerate in the central portion of the western outcrop area of the complex.

AGE AND CORRELATION

There is no direct evidence for the age of the East Pond Metamorphic Suite. However, rocks of the suite can be dated, indirectly, by regional correlation with dated units.

The Middle Arm metaconglomerate is identical to metaconglomerate zones localized in the Oody Mountain Amphibolite of the White Bay Group. The Oody Mountain Amphibolite is interpreted as the Eocambrian base of a sequence that unconformably overlies Grenville basement (see White Bay Group, Fleur de Lys Supergroup). In addition, Williams and Stevens (1969) suggested that the Middle Arm metaconglomerate is correlative with the basal conglomerate of the Bateau Formation that unconformably overlies Grenville basement on Belle Isle. De Wit (1972, 1974) concurred with this correlation and considered the remainder of the suite to be equivalent to the Long Range Grenvillian gneisses on the east side of White Bay.

Based on these correlations and the interpreted unconformity with predeformed rocks of the suite, the Middle Arm metaconglomerate is here also interpreted as a basal Eocambrian deposit (see also de Wit, 1972, 1974) overlying a predeformed basement that is most likely Grenvillian in age. The psammitic and semipelitic schists may represent more highly deformed correlatives of portions of the White Bay Group that immediately overlie the Oody Mountain Amphibolite. These schists are not considered to be Grenvillian as suggested by de Wit (1972, 1974, 1980).

FLEUR DE LYS SUPERGROUP

INTRODUCTION

The use of the name "Fleur de Lys" has had a long history in the stratigraphic nomenclature of the peninsula (Appendix II). It was originally proposed by Fuller (1941) for the series of schists and gneisses that outcrop near Fleur de Lys Harbour; subsequently, Baird (1951) included these rocks in the Fleur de Lys Group and extended the unit south to the latitude of Baie Verte community. Church (1969) raised the unit to supergroup status, and assigned to it all of the polydeformed and metamorphosed rocks of the Baie Verte (Burlington) Peninsula. He erected eastern and western divisions of the supergroup that were separated by less deformed rocks of the Baie Verte Group. De Wit (1972) separated "the basement," i.e. the East Pond Metamorphic Suite of this report, from the western division of the supergroup. In addition, some workers included, either in part or totally, rocks of the Advocate Complex within the western Fleur de Lys succession (Kennedy, 1973, 1975a; Bursnall and de Wit, 1975), whereas others excluded them from the supergroup (Church, 1969; Williams et al., 1977). The validity of Church's eastern division of the Fleur de Lys Supergroup was questioned and subsequently rejected by recent workers (DeGrace et al., 1975, 1976; Williams et al., 1977; Hibbard, 1982). Changes to the definition of the supergroup have been informal and not recognized by all workers; as a result, it has become increasingly unclear as to what rocks are included in the Fleur de Lys Supergroup. Therefore, formal redefinition of the supergroup is proposed here.

The Fleur de Lys Supergroup is defined as the collection of groups of dominantly metaclastic schists with interlayered amphibolite and greenschist, the main belt of which outcrops on the western portion of the Baie Verte Peninsula. In contrast to the East Pond Metamorphic Suite, layering in most of the supergroup is thought to represent bedding. The Fleur de Lys rocks extend northward, approximately 100 km along strike, to the Grey Islands (Kennedy et al., 1973), eastward to the Horse Islands and the area east of Ming's Bight, and southward to the Grand Lake area (Church, 1969; D. Knapp and D. Kennedy, personal communications, 1978). In this outcrop belt, the supergroup includes the White Bay Group (Betz, 1948), the Old House Cove Group (new), the Rattling Brook Group (Watson, 1947), the Birchy Complex (Fuller, 1941; redefined herein), the Ming's Bight Group (Baird, 1951), the Horse Islands Group (Hibbard and Bursnall, 1979) and unseparated schists on the Baie Verte Peninsula, as well as unnamed units on the Grey Islands (Kennedy et al., 1973) and in the Grand Lake area (Knapp et al., 1979). The total thickness of the supergroup is uncertain due to the structural repetition of units; the maximum observed outcrop width of the Fleur de Lys Supergroup is approximately 13 km, just south of Baie Verte, though the maximum width is probably much greater offshore, north of the peninsula.

Previous workers (Murray and Howley, 1881; Baird, 1951; Neale and Nash, 1963; de Wit, 1972) suggested that the supergroup on the peninsula forms a broad anticlinorium, with a core of rocks described herein as the East Pond Metamorphic Suite. Thus, de Wit (1972) suggested that rocks of the present Old House Cove Group form the base of the Fleur de Lys succession and that the Birchy Complex is at the top; the White Bay and Rattling Brook Groups have been considered equivalent units on opposing limbs of this structure. In the present study, the contacts between all of these groups are interpreted to be conformable and largely represent lateral interdigitations. The Birchy Complex, in part, represents a dismembered ophiolite suite (Bursnall, 1975) and, hence, only the uppermost portion of the complex could have been deposited on the Rattling Brook Group. In addition, the more seaward portion of the White Bay Group has been found to overlie a granitic gneiss basement (see Oody Mountain Amphibolite below), with the easterly remainder of the group lying within a synclinorium. Thus, the notion of a simple stratigraphy disposed in a broad anticlinorium for the main outcrop belt of the supergroup is significantly modified here. These contacts are described in detail with each respective unit, and the overall structure of the belt is summarized in a later section (see Chapter VII).

Amphibolite layers and pods characteristic of both the Old House Cove Group and the East Pond Metamorphic Suite were subdivided by de Wit (1972). In the present report, his divisions are not recognized; hence they are unseparated and described here following the Fleur de Lys rocks, with other prekinematic intrusive rocks. Small pods of serpentinite and talc-carbonate schist that occur in the Rattling Brook Group and the Birchy Complex are also discussed in the later section.

The Horse Islands and Ming's Bight Groups lie to the east of the main outcrop belt, but are believed to be rooted to the belt (Haworth et al., 1976; Hibbard, 1982). Their relationships to the main outcrop belt and to each other are uncertain, but these units are thought to be broadly correlative with the westerly groups.

The Fleur de Lys Supergroup in the Baie Verte Peninsula area is polydeformed, displaying at least three major structural phases, and is metamorphosed in the upper greenschist to lower amphibolite facies; locally, middle amphibolite facies metamorphism has been attained.

The Fleur de Lys Supergroup is in tectonic contact with both the East Pond Metamorphic Suite and the Baie Verte Belt, though in both cases there is evidence that the original contacts between portions of these stratigraphic units were depositional.

The age of the Fleur de Lys Supergroup is uncertain. A single brachiopod fragment recovered from marble in the White Bay Group (S. Stouge, personal communication, 1979) to the west of the area is the sole fossil occurrence known in the supergroup and it indicates that at least a portion of the unit is of Paleozoic age. This confirms earlier notions of Murray (Murray and Howley, 1881), Neale and Nash (1963), de Wit (1972), and Bursnall and de Wit (1975), who considered these rocks as Paleozoic in age and correlative with Paleozoic limestones on the western side of White Bay. Murray (Murray and Howley, 1881), Baird (1951), Neale and Nash (1963), and de Wit (1972) all considered the Fleur de Lys terrane to form a large anticlinorium and, thus, become progressively older toward the center of the outcrop belt. Based on regional correlation with rocks of western Newfoundland, de Wit (1972) suggested an age range of Eocambrian to Early Ordovician for the supergroup. Other workers interpreted these rocks to be Precambrian, based on deformation and metamorphism (Fuller, 1941; Watson, 1947; Baird, 1951), pre-Early Ordovician based on the inferred age of deformation (Neale and Kennedy, 1967; Church, 1969; Dewey and Bird, 1971; Kennedy, 1973, 1975a, 1975b; Kennedy et al., 1972; Kidd, 1974, 1977), and older than Middle Ordovician on stratigraphic and structural grounds (Bursnall and de Wit, 1975; Williams et al., 1977; Williams, 1977a).

In the present study, the Fleur de Lys Supergroup is considered to be Late Hadrynian to Early Ordovician in age based on: (i) the single fossil occurrence; (ii) the correlation with less deformed rocks of the Humber Zone; and (iii) regional structural constraints. The similarity of some of the Fleur de Lys schists with some of the psammitic-semipelitic schists of the East Pond Metamorphic Suite indicates that large portions of the suite may be the intensely deformed equivalent of the lower portions of the Fleur de Lys Supergroup.

WHITE BAY GROUP

Definition and Extent

The White Bay Group is the assemblage of metaclastic schists, marble, greenschist and amphibolite that outcrops in a linear north-northeasterly trending belt from Western Arm southward to the area north of Sandy Lake; two isolated patches of the unit occur at Sandy Lake. The group is disposed in a horizontal north-northeast trending synclinorium that is reclined and steeply west dipping in Western Arm; progressively more moderate dips occur southward and are reflected by the broader outcrop pattern of the group in the Big Chouse Brook area. This structure is informally termed the coastal synclinorium (see Chapter VII). The average width of the group is 3 km. It is bounded to the west by the Cabot Fault and White Bay and to the east by the Carrol Hill Slide, the Old House Cove Group, and the Wild Cove Pond Igneous Suite. It is not possible to erect a type section for the group due to its highly variable nature and complex internal structure. Instead, three representative sections that collectively encompass all of the rock types in the group are herein designated at Western Arm, Little Chouse Brook, and Rocky Brook.

The present definition of the group follows the original usage of the term by Betz (1948). Church (1969) informally divided the original group into three formations, and subsequently de Wit (1972) recognized seven formations in the Western Arm area. De Wit (1974) informally assigned the rocks in this area to the Bear Cove Group. In the present study, the original White Bay Group is used (Figure 1-1) and only two formations defined by de Wit (1972) are recognized, including the Garden Cove Formation (a name coined by Church, 1969) and a modified version of the Pigeon Island Formation (Appendix II). The remainder of de Wit's units were not found to be suitable for regional mapping because

(i) some units are limited to the small area of Western Arm and are not readily distinguishable to the south, and (ii) some units are unmappable on a regional scale due to structural and stratigraphic complexities. A new formation, the Oody Mountain Amphibolite, is herein proposed and defined. The balance of the group is undivided.

For the purposes of the following discussion, the group is informally separated and described as two distinct sequences, namely, the Outboard and Inboard sequences. They extend for almost the full strike length of the group and are subparallel to the White Bay coast. It appears that the Outboard sequence is overlain by the Inboard sequence (see Contact Relationships, below).

Outboard Sequence

This sequence encompasses the Oody Mountain Amphibolite and unseparated schists that outcrop along the coast of White Bay from the Penny Hills south to Hampden, and forms the western limb of the coastal synclinorium. Southward, the sequence occurs along the east side of the Cabot Fault and at Sandy Lake. The Outboard sequence is bounded to the east by the Penny Hills slide zone and the Pigeon Island Formation. Younging directions in metaclastics along the coast and a few younging directions found inland are predominantly east facing, suggesting that the sequence as a whole is largely overturned. The presumed base of the sequence is a distinct stratigraphic unit, the Oody Mountain Amphibolite, whereas the remainder of the unit is a highly variable, unseparated assemblage of mainly semipelitic, psammitic and graphitic schist, amphibolite, greenschist, marble and quartzite. The disposition of these rock types is not well defined, as they interfinger and form complicated lensoid outcrop patterns. Their distribution may be either primary or tectonic, or both.

OODY MOUNTAIN AMPHIBOLITE

Description: The name Oody Mountain Amphibolite is herein proposed for the sheetlike body of amphibolite that discontinuously forms the western boundary of the Outboard sequence between Sandy Lake and the Beaches (Figure 1-1). The thickness of the formation is uncertain, as it is truncated to the west by the Cabot Fault and is polydeformed. Its maximum outcrop width, near the southwest branch of the Hampden River, is approximately 1.5 km. The amphibolite is concordant with surrounding schists. The most complete section through the unit, at Rocky Brook, is here taken to be the reference section of the formation. Minor patches of granitic gneiss and metaclastics are closely associated with the formation and are described herein.

The amphibolite is massive, foliated, fine to medium grained and homogeneous throughout its outcrop area. White flecks of feldspar up to 1 cm long, usually oriented in the plane of the main fabric, impart a distinct texture to the rock. In outcrop, it has a deformed diabasic texture; in thin section, though, this texture is seen to be metamorphic. Locally, small, irregular pods and layers of jasper and carbonate up to 8 cm thick occur within the amphibolite.

The amphibolite is closely associated, in places, with outcrops of granitic gneiss. The pinkish to yellowish cream colored gneiss is prominently banded. The banding is defined by the distribution of quartzofeldspathic zones, and lenses and layers of biotite and epidote; in many places, biotite is nearly completely altered to chlorite. The granite gneiss is laterally discontinuous, with outcrops completely surrounded by amphibolite. On Rocky Brook, the gneissosity is obviously truncated by both thin dikes and more substantial portions of the main amphibolite body. Thus, I interpret the granitic gneiss as large screens of predeformed crystalline rock within the amphibolite. On the lower part of Rocky Brook, the granitic gneiss forms approximately one-third of the formation. It decreases eastward up the brook, which is presumably stratigraphically up-section. The granitic gneiss is absent near the contact of the amphibolite and structurally lower metaclastic rocks of the Outboard sequence.

Toward the top of the formation, the amphibolite is conformably interlayered with thin to medium banded, fine grained psammitic and semipelitic schists containing matrix-supported pebbles of quartz and alkali feldspar. Similar metasediments occur interlayered with the easterly portions of the amphibolite at Fox Point and north of the Beaches. Graded bedding in pebbly metaclastics near Fox Point indicates that these metasediments young to the east, which is consistent with the presumed stratigraphy of the sequence. Immediately north of the Beaches, one outcrop of cobbly metaconglomerate occurs as a lens, at least 3 m thick, within the amphibolite. The cobble sized, rounded clasts are matrix-supported and comprise quartzite and predeformed granitic gneiss very similar to the granitic gneiss on Rocky Brook.

Petrography: The amphibolite is characterized by irregular fibrous porphyroblasts of amphibole up to 3 mm long, defining a fabric within a fine grained matrix of untwinned plagioclase and epidote. The amphibole forms up to 50% of the rock and its low extinction angles (12-15°) combined with its ragged habit suggest that it is actinolite. Many of these porphyroblasts are incipiently altered to chlorite, and only locally does well defined chlorite occur. The plagioclase occurs as irregular aggregates that internally display a granoblastic to mortarlike texture in which individual crystal boundaries are vaguely defined. Considering these recrystallization textures, it is highly unlikely that the macroscopic diabaselike texture reflects any primary features. Fine grained subidioblastic to xenoblastic epidote peppers the rock. Minor carbonate locally fills interstices between feldspar aggregates. Locally, opaques comprise up to 5% of the amphibolite. Study of one thin section of granitic gneiss reveals it to be a foliated, mortared mosaic of quartz and feldspar, with finer grained intergranular wisps and more concentrated layers of biotite and epidote. Quartz constitutes approximately 40 to 45% of the rock. Undulose extinction and bent twinning are common in the feldspar, as well as local myrmekitic features. Greenish brown biotite constitutes 15% of the gneiss and is associated with fine grained subidioblastic to xenoblastic epidote. Sphene occurs locally.

OTHER ROCK TYPES

The remainder of the Outboard sequence is composed of an unseparated assemblage of metaclastic rocks, marble and amphibolite. In most places, these rock types alternate on a small scale, such that they are individually mappable only on detailed maps (1:25,000 scale or less). Likewise, little regional significance can be gleaned from measured sections.

METACLASTIC ROCKS: Pelitic, semipelitic, graphitic and psammitic schists, and minor quartzite, constitute the metaclastic rocks. Semipelitic and pelitic schists are the most common of the metaclastic rocks. Generally, they are thin to medium layered (<20 cm), fine grained and consist of any combination of the following minerals: quartz, feldspar, biotite, muscovite, chlorite and epidote, with local pyrite, rutile, sphene, garnet, tourmaline and magnetite. They range in color from buff to light gray through to dark gray, greenish gray and olive, the color dependent upon constituent mineral proportions. In the vicinity of mafic schist and amphibolite layers, these rocks are commonly enriched in chlorite, epidote and feldspar, such as in the area of Twelve Mile Point. Less commonly, distinct gray-green biotite porphyroblastic semipelite and pelite are interlayered with these rocks.

Graphitic schist is intimately and irregularly interbedded with the other metaclastics. This black, rusty to sulfurous weathering, fine grained schist ranges in thickness from thin, wispy, millimetre scale layers up to massive sections tens of metres wide, as on the south shore of Purbeck's Cove and in the area south of Little Chouse Brook. Locally, as at the mouth of Big Chouse Brook, the graphitic schist is interlayered with silvery, medium to coarse grained muscovite-quartz-garnet semipelite. Glossy black crystals of tourmaline are common in all of the metaclastic rocks.

Psammitic schists in the Outboard sequence are typically buff to gray weathering, thinly layered and quartz-rich, and occur as either single layers or groups of layers interbanded with more pelitic rocks (Plate 4-11). Massive psammite layers up to 1 m thick are rare, but occur in a number of layers at Purbeck's Cove. Locally, where the pelitic content of the psammite increases, it tends to have a schistose aspect, such as at the mouth of Little Chouse Brook, although massive varieties predominate. Psammitic schists associated with marble and carbonate schists weather a rusty buff color, as at Little Pumbly Cove.

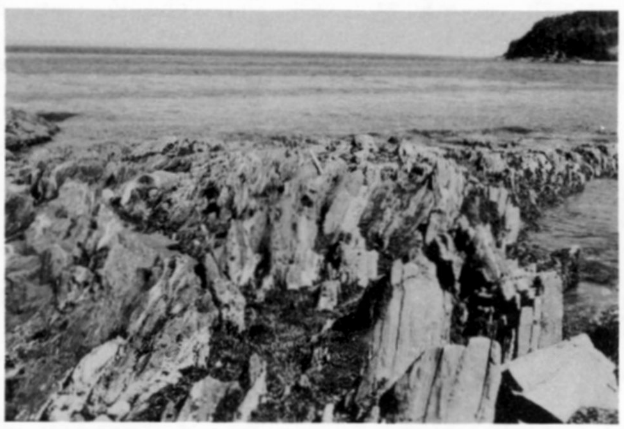


Plate 4-11: *Thin to medium banded psammitic schist interlayered with graphitic and pelitic schist; White Bay Group, Clay Cove, White Bay.*

In many places, thin quartzite layers generally less than 10 cm thick, though locally up to 1 m thick, are sporadically interbedded with the metaclastic rocks. They are generally frosty white to cream weathering, medium grained and contain only a minor pelitic component. Approximately 20 m of thinly layered quartzite is exposed at the mouth of Little Chouse Brook; it is bounded by metaclastics that are locally carbonate-rich. Another thick quartzite section that outcrops approximately 10 m up the brook may represent either a structural repetition of this member, or an independent unit.

Numerous, irregularly distributed, metamorphosed gritty and conglomeratic rocks characterize the Outboard sequence. In many places, they are graded (Plate 4-12), indicating that layering in the sequence largely represents original bedding. The clasts are generally rounded to

Plate 4-12: *Graded gritty metaclastic layer in center of photograph; from overturned White Bay Group rocks at the mouth of Little Chouse Brook.*

subrounded granules to cobbles composed of carbonate, blue quartz, vein quartz, quartzite, massive psammite, alkali feldspar, greenschist and pelite. Almost all of these metasediments are poorly sorted and matrix-supported. The clast content of the metasediments varies throughout the area. At Sandy Lake and in the wooded area southeast of Hampden, well sorted, almost pure quartz pebble metaconglomerate is found. At the Trans Canada Highway, the quartz pebble metaconglomerate contains isolated dark clasts of massive, fine grained psammite. Locally, the metaconglomerate is clast-supported. Metaclastic pebbly beds associated with marble generally display a strong carbonate fraction and also contain abundant clasts of both blue and vein quartz and quartzite; rarely, they contain buff psammitic clasts. Just west of White Point, white to pink weathering pebbly schists contain clasts of dominantly alkali feldspar and quartz. This is similar to a white to pale pinkish weathering pebbly conglomerate horizon that is associated with the coarse metaconglomerate of the Oody Mountain Amphibolite.

Petrography: In thin section, the semipelitic, pelitic and psammitic schists all display similar mineralogy and texture, but with different mineral proportions. The most common constituents include quartz, feldspar, biotite, muscovite, chlorite and garnet, with less abundant carbonate, amphibole, epidote, sphene, apatite, zircon, rutile, graphite and tourmaline. Quartz and feldspar in the metaclastic rocks are usually less than 2 mm in diameter and form a granoblastic to equigranular sutured matrix to any one of the following porphyroblasts: biotite, muscovite, feldspar and garnet. Feldspars include both twinned and untwinned plagioclase (albite to andesine) as well as microcline. The metaclastic rocks are generally foliated, with pelitic material defining the fabric, though locally it is delineated by amphibole. The biotite observed in thin sections is either reddish brown or dark olive green, but no obvious distribution of the coloration is recognized. The more mafic metaclastic rocks generally are enriched in chlorite, magnetite, epidote and carbonate, as well as locally containing both fine grained (<2 mm) idioblastic sphene and fine grained green hornblende.

CARBONATE SCHIST AND MARBLE: These are the most conspicuous variable rock types in the Outboard sequence. They occur in two general forms, (i) as schistose, impure carbonate schists and (ii) as massive, nearly pure marble and marble breccia. Locally, the two types are interlayered on a centimetre scale and form a massive, variegated variety (Plate 4-13). Carbonate schists occur throughout the unit, with the thickest sections found on the shoreline north of Penny Hills, at Shale Point, and at Little Pumbly Cove. These sections show little evidence of primary layering, although a local thin alternating muscovite-rich and carbonate-rich banding most likely represents bedding. The purest varieties, with only minor muscovite, appear as bright gold to buff colored zones, such as on the coast near the Penny Hills. The yellowish cast of this section indicates the presence of dolomite. Elsewhere, carbonate schists associated with amphibolite and greenschist are greenish brown due to contamination by chlorite, biotite, amphibole and epidote. In places, these schists completely enclose small enclaves of amphibolite up to 5 m long.

Massive varieties of marble occur as thick featureless layers, thinly banded zones, carbonate breccia and conglomerate beds. Some occurrences are apparently large scale blocks within metaclastic strata. These varieties range in color from steel blue to white, pink and buff; the pink and buff colors generally reflect a high content of dolomite. Thick, homogeneous layers up to 3 m wide are common along the White Bay coast south of Purbeck's Cove. At least 50 m of these layers with interspersed carbonate schist outcrop in a cliff face bordering the mouth of the brook just north of Otter Point. Generally, the massive marble is highly fractured, coarse grained and granoblastic, although thinner fine grained

Plate 4-13: *Variegated, impure marble of the White Bay Group on the southern shore of Purbeck's Cove.*

layers are present. Locally, near White Point, the massive marble is feldspar porphyroblastic and grades into carbonate schist. Zones up to 3 m wide of thinly layered (<3 cm), fine to medium grained, gray to white marble are interlayered with carbonate-rich metaclastic rocks on the upper portion of Rocky Brook and on the coast south of Clay Cove (Plate 4-14).

Marble conglomerate, breccia pods and breccia layers are common along the coast from Bear Cove to Big Chouse Brook, and inland along the brook and its tributaries. Generally, they form massive beds and lensoidal pods up to 2 m thick, but one massive unit approximately 12 m thick outcrops near the mouth of Big Chouse Brook. The least deformed breccias occur at White Point, Shale Point, Clay Cove, and near the mouth of Big Chouse Brook. These fragmental rocks range from polymictic metaconglomerate, with a high percentage of carbonate in either the clasts or the matrix, to pure carbonate breccias. The matrix is generally a fine to medium grained, buff gray to black marble. Locally, as in the Penny Hills area, the matrix of a metaconglomerate layer changes gradationally along strike from dominantly carbonate to dominantly pelite. The clasts range from granules to boulders up to 50 cm long. The metaconglomerate is generally ungraded, containing angular to rounded clasts of predominantly

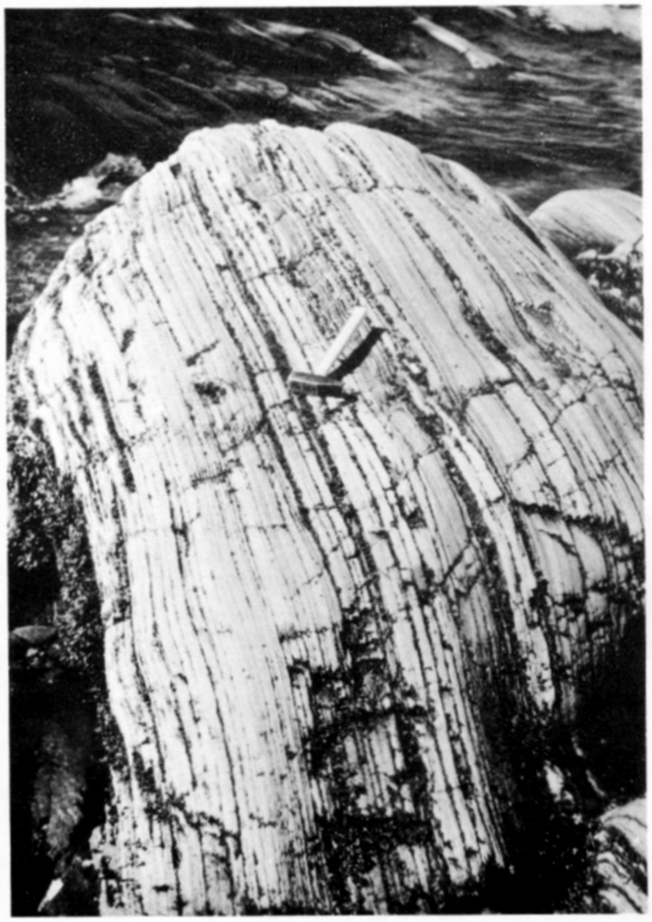


Plate 4-14: *Thinly layered, fine to medium grained marble of the White Bay Group along the coast south of Clay Cove, White Bay.*

white and pink carbonate, commonly accompanied by quartzite, vein quartz, psammite, pelite and greenschist clasts. At Purbeck's Cove, de Wit (1972) reported a metaconglomerate containing several rounded granitic clasts up to 30 cm in diameter. Carbonate clasts, particularly in the carbonate breccias, range in shape from flat, angular to subrounded (Plate 4-15) to nearly spherical. The former appear to represent remnants of thinly bedded carbonate, whereas some of the latter have forms that are reminiscent of colonial organisms (Plate 4-16). In places, such as Little Pumbly Cove, blocks of massive carbonate breccia up to 50 cm long are incorporated in the breccia beds.

Isolated massive blocks of buff, white, and pink marble occur in metaclastic rocks and in proximity to marble metaconglomerate beds. These blocks and pods are at least 15 m long, though at only one locality, near Twelve Mile Point, can it be shown that a massive buff dolomitic block is completely surrounded by semipelitic schist. At this locality, the metaclastic sequence appears to be continuous and nonrepetitive and the boundary of the block appears to be structurally intact. Elsewhere, as at White Point and on a lower tributary to Big Chouse Brook, pink and bluish white carbonate breccia and dolomitic layers, collectively ranging up to 7 m thick, are surrounded on three exposed sides by

Plate 4-15: *White Bay Group carbonate breccia composed mainly of flat angular slabs and minor subrounded carbonate clasts; White Point, Bear Cove.*

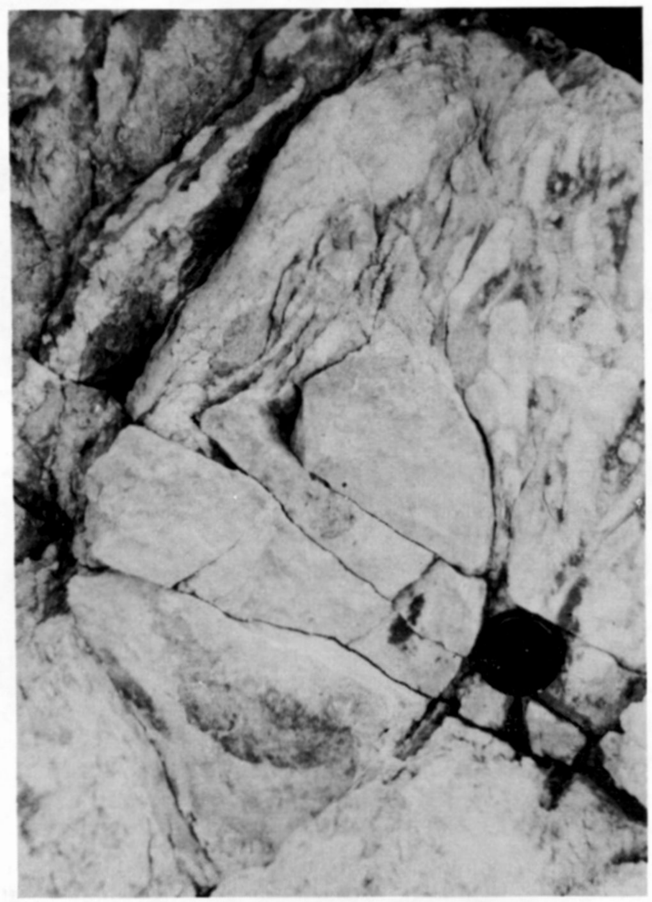


Plate 4-16: *Large subrounded clast (to left of lens cap), with a form reminiscent of colonial organisms. From the White Bay Group carbonate breccia at White Point, Bear Cove.*

metaclastic rocks. The nearest outcrops indicate that these carbonate units terminate abruptly on the unexposed sides. I consider these units to represent the tongues of large debris flows, channels, or depositional boulders within the metaclastic succession. De Wit (1972) reached the same conclusion for the marble outcrop at White Point (Plate 4-17). A large outcrop area of marble on the north side of Purbeck's Cove appears to be isolated within metaclastics, suggesting that it, too, may be a large block, but poor exposure precludes the interpretation of this occurrence.

Most of the marble is a xenoblastic mosaic of sutured calcite and dolomite, though locally quartz, feldspar and pelitic material are included. In the area south of the Beaches, where marble occurs near the Wild Cove Pond Igneous Suite, diopside is a significant constituent.

than 50 cm up to 10 m. At Twelve Mile Point, a continuous section of amphibolite and mafic schist is approximately 25 m thick.

In many places, the mafic schists are closely associated with marble, for example, at Twelve Mile Point, on the small island 1 km south of Little Pumbly Cove, and on the upper portion of Little Chouse Brook near the contact between the Outboard and Inboard sequences. At Twelve Mile Point, irregularly distributed, pea-sized concentrations of albite and epidote give the appearance of a primary fragmental texture to the amphibolite. In conjunction with the concordant and gradational nature of the metamafic rocks with surrounding strata as well as the occurrence of carbonate and metasedimentary interlayers, this suggests that schists were originally extrusive rocks.

Plate 4-17: *Pod of marble and marble breccia surrounded on three sides by metaclastic rocks at White Point, Bear Cove; this pod may represent a debris flow, a channel, or a depositional block; White Bay Group.*

AMPHIBOLITE AND MAFIC SCHIST: Massive dark to blackish green amphibolite and green-gray mafic schist layers and zones are common in the Outboard sequence. All observed metamafic rocks are concordant with the surrounding strata and, in many places, the contacts between these rocks appear to be gradational. A wide range of rocks exists between the pure, massive amphibolites and the metasedimentary rocks of the sequence. The metasedimentary rocks are intermediate in composition and are most common near the more massive amphibolite units.

The mafic schists range in form from massive amphibolitic layers, up to 10 m wide, to coarse grained schist zones. Locally, as in areas south of Shale Point and south of Otter Point, the amphibolites are garnetiferous. Layering, defined mainly by the concentration of epidote, is common; locally, metaclastic and carbonate pods and layers occur in these units. In the amphibolite, layers range in thickness from less

Petrography: Due to compositional variations of the metamafics, only generalizations can be made about their petrography. The most common constituents include amphibole, biotite, feldspar, epidote, chlorite, quartz, apatite, magnetite, ilmenite, sphene and rutile. Amphibole generally forms coarse porphyroblasts that define the main fabric in these schists; many show optical characteristics intermediate between actinolite and hornblende, though both end members have been identified. Biotite is porphyroblastic in almost all the mafic rocks of the sequence and ranges in color from reddish brown to olive green; locally, it is associated with subidioblastic garnet, as at Purbeck's Cove. Plagioclase is ubiquitous in the mafic rocks and appears to be mainly oligoclase. It occurs as both a fine grained, untwinned granoblastic groundmass and as idioblastic porphyroblastic crystals. Epidote and chlorite are relatively common, and locally define a millimetre to centimetre scale layering.

Chlorite commonly replaces amphibole, biotite and garnet in these rocks. Other minerals are accessory, though quartz is locally significant. In rocks gradational between amphibolite and the metaclastics, quartz forms a granoblastic to equigranular serrated groundmass for porphyroblasts of mafic minerals.

Inboard Sequence

The Inboard sequence of the White Bay Group forms the core of the coastal synclinorium. It outcrops mostly between the Penny Hills and Carrol Hill Slides, although a thin portion of the sequence outcrops to the east of the Carrol Hill Slide in the Western Arm area. Sparse younging evidence, within the sequence and nearby in the Outboard sequence (Figure 1-1), suggests that the east half of the Inboard sequence faces west, whereas the west half appears to face east. The stratigraphy of this sequence is more consistent than that of the Outboard sequence. The Inboard sequence is composed of a basal marble unit, and overlying units of amphibolite of the Garden Cove Formation, and metaclastic schists of Pigeon Island Formation. In general, the Inboard sequence is less pelitic and more homogeneous than the Outboard sequence.

De Wit (1972) mapped and described in detail the rocks that are here included in the Inboard sequence in the Western Arm area. He divided the rocks of the area into several units, including the White Point, Stuckless Cove, Walkers Cove, Pigeon Island, Back Cove, Rocky Point, and Garden Cove Formations. De Wit's White Point Formation included mainly carbonate rocks on the west side of the sequence; in this report, these have been left as an unseparated part of the Outboard sequence. During the present 1:50,000 scale mapping, the Stuckless Cove and Walkers Cove Formations were unmappable. The Back Cove and Pigeon Island Formations were not readily separable due to stratigraphic and tectonic interlayering. These four units are herein assigned to the expanded Pigeon Island Formation; local variations represented by de Wit's formations are considered as members. The mainly carbonate rocks of de Wit's Rocky Point Formation along the east side of the group are herein left unnamed, pending evidence that they form a distinct unit, separate from other carbonate rocks on the west side of the coastal synclinorium; the latter include his White Point Formation and other carbonate rocks along strike from it in the Outboard sequence. The Garden Cove Formation is unchanged from de Wit's definition, and has been extended southward into previously unmapped areas.

CARBONATE ROCKS

Mainly carbonate schist (Plate 4-18) and quartzite form the base of the Inboard sequence at Western Arm (Figure 1-1). Graded bedding in two layers at this locality suggests that the unit is westward facing. De Wit (1972) described these rocks in detail as follows:

> A thick sequence of about 250-300 meters of banded (1-20 cm) light yellow calcareous-quartzite schists is followed towards the top by about 100 m of a dominant yellow-brown carbonate schist sequence .... Sporadic graphitic and garnet schists occur and also some thin (10 cm - 2 m) gray marble bands. Throughout the formation, light green metavolcanic horizons (see next formation) occur up to 2 m thick [Plate 4-19], which at their contacts are interlaced on a mm-cm scale with discontinuous metasediment bands.

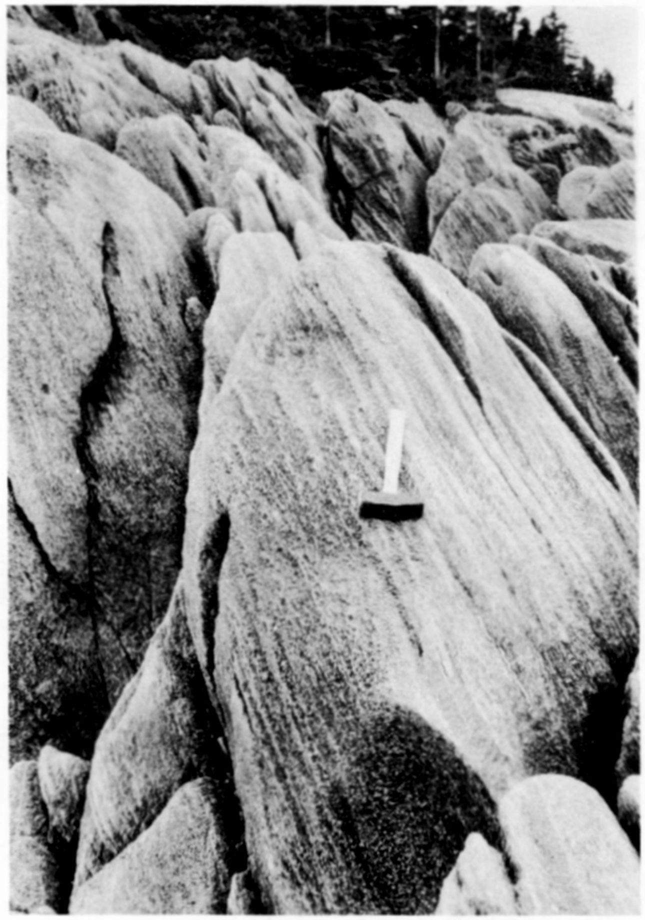


Plate 4-18: *Vaguely banded carbonate schist just southwest of Herbert Point, Western Arm; White Bay Group.*

> The quartzites have an extremely regular and evenly spaced schistosity.... Differential weathering leaves a very pronounced schistose platy quartzite, referred to in the field as "the paper schists". They range in color from white to orangey-yellow to an evenly or spotted reddish brown, the result of increase in carbonate and hæmatite respectively. Gradual or sharp increases of these and muscovite add to the definition of the bedding.
>
> Towards the top, in the carbonate schists, marble conglomerate bands occur. On the shore opposite Pigeon Island, several thin continuous bands are up to 40 cm in thickness. Pink, well rounded, coarse grained marble clasts up to 5 cm in diameter, are accompanied by pelitic schists, basic metavolcanic and quartzite clasts of more elliptical disc shape, up to 20 cm in length. These bands are identical to the White Point Formation marble conglomerate [included here with the Outboard sequence] but differ in that clasts other than marble are more common.

The uppermost carbonate schists and marble conglomerates are conformably overlain by massive black-green amphibolite of the Garden Cove Formation on the coast east of Pigeon Island (Plate 4-20), in Western Arm, on the Westport road, and at the pond immediately east of Purbeck's Cove. At the latter two locales, the carbonate units appear very thin (<2 m), though this may be a function of poor exposure. The carbonate schist unit appears to be absent on the west side of the synclinorium by the Penny Hills Slide, though portions of the Outboard sequence may represent this unit, as discussed below.

Plate 4-19: *Mafic metavolcanic lens surrounded by and interlayered with carbonate schist; White Bay Group on shoreline opposite Pigeon Island, White Bay. The lens is approximately 1.5 m thick.*

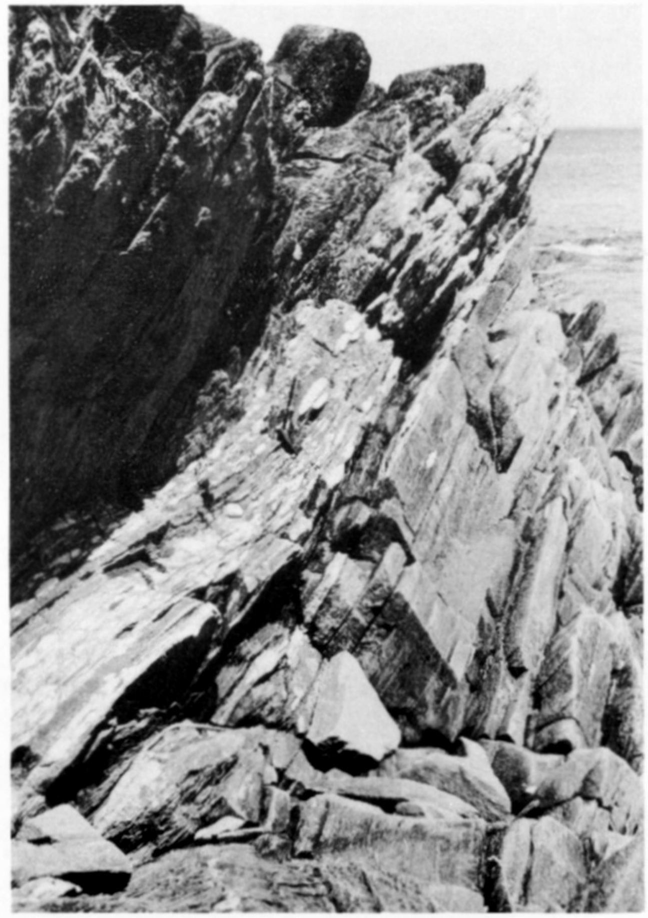


Plate 4-20: *Conformable contact between White Bay Group marble and carbonate metaconglomerate and the structurally overlying amphibolite of the Garden Cove Formation. On the coast just east of Pigeon Island, White Bay.*

GARDEN COVE FORMATION

The Garden Cove Formation (Church, 1969; de Wit, 1972) comprises layered, massive and schistose amphibolite, epidosite, epidote-biotite schist, and minor metasedimentary rocks that outcrop in two bands, up to 500 m wide on opposing limbs of the coastal synclinorium (Figure 1-1). Both belts are concordant with the surrounding stratigraphy. The eastern band is discontinuous and outcrops for 5 km south from the shore of White Bay where it pinches out; it also outcrops for approximately 3 km strike length at the pond immediately east of Purbeck's Cove. This distribution may be primary, but more likely reflects the effects of the Carrol Hill Slide (see Chapter VII). More detailed mapping southward along strike may prove this belt to be more extensive. The western belt of the formation is continuous for 15 km from Bear Cove to the area south of Purbeck's Cove. Possible equivalents of this belt occur on Little Chouse Brook, though poor exposure over the wide intervening area precludes the linkage of these outcrop areas on the map.

Banding within the unit ranges from millimetre scale layers of epidote to massive zones up to 10 m wide defined by either massive amphibolite or amphibolitic clasts set in an epidote-rich matrix. Both varieties occur at Wild Cove, Western Arm as well as in Big Cove. At the mouth of Wild Cove Brook, Western Arm, elliptical amphibolite and epidote clasts up to 40 cm in length are set in an apple green to yellow-green, coarse grained, epidote-rich matrix (Plate 4-21). Locally, rounded pink carbonate and quartzite clasts occur in this unit. The fragmental rock appears to be conformable and locally gradational with massive amphibolite to the east. The fragmental nature of this rock as well as the inclusion of metasedimentary clasts suggests that this member of the Garden Cove Formation represents a volcanic conglomerate. Elsewhere in the formation, de Wit (1972) reported the occurrence of well defined, light green epidote ovoids (<5 mm diameter) with dark massive amphibolite; he suggested that these are relict amygdules. Locally, thin zones up to 2 m wide of metasediments are interlayered with the amphibolite. At Garden Cove and on the pond east of Purbeck's Cove, layers of marble and marble conglomerate up to 25 cm thick occur within the unit; their extent is unknown due to poor exposure. Similarly, a few metres of feldspathic psammite are interlayered within the metamafics in Stuckless Cove. It appears from the foregoing description that part, if not all, of the

Plate 4-21: *Fragmental aspect of the Garden Cove Formation at the mouth of Wild Cove Brook, Western Arm; White Bay Group.*

Garden Cove Amphibolite was originally an extrusive mafic sequence with minor sedimentary interlayers.

Petrography: The most common constituents of the formation include hornblende, plagioclase, epidote, biotite, quartz, garnet, sphene, rutile, ilmenite, magnetite and apatite. De Wit (1972) described the occurrence of these minerals as follows:

> Hornblende... textures vary from decussate to nematoblastic.... In addition, large poikiloblastic hornblendes (3-4 mm) occur in the thin amphibolite bands where interlaminated with calcite bands. The hornblende is commonly zoned to a pale colored actinolitic amphibole.
>
> Clear plagioclase (albite and/or oligoclase) occurs as part of a granoblastic mosaic with quartz and hornblende or as small equant porphyroblasts (<0.5 cm) and aggregates thereof. Grain boundaries are irregular... At the edge of the plagioclase, quartz may be vermicularly intergrown. A slight zoning in the plagioclase is common. Equant epidote granules (<0.1 mm) are the most common inclusions, frequently aligned as well developed earlier relic banding. Often the epidotes are restricted to a central zone, leaving a clear outer rim. Elsewhere, light yellow epidote occurs as small subidioblastic grains, dispersed throughout the matrix, or subpolygonally in folia.
>
> Biotite (green and brown) is common, and concentrates in particular bands. It displays similar habits as the hornblende, with which it is commonly intergrown. Chlorite is nearly always present as an accessory, interleaved with hornblende or as large random flakes overgrowing the matrix.
>
> The absence of garnet throughout most of the amphibolite is conspicuous. It is only seen to be well developed at Big Cove. Here, it occurs as small idioblasts or coagulates in masses in veins. The absence elsewhere is probably a reflection of the original rock composition.
>
> Rutile is consistently present in all but the calcite rich bands. It is commonly present in large quantities, as single crystallites or small dispersed aggregates, but often common as long trains or thin (0.2 mm) bands, parallel to the banding, and frequently associated with the epidote bands. Some bands are rich in rutile, others have little. Sphene is also common, but not in bands. Ilmenite and magnetite are frequent. The latter is concentrated in several bands, as large octahedra or xenoblastically dispersed throughout the matrix.... Apatite is a ubiquitous accessory.

PIGEON ISLAND FORMATION

This formation is a varied, mainly metaclastic assemblage of garnetiferous semipelitic schist, feldspathic psammite, graphitic schist and quartzite with only minor carbonate schist and amphibolite. It forms the core of the coastal synclinorium and extends for the length of the White Bay Group, north of the Wild Cove Pond Igneous Suite. Its maximum outcrop width is approximately 5 km in the area of the Beaches. Psammitic portions of the formation are rich in magnetite; this is reflected in the distinctly high magnetic signature of the unit on the aeromagnetic map of the area (Geological Survey of Canada, 1970).

Semipelitic schist and feldspathic psammite form most of the unit, but are not divisible on a regional scale. They are in tectonic contact with the Garden Cove Formation in Bear Cove whereas in Wild Cove the contact is transitional over 1 m in a zone of high strain. Quartzite and graphitic schist tend to form distinct members that directly overlie the Garden Cove Formation on the east limb of the coastal synclinorium in Western Arm; however, here, the section is disrupted by the Carrol Hill Slide, approximately 10 m above the amphibolite. The quartzite and graphitic schist also occur elsewhere in the formation.

Semipelite and psammite form regularly layered beds up to 50 cm thick and locally include thin interlayers of gray to brown pelitic schist. Light gray to rusty weathering semipelitic schist is the most conspicuous rock type of the formation, and tends to occur in the central to eastern portions of its outcrop area. It is generally a coarse garnetiferous quartz-muscovite schist with minor biotite, chlorite and tourmaline. In thin section, accessory calcite, magnetite, ilmenite, epidote, chloritoid, sphene and apatite are also present. Locally, south of Little Chouse Brook, sillimanite occurs where the semipelite borders the Wild Cove Pond Igneous Suite. The following two major features characterize the semipelitic schist: (i) large, burgundy colored garnets, spectacularly developed (up to 3 cm) on Pigeon Island, are typical in these schists, though in the area south of Little Chouse Brook, only minor pinhead-size garnets were found within the semipelite, and (ii) ubiquitous thin white quartz veins and pods up to 2 m long parallel the schistosity in these rocks.

The feldspathic psammite weathers gray, gray-green, buff and brown on the coast, but inland is chalky white, with prominent feldspar porphyroblasts. Typically, it is interlayered with brown semipelite (see Plate 7-3). Generally, magnetite occurs as both isolated octahedra and thin beds and clusters in the psammite. In thin section, the plagioclase ranges from albite to oligoclase (de Wit, 1972) and occurs as porphyroblasts and poikiloblasts that generally enclose fine grained quartz and garnet. Chlorite, biotite and muscovite are common in these generally massive schists.

Graphitic schist is common along the Carrol Hill Slide; it occurs west of the slide in Garden Cove and on the Westport road and locally along the slide to the south. The slide appears to be keyed to this incompetent unit. Graphitic schist is also interlayered with garnetiferous semipelitic schist on the northeast side of Pigeon Island.

The graphitic schist is plagioclase porphyroblastic with a heavy dusting of graphite clearly defining inclusion fabrics within the feldspar. In addition, muscovite, biotite, epidote, chlorite, quartz and garnet are common.

At Garden Cove, to the east of the Carrol Hill Slide, a 2 m section of thinly banded light buff quartzite is interlayered with Pigeon Island psammite and semipelite that directly overlie the Garden Cove Formation (Plate 4-22). De Wit (1972) noted a similar quartzite member at Stuckless Cove.

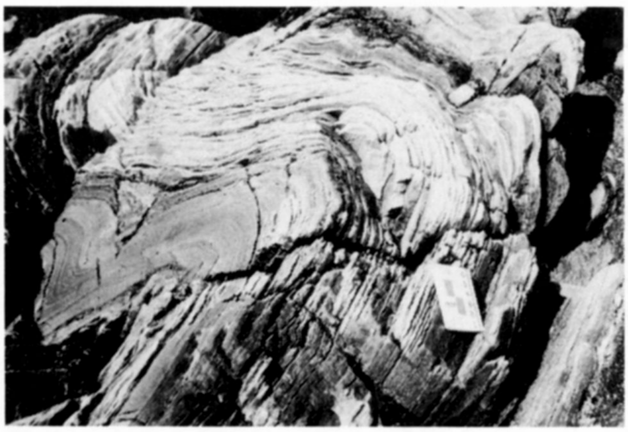


Plate 4-22: *Thinly banded, fine grained Pigeon Island quartzite at Garden Cove, Western Arm; White Bay Group.*

Minor amphibolite layers and carbonate schist zones occur sparingly within this mainly metaclastic formation. The amphibolite layers range up to 5 m wide, and outcrop generally in the area north of Carrol Hill. Approximately 200 m of gray to buff weathering carbonate schist with metaclastics and minor marble conglomerate are found on the coast at Dark Gulch Point. The inland extent of this member is uncertain, though it is probably terminated along a fault on its east flank (Figure 1-1).

Contact Relationships

INTERNAL CONTACTS

The Penny Hills Slide forms the contact between the Outboard and Inboard sequences in the northern portion of the White Bay Group. It extends from Back Cove to the area just south of Purbeck's Cove. In many places along the slide, marble and carbonate schist of the Outboard sequence are juxtaposed either against, or close to, the Garden Cove Formation; this forms a succession which closely resembles that at the base of the Inboard sequence on the east limb of the coastal synclinorium. South of the latitude of Purbeck's Cove, the contact between the sequences is unexposed, nor have slide zone rock types been encountered in the vicinity along strike. The contact between the two sequences appears to be conformable at Little Chouse Brook. Here, a section of over 150 m of greenschist, carbonate schist, marble and garnetiferous semipelite appears to be conformable with structurally underlying amphibolite to the east. The amphibolite is not well exposed, but appears to span greater than 30 m of section; it is conformable with Pigeon Island psammite upstream to the east. Thus, this contact suggests that the two sequences were originally conformable. The carbonate schist on Little Chouse Brook is probably equivalent to the basal carbonate schist unit on the east side of the Inboard sequence at Western Arm, though on the opposing limb of the coastal synclinorium. From this it can be inferred that the Penny Hills Slide represents a bedding plane translation, wherein carbonates of the Outboard sequence, along the fault, may be close to their original stratigraphic position. In addition, it indicates that the Outboard sequence is largely older than the Inboard sequence. This line of reasoning is confirmed by the contact relationships of the White Bay Group with surrounding rock units.

EXTERNAL CONTACTS

It appears that the Outboard sequence was deposited on a predeformed crystalline basement. Relationships of the amphibolite and granitic gneiss in the Oody Mountain Amphibolite indicate that the amphibolite intrudes and includes screens of a predeformed terrane. I suggest that the Oody Mountain Amphibolite either masks an unconformity or marks an nonconformity between an older gneissic terrane, that was probably removed along the Cabot Fault, and the overlying White Bay Group.

The contact between the Outboard sequence and the Old House Cove Group is ambiguous in the Westport area. The contact occurs in a highly tectonized zone between Pound Cove and Eastern Head which is characterized by numerous faults of uncertain significance. In this tectonized zone, the strata are transitional from the more pelitic and varied rocks typical of the White Bay Group to the more psammitic Old House Cove Group. Carbonate and graphitic schist typical of the White Bay Group, but either rare or absent in the Old House Cove Group, decreases in abundance northward through the contact zone. Thus, the contact between the groups is somewhat arbitrarily drawn along an obvious tectonic break south of which White Bay Group rock associations predominate. I suggest that the original contact between these units was a conformable, lateral transition, based on the nature of this zone.

The Inboard sequence is conformable with the Old House Cove Group on the eastern limb of the coastal synclinorium. The contact is exposed near Rocky Point, east of Garden Cove, and on the coast east of Pigeon Island. At these localities, there is a transition, over approximately 100 m, from psammitic schist of the Old House Cove Group to the basal calcareous schist and quartzite of the Inboard sequence. Facing directions from pebbly layers in the Inboard sequence on the shores of Western Arm and on the White Bay coast all indicate that the White Bay Group, here, overlies the Old House Cove Group. South of the Westport road, the contact between the two units is the Carrol Hill Slide. On the west limb of the coastal synclinorium, the Inboard sequence is faulted along the Penny Hills Slide against a small sliver of the Old House Cove Group.

The relationships outlined above indicate that the White Bay Group has two different substrates, namely, the crystalline basement and the Old House Cove Group, and yet, apparently, is laterally transitional with the Old House Cove Group. These observations lead to the interpretation that the Old House Cove and White Bay Groups are at least partial lateral equivalents representing different depositional facies resting on crystalline basement (see Old House Cove Group). Thus, the White Bay Group is here considered to be diachronous with the older Outboard sequence overlying basement and interdigitating with the Old House Cove Group, whereas the Inboard sequence represents a stratigraphically younger tongue of the White Bay Group that, in part, overlies the Old House Cove Group. This interpretation is in accord with the stratigraphy erected within the White Bay Group and the correlation of the carbonate units on opposing limbs of the coastal synclinorium.

Depositional Environment

The White Bay Group contains the best preserved primary structures in all of the Fleur de Lys Supergroup; hence, it divulges the most information concerning the original depositional environment of Fleur de Lys protoliths. In particular, the occurrence and general features of the marble and carbonate schists are most revealing. The spectacular calcareous metaconglomerates, large marble blocks, and thinly bedded marble and calcareous schists interlayered with dominantly thinly layered metaclastics and sulfurous graphitic schists closely correspond with descriptions of carbonate slope facies marginal to a carbonate bank (e.g. Wilson, 1969; McIlreath and James, 1979). The marble breccia layers of the White Bay Group most likely represent debris sheets emplaced in a slope environment; the thin, flat nature of many of the carbonate clasts suggests that the deposits were slope-derived, rather than platformally derived (McIlreath and James, 1979). The large isolated pods of marble breccia and massive marble may be tectonized equivalents of either debris flow channels or allochthonous submarine glide rafts. In the light of these associations, the pelitic carbonate schists most closely resemble hemipelagic deposits of periplatformal ooze (McIlreath and James, 1979), and more massive marble layers may originally have been turbidite-derived calcarenites. The slope environment envisioned for the White Bay Group is also suggested by the thinly interlayered metaclastics and graded pebbly metaclastics that are reminiscent of distal turbidite sequences.

The greenschist and amphibolite in the sequence are generally associated with the carbonate rocks, and indicate that there was an igneous, probably extrusive, influence in the slope environment. Even the basal Oody Mountain Amphibolite lies within 100 m of a marble layer at the Beaches, indicating that the presumed older parts of the sequence are most likely submarine. The petrochemical features of the amphibolite, as outlined later in this report, most closely resemble those of rift-type tholeiites. Hence, the White Bay Group most likely represents submarine rift and slope deposits that appear to have bordered a substantial carbonate bank.

Age and Correlation

Based on regional correlation, Murray (1881), Howley (1902), and de Wit (1972) have all considered rocks of the White Bay Group to be largely Paleozoic. One dorsal valve of an inarticulate brachiopod with an affinity to the Order Acrotretida has been recovered from the marble breccia block at White Point (S. Stouge, personal communication, 1979). This fossil indicates that the marble is not older than Hadrynian. From the foregoing discussion of stratigraphy, this would imply that the whole Inboard sequence is Cambrian or younger, whereas the Outboard sequence could be older.

More specific dating of portions of these rocks is only possible through regional lithostratigraphic correlations of the group with less deformed, fossiliferous sequences of the Humber Zone. The basal section of the Outboard sequence in the Hampden area bears a marked resemblance to the lower portion of the Labrador Group (as defined by James et al., in preparation) on the Great Northern Peninsula of Newfoundland. The Oody Mountain Amphibolite is strikingly similar to the basal Bateau Formation and Lighthouse Cove Volcanics of the Labrador Group. Both assemblages appear to overlie crystalline basement; the Bateau Formation overlies Grenville basement on Belle Island (Williams and Stevens, 1969), whereas the Oody Mountain Formation contains screens of granitic gneiss that are most likely the remains of a more extensive basement complex. The screens of granitic gneiss are lithically very similar to Grenville gneisses of the Long Range Inlier on western White Bay. Conglomerates in both the Oody Mountain and Bateau Formations contain predeformed gneissic boulders and cobbles that were probably derived from the underlying gneissic terranes. In addition the Oody Mountain Amphibolite is geochemically correlative with the Lighthouse Cove basalts (see Chapter VI) that overlie the Bateau Formation. There are also other rocks associated with the Oody Mountain Amphibolite that resemble other parts of the Labrador Group. Quartz-rich pebbly metaconglomerate of the White Bay Group, that occurs along the Trans Canada Highway, and inland directly east of the Oody Mountain Amphibolite, bears close resemblance to the Bradore grits of the Labrador Group in Canada Bay; locally, the grits in Canada Bay unconformably overlie Grenville basement (Smyth, 1973).

Based on these lithic and geochemical similarities, I consider the White Bay Group in the Hampden area as correlative with the lower part of the Labrador Group. The lower part of the Labrador Group underlies the Lower Cambrian Devils Cove Formation (Betz, 1939), which is the oldest known carbonate in the cover sequences of western Newfoundland. In the White Bay Group, the Oody Mountain Amphibolite and associated metaclastics appear to underlie the lowest carbonate unit in the Outboard sequence. Considering this relationship and the correlation of part of the White Bay and Labrador Groups, I suggest that the Oody Mountain Amphibolite and associated metaclastics are most likely Late Hadrynian to earliest Cambrian in age. Thus, the remainder of the group is probably Early Paleozoic. Regional structural constraints pertaining to the Early to Middle Ordovician transport of ophiolites over the Fleur de Lys terrane indicate that these rocks can be no younger than Early Ordovician (see Chapter VII).

Correlatives of the White Bay Group are also found in the Fleur de Lys Belt. The Middle Arm metaconglomerate (see East Pond Metamorphic Suite) is identical to the metaconglomerate in the Oody Mountain Amphibolite; both are in the same setting, close to predeformed crystalline basement, and have similar clast assemblages. I consider the metaconglomerates to be mutually correlative. Thinly layered, quartz-rich psammitic and semipelitic schists of the East Pond Metamorphic Suite may thus be in part correlative with similar rocks of the White Bay Group.

Lithic correlatives of part of the White Bay Group also occur on Glover Island to the south of the Baie Verte Peninsula (personal observation). There, the section consists of amphibolite, marble, graphitic schist and pebbly metaclastics that structurally overlie a granitic gneiss (Knapp et al., 1979). The amphibolite, similar to the Oody Mountain Amphibolite, directly overlies the gneiss and is texturally similar to the Oody Mountain rocks. It is structurally overlain by other rock types that are very similar to the Outboard sequence on White Bay. The setting of the Glover Island rocks is different from that of the White Bay Group; the Glover Island sequence lies along the east side of the Fleur de Lys Belt against the southward extension of the Baie Verte Line (Knapp, 1980), whereas the White Bay Group lies on the west side of the Fleur de Lys Belt.

OLD HOUSE COVE GROUP

Definition and Extent

The Old House Cove Group is a new name proposed here for the dominantly psammitic and semipelitic schists that outcrop over a large portion of the west half of the Baie Verte Peninsula (Figure 1-1). The group extends along the White Bay coast from Partridge Point south to Big Cove, and from Tom Cod Point south to Fox Cove. In the area of Western Arm, it outcrops on both sides of the White Bay Group, thus accentuating the definition of the coastal synclinorium.

Very few primary structures are preserved in the group. Considering this feature as well as the polydeformed nature of these rocks, erection of a type section is impractical for this unit without more detailed work. Typical exposures of the group at Old House Cove and at the mouth of Southern Arm on White Bay are identified here as representative sections. The stratigraphic thickness of the group is not known, though the maximum outcrop width is approximately 10 km just south of the Little Lobster Harbour Fault. Amphibolite pods and layers, common throughout the group, are discussed in a later section of this report with other prekinematic igneous rocks.

Fuller (1941) first considered these rocks in the area of Partridge Point as a part of his Headlands gneiss. Kennedy (1969, 1971) later referred to the same rocks as the Plat Bay Formation, a division of his White Bay sequence. Along strike, in the area of the Westport road, the same schists were considered as part of the Middle Arm Brook and Western Arm Formations of the Rattling Brook Group by Church (1969). De Wit (1972) reassigned these rocks to the Seal Cove Group. Since the names White Bay, proposed by Kennedy (1971), and Seal Cove, proposed by de Wit (1972), are pre-empted by their stratigraphic usage elsewhere in Newfoundland, the name Old House Cove Group is herein proposed for these rocks.

De Wit (1972) recognized three divisions of his Seal Cove Group, including the Southern Arm, Middle Arm, and Middle Arm Pond formations. In this study, the first two formations were found to be unmappable outside of de Wit's area, as well as over large parts of the inland area. Hence, these formations are not recognized in this report. The Middle Arm Pond metaconglomerate has been referred to herein as the Middle Arm metaconglomerate and included in the East Pond Metamorphic Suite. De Wit (1972) also separated small inliers along the White Bay coast, notably at Pound Head, Crow Head, near Fish Point, and near the Penny Hills, from the Fleur de Lys rocks, and assigned them to his predeformed "basement." However, these rocks appear identical to those designated here as the Old House Cove Group, with the exception of a unique felsic stock at Crow Head. All of these rocks, except the felsic stock, are here included in the Old House Cove Group.

The area between Baie Verte and the Wild Cove road is largely underlain by rocks, designated here as Old House Cove Group, that are bounded by the Rattling Brook Group. Exposures in this area are very limited, though the few present display metasediments and associated amphibolite typical of Old House Cove Group. It is possible that these rocks represent a facies similar to the Old House Cove Group within the Rattling Brook Group, but the overall aspect of the psammite and the presence of abundant amphibolite in this area suggest that the rocks more likely belong to the Old House Cove Group, and have been infolded with the surrounding Rattling Brook Group.

Description

The Old House Cove Group is an extremely monotonous sequence of alternating coarse feldspathic psammite and semipelite with minor interlayers of graphitic schist, pelite, metaconglomerate, garnetiferous semipelite and quartzite; minor marble has been reported from the group (Kennedy, 1969). In the area south of Tom Cod Point, the Old House Cove metasediments are generally more thinly banded and more graphitic than the main outcrop belt to the east.

Psammite of the unit is typically buff to gray weathering, coarse quartz-feldspar-biotite-muscovite schist. Locally, as at Seal Cove, the more massive layers are interbedded with thin graphitic schist beds and contain angular fragments of graphitic schist (Plates 4-23, 4-24). The semipelites characteristically contain large (up to 1 cm) plagioclase porphyroblasts. Pebbly metaconglomerate is common near the mouth of Southern Arm; it is composed of small clasts (up to 3 cm) of quartz, feldspar and graphitic schist in a psammitic matrix. Quartzite, as massive ledges up to 2 m thick, is rare on the coast north of Seal Cove. Bursnall (1979) reported a 20 m thick section of quartzite from the White Bay coast, approximately 2 km north of Little Lobster Harbour. It is fine grained, strongly foliated, concordant with surrounding metasediments, and contains ribbon quartz and white mica. He noted that it has the character of a tectonic slide, although its origin cannot be confirmed without more detailed work.

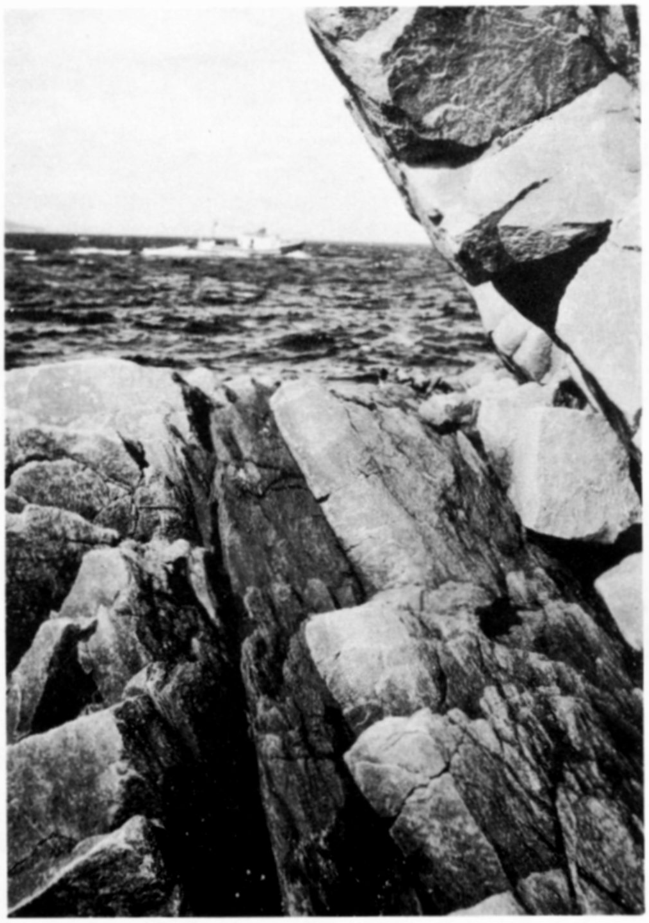


Plate 4-23: *Graphitic schist layer, approximately 30 cm wide, interlayered with massive psammite at the mouth of Seal Cove, White Bay; Old House Cove Group.*

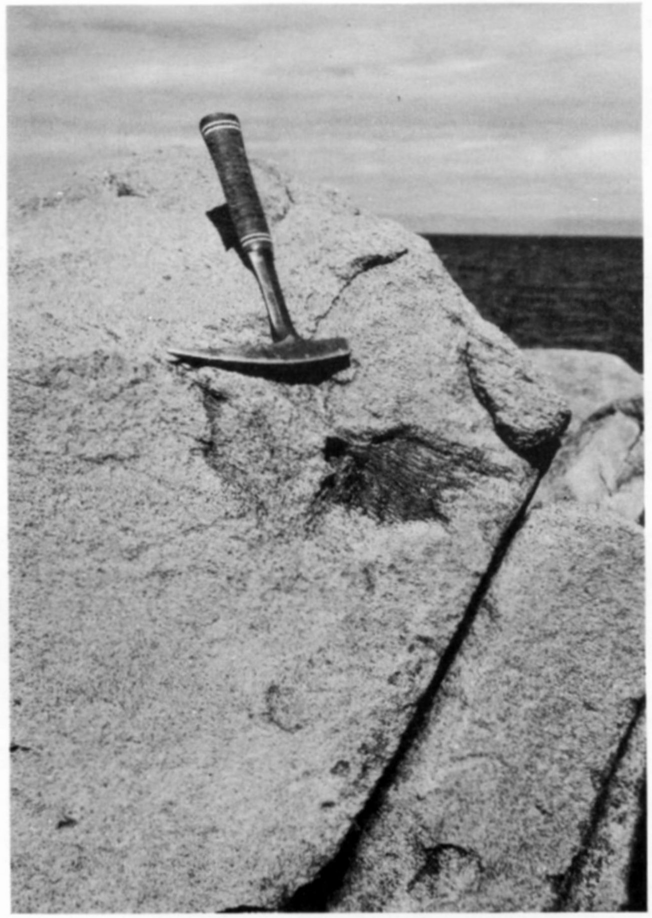


Plate 4-24: *Angular clast of graphitic schist within massive Old House Cove psammite layer at the mouth of Seal Cove, White Bay.*

Garnetiferous semipelite, identical to that in the Pigeon Island Formation of the White Bay Group, occurs sporadically throughout the Old House Cove Group. In particular, it forms a thin, discontinuous zone for approximately 20 km along strike near the contact of the group with the Inboard sequence of the White Bay Group. Similar semipelite occurs in Hard Bay, where Kennedy (1969) reported impure marble. This association in Hard Bay may indicate that the White Bay Group, though unexposed, is stratigraphically nearby. De Wit (1972) reported limited, thin bands of apatite and of heavy minerals from the group.

Petrography

The metasediments contain feldspar, quartz, muscovite, biotite, chlorite and garnet with subordinate epidote, clinozoisite, calcite, apatite, sphene, tourmaline, rutile, allanite, zircon and opaques. In addition, kyanite, sillimanite and staurolite occur rarely within the group. Plagioclase is the most abundant feldspar; it ranges from albite to oligoclase in composition. It is the most conspicuous component of these rocks, as it commonly forms porphyroblasts and poikiloblasts, generally <5 mm across, that either predate or overprint the main foliation in the metaclastic rocks. The poikiloblasts are generally studded with small (<1 mm) idioblastic garnets and many of the poikiloblasts contain inclusion trails defined by quartz and muscovite. Locally, the plagioclase forms an equant granoblastic groundmass with quartz. Quartz is very abundant, either as a sutured, equigranular mosaic or as granoblastic to sutured layers and pods. It is particularly common within plagioclase poikiloblasts and as the main constituent of pressure shadows on feldspar porphyroblasts.

Sparse microcline has been observed in three samples from the area south-southwest of the pond just east of Purbeck's Cove, and in the area west of Purbeck's Brook; otherwise, alkali feldspar is conspicuously absent from the Old House Cove Group. Small garnets are very common in all of the metaclastic rocks. Where garnet forms individual porphyroblasts independent of the plagioclase, it is locally altered to chlorite and biotite.

The platy minerals biotite, muscovite and chlorite define the main foliation in the semipelitic and pelitic rocks; in the massive psammite, the main fabric is generally a lineation defined by elongated quartz and feldspar crystals. Minor amounts of small idioblastic epidote and clinozoisite are present in places. Calcite is very rare, but forms an interstitial matrix between quartz and feldspar in some of the psammite.

The remaining minerals, including sphene, tourmaline, rutile, allanite, zircon and opaques, are common accessories.

Contact Relationships

The Old House Cove Group is in conformable contact with the White Bay and Rattling Brook Groups and in tectonic contact with the East Pond Metamorphic Suite. Detailed descriptions of its contacts with the White Bay and Rattling Brook Groups are included with the descriptions of these units. A distinctive mappable unit of tectonic schist occurs along the contact of the Old House Cove Group and East Pond Metamorphic Suite, and is described in a section following the prekinematic rocks. Where these schists are absent, the contact is assumed to be faulted, although locally, northeast of the Wild Cove Pond Igneous Suite, these units may be infolded (see East Pond Metamorphic Suite).

Depositional Environment

Since most primary features of the Old House Cove Group have been obliterated by regional tectonism, it is difficult to assess its depositional environment. The alternation of layered psammite and semipelite with minor graphitic schist and the overall monotony of the sequence are reminiscent of flysch deposits. Considered with the medium to thick layering so prevalent in the unit, this suggests that the Old House Cove Group originated as a basinal turbidite sequence.

Age and Correlation

The age of the Old House Cove Group is uncertain. Metaclastic rocks similar to those of the Old House Cove Group are associated with the East Pond Metamorphic Suite to the northeast of Wild Cove Pond, although those of the East Pond Metamorphic Suite are thinner, and generally less feldspathic. Since the East Pond metaclastics appear to overlie Grenville basement, lithic correlation of these units would imply a Hadrynian to Cambrian age for the Old House Cove Group. The apparent lateral equivalency of the Old House Cove Group with the Outboard sequence of the White Bay Group also suggests a similar age.

De Wit (1972) correlated the Old House Cove Group with less deformed rocks of inferred Eocambrian-Cambrian age found further west in the Humber Zone; specifically, he cited the Bateau and Bradore Formations and the allochthonous Maiden Point Formation (Smyth, 1973) as lithic correlatives of the group. The occurrence of basalt in the Maiden Point Formation and amphibolite in the Old House Cove Group strengthens this correlation. Furthermore, all of these basic rocks are petrochemically similar to the Eocambrian-Cambrian Lighthouse Cove Volcanics (see Chapter VI; de Wit and Strong, 1975; Smyth, 1973). On the basis of these lithic and geochemical correlations, the Old House Cove Group is probably of Eocambrian to Cambrian age.

RATTLING BROOK GROUP

Definition and Extent

The Rattling Brook Group is the varied assemblage of medium to coarse grained, mainly metaclastic schists with subordinate marble, amphibolite and greenschist that outcrop near the eastern boundary of the Fleur de Lys Belt. The overall character of the Rattling Brook Group changes across strike. From west to east, the unit is progressively more pelitic, more diverse and more strongly foliated and tectonically disrupted; however, its amphibolite content decreases.

The unit outcrops in two distinct structural belts separated by the Little Lobster Harbour Fault. The southerly belt is generally north-northeast striking and steeply dipping, and outcrops in a Y-shaped pattern (Figure 1-1). Its maximum outcrop width is approximately 4 km at the junction of the two outcrop limbs along Baie Verte Brook. The group thins in the Middle Arm Brook area to less than 1 km width. In this region, termed the Eastern Slide Zone by de Wit (1972) (see Chapter VII), deformational effects are extremely intense, so that the Rattling Brook and Old House Cove Groups are largely indistinguishable. Where the Rattling Brook is recognizable in this area, its lithic divisions are inconsistent along strike.

The northerly belt of the Rattling Brook Group is disposed in a major monocline that Kennedy (1969, 1971) termed the Baie Verte fold. The group has moderately to steeply dipping structures in eastern portions of this belt that dip progressively more moderately toward the west (Figure 1-1). This fanning of dips largely accounts for the wider outcrop pattern (up to 8 km) of the northern belt than the southern belt. In addition, J.T. Bursnall (personal communication, 1980) speculated that there may be a fold core containing Old House Cove Group rocks in the area north of Fleur de Lys Harbour, but rocks diagnostic of either group are absent in this area. Because of this uncertainty, the rocks in question are left in the Rattling Brook Group.

All characteristic rock types of the group occur in both belts. These include feldspathic psammite and semipelite, garnetiferous semipelite, quartz-muscovite semipelite, pelite, graphitic schist, amphibolite, greenschist and marble. In places, these rock types form definable belts, such as the graphitic schist units to the east of the Wild Cove road and on the Fleur de Lys road; generally though, their distribution is less clearly defined. The complicated interfingering and lensing of these distinct units may be attributed to either complex depositional facies relations or structural transposition, or both. Due to this complex pattern, neither the top nor the base of the group is defined.

Besides being more varied, the group is more pelitic than the Old House Cove Group. The overall lithic character and stratigraphic associations bear a remarkable similarity to those of the White Bay Group. However, the geographic separation of these units, as well as the less consistent stratigraphy of the Rattling Brook Group and its relationships with the Birchy Complex, justify the separation of the Rattling Brook and White Bay Groups.

The Rattling Brook Group, as defined here, follows the original usage of Watson (1947). He considered the group equivalent to the Fleur de Lys area "gneisses" (Fuller, 1941). Kennedy (1969, 1971), mapping in the Fleur de Lys area, disregarded the term Rattling Brook Group, though it encompasses rocks he mapped as the Harbour sequence and portions of his White Bay and Eastern sequences. One of his units, the Flat Point Formation of the Eastern Sequence, is recognized here. Church (1969) informally extended the Rattling Brook Group southward to the latitude of Flat Water Pond and expanded the group to include the mafic Birchy

Schist (Fuller, 1941). De Wit (1972), in turn, reduced the group by recognizing and separating the units herein renamed the East Pond Metamorphic Suite and the Old House Cove Group; however, he retained the Birchy Schist as a formation within the Rattling Brook Group. Bursnall (1975) informally divided the group in the area north of the Little Lobster Harbour Fault and followed Watson (1947) by excluding the Birchy Schist from it. Bursnall's (1975) informal divisions of the group are noted below, where pertinent.

Semipelitic and Psammitic Schist

DESCRIPTION

Semipelitic and psammitic schists are the most common rock type in the group. Typically, they are white to buff weathering, medium to coarse grained, gray schists that are randomly interlayered; the psammite layers range up to 50 cm thick. Locally, subordinate pelitic zones, up to 40 cm thick, are intercalated with these schists. Primary features are generally lacking though local pebbly layers indicate that the layering at a number of localities is most likely a primary feature. The pebbly psammites are most plentiful in westerly portions of the group and are best exposed along the coast in the vicinity of Bishie Cove. A strong fabric, defined primarily by muscovite, biotite or chlorite, is developed in these schists. In many places throughout the group, and particularly in the area south of Middle Arm Brook, the foliation in the semipelite and pelite has obliterated any sign of layering. In these areas, only the more competent psammite defines layering. Commonly, where layering has been destroyed, lensoid quartz veins occur parallel to the main fabric. Metaclastic schists that occur near greenschist and amphibolite are usually enriched in plagioclase, chlorite and epidote. Magnetite-bearing psammite and semipelite identical to those of the White Bay Group occur throughout the unit. At Baie Verte Brook and in the hills to the north, thin, fine grained quartzite interlayers contain up to 80% quartz, with subordinate muscovite and plagioclase.

Psammite and semipelite that outcrop on the eastern side of the Birchy Complex and which are locally infolded with the complex have been informally termed the Flat Point Formation (Kennedy, 1969, 1971; Plate 4-25). Bursnall (1975) also recognized this formation and included it in the Rattling Brook Group. The psammite alternates rhythmically with the semipelite, in layers up to 30 cm thick. Kennedy (1971) described this rock unit as follows:

> It consists of muscovite-albite psammites with thin intercalated semipelites and some massive semipelites. The psammites are pebbly in some exposures and in a few outcrops show graded bedding. They are calcareous in places and bands of feldspathic quartzite are locally intercalated with them. An impure siliceous marble is exposed at Green Point. On French Island, the semipelites contain rare rounded to subangular boulders of calcareous psammite. An impure albite-muscovite marble occurs at the junction between this formation and the Birchy Schist in Coachman's Harbour. The psammites here are also calcareous and several bands of actinolite-chlorite schist are present. At Birchy Cove and Seal Cove, the marbles, interbedded psammites and actinolite-chlorite schist, are absent and the junction with the Birchy Schist is sharp.

Bursnall (1975) noted possible soft-sediment mixing features within regularly layered psammite and pelite in the sequence at Flat Point.

Plate 4-25: *Regularly layered Flat Point formation psammitic and semipelitic schist at Flat Point, Coachman's Harbour; Rattling Brook Group. P. Kennan steadies outcrop.*

PETROGRAPHY

The platy minerals in all of these metaclastic rocks display a distinct lepidoblastic texture in thin section. Commonly, the biotite is intergrown with chlorite. Quartz and plagioclase are generally equigranular, with serrated boundaries. Plagioclase ranges from albite to orthoclase and, in many places, occurs as porphyroblasts up to 5 mm long. Small garnets (1 mm) are common and epidote, allanite, tourmaline and opaques occur as accessories.

Marble - Amphibolite/Greenschist - Graphitic Schist

The most distinctive division of the group is an assemblage of marble - amphibolite/greenschist - graphitic schist, with minor interlayered pelite. In the area west of Coachman's Harbour, Bursnall (1975) informally referred to this assemblage as the Long Pond formation. The three constituent rock types occur independently of each other in minor amounts throughout the group, but most commonly they occur together. The order of succession of these members is not regionally consistent and in many places all three are interlayered and compositionally gradational. The assemblage is best exposed at Cook In Cove, on the north side of Fleur de Lys Harbour, along the Fleur de Lys road east of Coachman's Cove, and in the area between Castor's Pond and Wild Cove Brook. It is less well developed in the areas of First Pond and southwest of Cat Path Pond. It is uncertain if these occurrences represent the same stratigraphic level or independent stratigraphic levels throughout the group. Its maximum outcrop width is approximately 0.5 km in the area of Wild Cove road and along the Fleur de Lys road; in the area of Wild Cove road, it has a strike length of at least 8 km on the west limb of a major fold (see Chapter VII).

The occurrence of buff to white carbonate rock varies considerably from pelite-rich carbonate schists to marble laminae (<3 cm) and massive marble ledges up to 10 m wide. Carbonate schists along the Wild Cove road are closely associated with chloritic pelite and are ribbonlike in appearance; the banding is defined by the alternating concentration of platy minerals and carbonate. Elsewhere, near its boundaries with interlayered greenschist and pelite, the carbonate is commonly enriched in epidote, quartz or feldspar. Marble and carbonate schist occur near graphitic schist, on both limbs of the major fold in the area of Wild Cove road. Also, some outcrops of marble are present in the core of this structure, indicating that there may be either two stratigraphic levels of marble or else structural repetition of the sequence in this area. In many places, the marble is discontinuous along strike but, due to poor exposure, it is uncertain if this reflects primary depositional patterns or later tectonic dismemberment. Locally, in the area of Coachman's Pond and east of the Wild Cove road, the marble has a coarse clastic texture; de Wit (1972) and Bursnall (1975) also reported separate occurrences of single, rounded marble clasts in massive marble.

The amphibolite-greenschist member is very variable, with greenschist predominating over amphibolite. These rocks are most abundant in the area directly east of the Wild Cove road. The amphibolite is generally massive to slightly foliated, blackish green, medium grained and feldspar porphyroblastic. It is concordant with surrounding rocks and, in most places, is compositionally gradational with them. Near Coachman's Pond, a body of epidote amphibolite more than 15 m wide is interlayered with the graphitic and calcareous schists. Along the Seal Cove road, near the Wild Cove road, vague pillowlike forms occur in massive amphibolite (Plate 4-26). Here, a rind of well foliated amphibolite defines bulbous forms of the more massive amphibolite. Amphibolite is common on the tributary to Wild Cove Brook at the north end of the Wild Cove road outcrop belt; here, it forms up to 50% of the section, with subordinate greenschist and pelitic schist.

In the amphibolites, actinolite is the common amphibole, though hornblende is present locally; the actinolite is nematoblastic, whereas the hornblende displays decussate textures. The amphibole is generally less than 5 mm long. It forms up to 60% of the rock and generally defines the main fabric. In addition, plagioclase, garnet, biotite, epidote, chlorite, rutile and opaques are present. Plagioclase (albite to oligoclase) is primarily porphyroblastic, and most commonly forms large (up to 1 cm) masses that overprint the main fabric. Garnet is less abundant, but forms subidioblastic porphyroblasts generally altered to chlorite and biotite at their margins.

The greenschists display a wide compositional range in the field, but in most cases epidote, chlorite, amphibole and

Plate 4-26: *Pillow like forms within amphibolite; note how foliated rind directly beneath hammer appears to wrap around the bulbous forms. At the Seal Cove road - Wild Cove road junction; Rattling Brook Group.*

biotite comprise more than 40% of the rock. Plagioclase and, locally, carbonate form significant components of these rocks while quartz, muscovite, chloritoid, opaques, apatite and zircon are also present. The greenschists are well foliated, medium to coarse grained, and commonly compositionally gradational with amphibolite, carbonate schists, or metaclastic rocks. Most notably, chlorite-epidote-plagioclase ± biotite ± carbonate nodules and segregations occur locally, and may form up to 25% of an individual outcrop. These prominent, knobby weathering nodules range from subrounded spheres to flattened disks up to 8 cm in diameter; they impart a fragmental aspect to the greenschists. Greenschists containing these nodules form a poorly defined belt of undeterminable width that is crossed by the Wild Cove road approximately 1.5 km from the Seal Cove road. Thin (10 cm) marble bands are interlayered with the greenschist in this belt. The greenschist is commonly rich in magnetite and ilmenite, thus accounting for a distinctively high aeromagnetic relief over these rocks in the Wild Cove road area (Geological Survey of Canada, 1968).

Notable features of the amphibolite-greenschist sequence include (i) the presence of pillowlike forms, (ii) local concentrations of nodules, and (iii) compositionally gradational contacts with surrounding rocks. These strongly suggest that this sequence was originally a volcanic-volcaniclastic sequence.

Gray, brown and black graphitic schist forms an apparently continuous marker unit up to 100 m wide within the marble-amphibolite-graphitic schist assemblage. It defines the major fold to the east of the Wild Cove road. The schist is typically sulfurous and rusty black weathering with prominent gray to black plagioclase porphyroblasts up to 5 mm long. Layering is discontinuous, and where observed rarely exceeds 15 cm. Typically, the graphitic schists are interlayered with biotite and chlorite pelite and semipelite of variable thickness. On the Wild Cove road and in the woods to the east, marble bands up to 50 cm wide are intercalated with the schists. The graphitic schist is composed of muscovite, biotite, chlorite, plagioclase, quartz and epidote, with abundant fine grained graphite. Schistosities are well defined by the platy minerals and, within feldspar porphyroblasts, by graphitic inclusion trails.

Graphitic schist and amphibolite also occur independently of carbonate rocks throughout the Rattling Brook Group. Graphitic smears and layers are common throughout the metaclastic sequence. Amphibolite within the metaclastic rocks forms mappable units, particularly west of Fleur de Lys, where it is most abundant. Its boundaries are abrupt and well defined, in contrast to the gradational margins of amphibolite in the graphitic schist - amphibolite - marble sequence. Bursnall (personal communication, 1980) indicated that there are two varieties, including plagioclase porphyroblastic amphibolite with only minor garnet, and garnet porphyroblastic amphibolite. The latter occurs either in association with meta-ultramafic rocks or independently. It also contains plagioclase porphyroblasts and is generally more schistose than the solely feldspar porphyroblastic variety. The garnetiferous amphibolite is limited to the area west of the fault which trends southwestward from Woody Cove, and the area west of the prekinematic ultramafic bodies southwest of Cat Path Pond (Figure 1-1). This does not represent a regional metamorphic isograd, as garnet is common in metaclastics to the east of this boundary. The feldspar porphyroblastic amphibolite is characterized by prominent equant plagioclase crystals (up to 5 mm), though locally the plagioclase is lathlike, imparting a subophitic appearance to the amphibolite. However, the lathlike texture is a recrystallization feature. This amphibolite, which resembles a metagabbro, most commonly occurs in association with the prekinematic ultramafic rocks in the group.

Garnetiferous Semipelite

Garnetiferous quartz-muscovite semipelite forms a poorly defined belt through the center of the Rattling Brook Group. It extends from the area south of Cat Path Pond southwestward along flat-topped hills to the Little Lobster Harbour Fault; south of the fault, the belt is discontinuous from Lower Duck Island Cove Brook to the area of Castor Pond. The semipelite also occurs sporadically throughout the group. This member was informally termed the Drover's formation by Bursnall (1975). It is characteristically a coarse grained quartz-muscovite semipelite containing porphyroblasts of garnet that locally exceed 1.5 cm in diameter. Where present, large garnets are generally sparsely distributed throughout the schist and compose a maximum of 15% of the rock, whereas smaller garnets tend to form clusters that form up to 60% of the rock (Bursnall, 1975). The main fabric in the semipelites forms augen of the porphyroblasts and, in places, an internal fabric is mesoscopically visible within the larger garnets. The semipelite is interlayered with thin psammitic and quartzitic bands that most likely represent bedding; otherwise, primary features are lacking. A few small (< 1 m wide) pods of epidote amphibolite are also associated with the semipelite.

Minor biotite occurs in the schist, in many places intergrown with chlorite. In addition, epidote, apatite, tourmaline and magnetite are common accessories.

Contact Relationships

The stratigraphy of the Rattling Brook Group is uncertain due to poor inland exposure and a high degree of tectonic disruption. Local stratigraphic correlations in the area east of the Fleur de Lys road were outlined by Bursnall (1975). Externally, the group is conformably interlayered with the Old House Cove Group and the Birchy Complex, and is assumed to be faulted against the East Pond Metamorphic Suite.

Bursnall (1979; personal communication, 1980) reported a transitional boundary between the Old House Cove and Rattling Brook Groups in the area west of Fleur de Lys Harbour. The boundary there is imprecise, due to discontinuous or poor exposure of marker units. The contact zone is marked by a gradational decrease of pelitic material concomitant with an increase in layer thickness from the Rattling Brook Group westward into the Old House Cove Group. Bursnall (personal communication, 1980) suspected that recumbent to moderately inclined folds affect this boundary; however these have not been precisely defined, again due to lack of marker horizons. Thus, there may be some infolding of Old House Cove and Rattling Brook Groups along the contact. The lack of reliable facing evidence and the suggestion of complex structures at this contact makes interpretation of the original stratigraphic order of these units impossible. The contact between the Rattling Brook Group and Birchy complex is discussed following the description of the complex.

The contact between the Rattling Brook Group and East Pond Metamorphic Suite is unexposed, but change in rocks types is abrupt. Rocks in the area of the contact are generally fine grained and intensely foliated; hence, this contact is assumed to be tectonic. Kidd (1974) informally termed this unexposed contact the Noseeum boundary slide (see Western Orthotectonic Block, Chapter VII).

Depositional Environment

Primary features are conspicuously lacking in rocks of the Rattling Brook Group; hence, its environment of deposition is uncertain. The rock associations within the group are remarkably similar to those of the White Bay Group; the latter is interpreted as a continental slope deposit at the edge of a carbonate shelf. A similar environment of deposition is here suggested for the protoliths of the Rattling Brook Group.

Age and Correlation

The age of the Rattling Brook Group is uncertain. The age of the group relative to the flanking Old House Cove Group and Birchy Complex is uncertain due to the paucity of reliable younging evidence near the contacts of these units. The best evidence for the age of the group comes from regional correlation. The Rattling Brook Group displays remarkable similarities to the White Bay Group, as first noted by de Wit (1972). Specifically, the graphitic schist - amphibolite - marble sequence of the Rattling Brook Group is very similar to the sequence proximal to the Garden Cove Formation of the White Bay Group. Significantly, the Rattling Brook Group amphibolite and that of the Garden Cove Formation are petrochemically correlative (see Chapter VI). Also, the garnetiferous semipelite of the Rattling Brook Group is lithically identical to that of the Pigeon Island Formation in Western Arm. Unfortunately, the succession of units within the Rattling Brook Group is not as obvious as that of the White Bay Group; hence, only lithic correlations can be made at this time. There are major differences between the two groups, including: (i) the Rattling Brook Group appears to lack a thick basal amphibolite corresponding to the Oody Mountain Formation of the White Bay Group, (ii) prekinematic ultramafic rocks occur only in the Rattling Brook Group, and (iii) the Rattling Brook Group is interlayered to the east with a thick mafic succession, the Birchy Complex, whereas no analogue to the Birchy Complex has been found in contact with the White Bay Group. These contrasts may reflect differences in the substrate upon which these units were deposited; this idea is expanded in a later section of this report (also see White Bay Group and Birchy Complex).

Nevertheless, the overall similarities of the two groups are striking enough that the two groups are here considered to be in large part correlative (see also de Wit, 1972). This lithic correlation of the groups suggests that lower portions of the Rattling Brook Group, like those of the White Bay Group, are laterally equivalent to the Old House Cove Group. This may explain the apparently gradational contacts between these groups in the area west of Fleur de Lys. Also, the correlation of the Rattling Brook and White Bay Groups suggests that the Rattling Brook Group is Eocambrian to Lower Ordovician. The age correlation may not be reliable, however, as the two groups could represent either a diachronous sequence or chronologically distinct sequences deposited in similar environments.

BIRCHY COMPLEX

Definition and Extent

The Birchy Complex consists of greenschist (with lesser amphibolite), metagabbro, and graphitic schist that outcrop as a structural assemblage along the eastern margin of the Fleur de Lys Belt on the Baie Verte Peninsula. The complex is distributed in two outcrop belts separated by the Little Lobster Harbour - Advocate Fault system (Figure 1-1). The southerly belt is generally less than 1 km wide. South of Flat Water Pond it bifurcates, with the thicker westerly part separated from the easterly part by metaclastics typical of the Rattling Brook Group. The southerly belt is composed mainly of monotonous greenschist. The northerly belt has a maximum outcrop width of more than 2 km in the area west of Marble Cove. The belt is also bisected by the Rattling Brook Group in the area north of Coachman's Harbour.

The northerly belt contains a greater variety of rocks than the southerly belt, and is ophiolitic in character. Ultramafic rocks in the complex are indistinguishable from other ultramafics in the Fleur de Lys Supergroup and are therefore described later in this chapter. The other rock types are distinct from other portions of the Fleur de Lys Supergroup and are described below. The stratigraphy of the complex is very tenuous because it is a structural assemblage. Bursnall, however, demonstrated a stratigraphic linkage of rock types in the complex; his thesis (Bursnall, 1975) gives a detailed stratigraphic assessment of these rocks between Marble Cove and Deep Cove.

The section of the complex exposed along Slaughter House Cove Brook is herein designated as a reference section, because most typical rock types of the unit are exposed there. The rock types are described below in three groupings, including calc-silicate schist (South Cove schist) and metagabbro, graphitic schist and mélange (Slaughter House Schist), and greenschist and subordinate metaclastics.

These rocks were first referred to as the Birchy Schist by Fuller (1941) in the area of Green Point. The term was later adopted by other workers in the area, although its stratigraphic rank and association have not been agreed upon. Watson (1947) considered the schists as part of the Baie Verte Group and separated them from his Rattling Brook Group. Neale (1959a) also assigned these mafic rocks to the Baie Verte Group, but Neale and Kennedy (1967) included some of them, to the northwest of Coachman's Harbour, in the Fleur de Lys Group.

Church (1969), de Wit (1972) and Kidd (1974) subsequently included the Birchy Schist Formation in the Rattling Brook Group. Kennedy (1969, 1971) considered these rocks to be a formation within his Eastern Sequence of the Fleur de Lys Group. Later, he included parts of the Pelée Point schist (of this report) in the formation (Kennedy, 1975a). Bursnall (1975), Williams et al. (1977) and Williams (1977a) agreed with Watson (1947) in separating these rocks from the Rattling Brook Group, but these workers all included the Birchy Schist in the Fleur de Lys Supergroup. Bursnall (1975) mapped the unit north of Baie Verte and termed it the Birchy Group. Since the greenschist contains ophiolitic remnants and zones of early tectonic disruption that predate the regional polydeformation of the area, Williams et al. (1977) and Williams (1977a) preferred to call it the Birchy Complex. Likewise, the term Birchy Complex is applied to these rocks in this report.

Detailed mapping of the Birchy Complex has been carried out by many workers (Kennedy, 1969, 1971; de Wit, 1972; Kidd, 1974; Bursnall, 1975), though only locally can the unit be consistently divided (Kennedy, 1969, 1971; Bursnall, 1975). Rocks included in the Birchy Complex in this report were divided into three formations by Kennedy (1969, 1971) and four formations with associated igneous rocks by Bursnall (1975). At the present scale of mapping, the following two of these divisions were separable from the rest of the complex: (i) graphitic schist informally referred to as the Slaughter House Schist by Kennedy (1969, 1971) and included in Bursnall's (1975) Beaver I formation, and (ii) metagabbro with associated calc-silicate schist referred to by both these workers as the South Cove Schist. Much of the following description of the complex is taken from the detailed work of Kennedy (1969, 1971) and Bursnall (1975).

Metagabbro and Calc-silicate Schist

This unit includes the belt of metagabbro and calc-silicate schist that outcrops near and along the coast of Baie Verte from Seal Cove north to Coachman's Cove (Figure 1-1). The metagabbro also occurs inland as a fault-bounded block to the west of Slaughter House Cove Point. The calc-silicate schist, informally termed the South Cove schist (Kennedy, 1971), outcrops along the coast from the south side of Slaughter House Cove northward to Coachman's Harbour. Both the metagabbro and the South Cove schist are tectonically bounded (Bursnall, 1975).

Bursnall (1975) described the metagabbro as follows:

> ...(In the metagabbro) textural variation is great in detail but in general terms the field appearance is of a medium- to coarse-grained rock with off-white to cream coloured base in which are set dark green amphibole felted aggregates or single crystals. The high contrast between these two constituent fractions emphasizes the presence of any tectonic fabric or predating inferred igneous foliation. These are normally seen as very thin (<5 mm), locally laterally persistent, leucocratic laminae that appear to truncate inferred pseudomorphs of original igneous grains but are affected by the early structures. The rock is prominently schistose... in thin section, the leucocratic aggregates are composed of (variably) clinozoisite, zoisite, epidote, albite and quartz with subordinate actinolitic amphibole.
>
> Melanocratic areas, which are sometimes densely distributed, consist of medium brownish-green to light-green hornblende and accessory actinolite-tremolite. Sphene is a ubiquitous accessory. Towards the centre of the body, less schistose varieties sometimes occur and a typical garbenschieffer texture obtains. Here, the melanocratic fraction is sometimes composed of coarse-grained single amphibole crystals (<1.5 cm), rather than a felted aggregate, set in a leucocratic matrix of zoisite and quartz. Textures in the area may be relic of original igneous grain distribution.
>
> Very rare narrow zones of amphibolite, composed of at least 80% hornblende with accessory albite, sphene, epidote and opaques, are undoubtedly primary in origin and may indicate relict igneous banding. In Slaughter House Cove, a tectonic lens of melanocratic metagabbro and associated hornblendite are spatially close to a small serpentinite body.

In contrast to the metagabbro, the protolith to the South Cove schist is more ambiguous, as described by Bursnall (1975) below:

> An unusual, regularly banded, amphibolitic schist outcrops for some 4 km along strike south of Coachman's Harbour.
>
> The formation is rarely more than 300 m thick and is composed of laterally persistent melanocratic layers of amphibole-rich rock interbanded with leucocratic amphibole-poor layers... Some layer boundaries appear gradational while others are sharp. It is apparent at some localities that two contrasting adjacent bands comprise a single unit with sharp upper and lower boundaries but gradational internal contact. This association may be rhythmically repeated through thicknesses of 10 m or more. Individual layers vary in thickness from 1-2 cm to as much as 40 cm and there may be considerable grain size variation throughout a unit. Thin, layer parallel epidote laminae are common, particularly on the leucocratic external margin of units.
>
> In thin section, the melanocratic fraction may be true amphibolites with as much as 85% amphibole (normally actinolite but, rarely, a brownish hornblende). This contrasts with the dominantly epidote (and/or clinozoisite)+albite+quartz+actinolite (+ tremolite) assemblage of adjacent layers; gradational contacts show intermediate distributions between these extremes. Sphene and calcite are almost ubiquitous accessories... Muscovite, apatite, and magnetite are uncommon.
>
> The origin of these rocks is enigmatic. Watson (1947) suggested that they were 'metadiorite'. Their field appearance is remarkably similar in places...to layered mafic gabbros....this superficial similarity is emphasised by numerous thin dikes oblique to the banding, 'pegmatitic' fractions, epidote-quartz veins and small homogeneous metagabbro bodies present in the Slaughter House Cove section... Exceptionally small areas of gabbroic textured epidote-amphibolite are included within the layering....it is difficult to definitively state whether these are fragmentary or small intrusive bodies....

In contrast to Watson (1947), Kennedy (1969, 1971) interpreted the schist to be sedimentary in origin. Bursnall (1975) was noncommittal and considered both ideas, whereas Williams (1977a) supported an igneous origin. Indeed, the characteristics of the schist, described above by Bursnall, and its occurrence along strike with definite metagabbro of the complex, support an igneous origin for the schist. The protolith is herein interpreted to be gabbroic.

Minor intrusive phases other than the metagabbro and possibly the South Cove schist were noted by Bursnall (1975). Along the coast from Seal Cove to Coachman's Harbour, he observed two sets of metamafic dikes, one pre- to synkinematic, the other interkinematic. The dikes crosscut banding in the complex and petrographically resemble the metagabbro described above, though the dikes are more mafic. Bursnall (1975) also reported a sill-like leucocratic intrusive on the north shore of Seal Cove. It is trondhjemitic in composition and probably interkinematic.

Graphitic Schist and Mélange

Graphitic schist, usually associated with chloritic semipelite, quartzite or calcareous pelite, forms mappable marker units within the greenschist terrane. The thickest of these units, generally less than 100 m wide, occurs immediately west and southwest of Slaughter House Cove. The schist is brownish black, strongly foliated, and characterized by centimetre scale alternations of quartz-rich and quartz-poor layers. In addition, thin graphitic interlayers and smears are common throughout the complex.

Significantly, the graphitic schist is commonly associated with mélange zones in the complex. Bursnall (1975) first described the close association of a graphitic schist zone with a mélange zone on the lower portion of Slaughter House Cove Brook as follows:

> ...well-exposed sections (of the mélange) occur in Slaughter House Cove Brook 300 m west of Slaughter House Cove and in the southern branch of this brook 1 km to the west. It is invariably closely associated with thin brownish black striped graphitic pelites (equivalent to those mentioned above) which to the south are somewhat calcareous. Actinolite schist pods, spessartine garnet aggregations, small disrupted blocks of metagabbro, psammitic schists, metabasite lenses, and small serpentinous masses all occur as angular to rounded fragments (up to 10 × 3 m) in a gray-green schist matrix... The matrix varies from quartz-rich to quartz-poor with much chlorite: graphitic horizons are only occasionally present... On the present interpretation the graphitic pelitic schists occur high in the Birchy Schist Group succession in this area. The locally associated mélange is considered to be dominantly tectonic in origin...

Subsequent to this initial recognition, many more mélange zones were mapped in the complex (Williams et al., 1977; Williams, 1977a) (see Figure 7-2). The best exposures are in Coachman's Harbour, where the zones have informally been termed the Coachman's mélange (Williams, 1977a). The zones, ranging up to 50 m wide, are very similar in character to the Slaughter House Cove Brook occurrence, though black graphitic pelite appears to be more common in the matrix at Coachman's Harbour. All of these occurrences are characterized by the inclusion of serpentinized ultramafic rocks or coarse actinolitic schist of ultramafic derivation (Kennedy, 1971; Bursnall, 1975). In addition, the mélanges display all of the regional minor structures evident in other rock types of the complex; hence, they must have formed either before or very early in the Fleur de Lys deformational history. These graphitic mélange zones appear to be confined to the northerly outcrop belt of the complex but, in roadside exposures west of Slink Pond, bright green actinolite-fuchsite schist pods (up to 2 by 1 m) occur in chloritic and graphitic pelites of the southerly outcrop belt (see Roadside Slide, Chapter VII).

Greenschist and Metaclastic Rocks

The Birchy Complex is composed mainly of strongly foliated chlorite-epidote-actinolite schist with subordinate amphibolite and local interlayers of psammitic, pelitic and graphitic schist. It is usually banded on a centimetre-millimetre scale, with individual layers defined largely by variations in the amounts of chlorite, epidote, feldspar and quartz. The greenschists range in color from pale to dark green. Primary features are not common in the unit, though, in places, the greenschist contains mafic clasts ranging up to 30 cm across and resembles an agglomerate (Plate 4-27). Bursnall (1975) and Kennedy (1969, 1971) reported such fragmental members from the southeastern part of Coachman's Harbour, and Kidd (1974) reported similar horizons to the northwest of both Kidney Pond and Flat Water Pond.

Plate 4-27: *Birchy Complex greenschist containing subangular epidote-albite clast, Coachman's Harbour.*

Also in the greenschist, local epidote-chlorite-plagioclase pods and lenses up to 8 cm across and small (6 cm) mafic pods that display a diabasic texture may represent primary clasts. Bursnall (1975) traced a "clastic" greenschist horizon, generally less than 5 m thick, from the area of Slaughter House Cove Brook southwestward toward the Fleur de Lys road. He indicated that it maintains a consistent position relative to the nearby Rattling Brook Group, and described the unit as follows:

> ...cream to buff colored angular epidote/quartz 'clots' are set in a mesocratic laminated greenschist matrix... The clasts may account for as much as 60% of the rock. In one loose block, they appeared to be graded. The particles are generally less than 2 cm in greatest dimension and their angular nature suggests a fragmentary origin. The lithology may have originated as a coarse-grained tuff.

A fine grained chlorite-epidote-albite ± actinolite assemblage prevails in the greenschist; igneous textures are absent. Chlorite and actinolite generally define the well developed fabric in these rocks, with albite and quartz forming a fine to medium grained granoblastic mosaic. Locally, hornblende is present in place of actinolite. Yellow epidote and/or clinozoisite is ubiquitous. Muscovite, biotite, sphene, rutile, magnetite, calcite and apatite are also present as accessories.

Massive, fine grained, dark green amphibolite is interlayered locally within the greenschist; in places, it is plagioclase porphyroblastic. Amphibolite layers are generally less than 4 m thick and, in most places, concordant with the surrounding greenschists. Bursnall (1975) and Kidd (1974) both noted two locations where the amphibolite forms dikes that crosscut banding in the greenschist. Bursnall (1975) reported pillowed forms from a massive amphibolite, approximately 2 km southwest of Slaughter House Cove. He also summarized the petrography of the amphibolites:

> In thin section, these compact mafic rocks are notably quartz-poor, and consist mainly of fine-grained nematoblastic amphiboles (ferro-actinolite + tremolite/actinolite) with abundant epidote and occasional small (<2 mm) lensoid aggregates of albite and quartz; sphene and chlorite are common accessories, opaques are rare.

Metaclastic schists form conspicuous members within the monotonous greenschist terrane. They include quartz pebble zones, coticules and psammitic and semipelitic schists. Quartz pebble layers occur within the greenschist in roadside outcrops 5 km north of Flat Water Pond. Locally, distinctive coticule (fine grained garnetiferous quartzite) occurs within the greenschist (Plate 4-28). De Wit (1972) and Kidd (1974) both reported this rock type from the southerly outcrop belt of the complex, and Bursnall (1975) noted several occurrences between Slaughter House Cove and Coachman's Harbour. The coticules are generally less than 10 cm thick and are pinkish red to purplish or black in color; typically, they are strongly laminated. Bursnall (1975) noted that they are commonly in close association with graphitic schists. These rocks most likely represent metacherts. One ellipsoidal quartz-garnet pod, 40 cm long, of uncertain origin was noted in greenschists exposed along the transmission line west of the Baie Verte highway.

Bursnall (1975) reported that the coticules are composed of quartz, garnet and magnetite, with minor stilpnomelane, chlorite, actinolite and apatite; he noted the presence of

Plate 4-28: *Distinct coticule (quartz-garnet) band (left of hammer) infolded with greenschist at Coachman's Harbour; Birchy Complex.*

riebeckite in one thin section. He also reported the following garnet composition determined by X-ray fluorescence: pyrope (6%), almandine (35%), spessartine (46%), and grossular (15%).

Thin zones of psammitic to pelitic schists are commonly interlayered with the greenschists. Locally, the zones range up to 100 m thick, such as the one that extends from Tom's Pond to the area northeast of Jack's Pond (Kidd, 1974). Typically, these metaclastics are thinly bedded (<30 cm) but, otherwise, primary structures are rare. In the northerly outcrop belt, Bursnall (1975) mapped calcareous pelites within the Birchy Complex greenschist near its contact with the Rattling Brook Group and in the area one kilometre southwest of Slaughter House Cove.

Contact Relationships

The Birchy Complex is in comformable, gradational contact with the Rattling Brook Group (Kennedy, 1969, 1971; de Wit, 1972; Kidd, 1974; Bursnall, 1975), though locally the contact is faulted. It is in tectonic contact along most of its eastern boundary with rocks of the Baie Verte Belt (see Chapter VII). Although the contact between the Rattling Brook Group and Birchy Complex is conformable, the relative stratigraphic order is uncertain. Kennedy (1969, 1971, 1975a), Kidd (1974) and Bursnall (1975) maintained that the Rattling Brook Group, in part if not totally, overlies the Birchy Complex. Their interpretation is based on younging directions in the Flat Point Formation near its contact with the Birchy Complex at Flat Point. In contrast Church (1969) and de Wit (1972) interpreted the Birchy Schist to lie at the top of the Rattling Brook Group, based mainly upon regional structure and stratigraphy. The Flat Point contact was revisited during the present study in the company of J. Bursnall and M.J. Kennedy; however, both visits failed to confirm the previously stated relationships.

It is doubtful if enough conclusive evidence is preserved to establish definitely the stratigraphic position of the complex. Even if the Flat Point psammites are conclusively shown to overlie the Birchy Complex, the stratigraphic relationship of the Flat Point rocks to the remainder of the Rattling Brook Group is uncertain; the Flat Point Formation could be a tongue of the group within the Birchy greenschists.

The disagreement between previous workers on the stratigraphic order of the group and the complex may be because they interfinger. However, regional interpretation and correlation suggest that at least a portion of the Birchy Complex greenschists overlie the Rattling Brook Group and that the portion probably represents the top of the Fleur de Lys Supergroup. Mélange zones within the Birchy Complex have been interpreted as marking the tectonic transport of ultramafic rocks through the complex prior to its polyphase deformation (Bursnall, 1975) (see Chapter VII). Williams (1977a) correlated them with ophiolitic mélange at the base of allochthonous ophiolites in western Newfoundland, and he related the Birchy mélanges to the transport of these allochthons. Since the loading of these ophiolites onto the Fleur de Lys terrane and their movement across it probably terminated Fleur de Lys deposition and initiated deformation of the supergroup (Williams et al., 1977), the ophiolitic mélange of the Birchy Complex may represent the upper surface of the Fleur de Lys depositional pile at the time of obduction. Significantly, greenschists identical to those of the Birchy Complex occur as a basal dynamothermal aureole to the allochthonous ophiolites of western Newfoundland (Church and Stevens, 1971). The aureoles are named the Goose Cove Schist in the Hare Bay Allochthon (Smyth, 1973; Williams and Smyth, 1973), the Old Man Cove Formation in the Humber Arm Allochthon (Williams, 1975, 1977a), and the Murray's Cove Schist beneath the transported Coney Head complex in western White Bay (Williams, 1977b).

Birchy Protoliths

Bursnall (1975) summarized the probable primary characteristics of the complex based on descriptions given above. It should be emphasized that the stratigraphy he referred to is tenuous, and was not apparent from the regional mapping carried out during the present work:

> The inferred primary lithologies within the Birchy Schist Group included occasionally pillowed mafic volcanic lava, tuffaceous and agglomeratic sediments and thick pelitic volcaniclastics with increasing dominance of thin pelitic and semipelitic units upwards. Mafic intrusives, including probable basaltic dikes and gabbroic sheets, occur towards the base of the group. The succession was cut by tectonically emplaced ultramafic (serpentinite) bodies and a large gabbro complex—the Slaughter House metagabbro. This lithological association and the inferred lithostratigraphy suggest a submarine volcanic pile with decreasing eruptive dominance upwards; the present rock association may be described in classical terms as 'ophiolitic'.

Thus, a major part of the complex, including some of the greenschists, the metagabbro and calc-silicate schists, and tectonically emplaced ultramafic rocks (see Meta-ultramafics below) represent the dismembered components of an ophiolite suite. Some of the greenschists in the complex appear to be conformably interlayered with metaclastic rocks of the Rattling Brook Group; this infers that at least a part of that group was deposited on oceanic basement. The ophiolitic mélanges are of uncertain origin; they may be either tectonic, olistostromal or a combination of these. They do, however, mark significant structural zones that have been related to the structural dismemberment of the ophiolitic rocks of the complex (see above and Chapter VII).

Age and Correlation

There is no direct evidence for the age of the Birchy Complex. Based on lithic correlations of greenschists and mélanges of the Birchy Complex with aureole rocks and mélanges at the base of allochthons in western Newfoundland (Williams, 1977a) as well as regional tectonic constraints (Williams, 1975) (see Chapter VII), the Birchy Complex can be no younger than late Early Ordovician. In addition, its relationships with other Fleur de Lys rocks indicate that it is younger than Eocambrian. Closer estimates of the age of the complex can be made from regional correlation.

Lithically and geochemically, the complex is similar to nearby ophiolitic rocks of the Baie Verte Belt, though it is more highly deformed (see Chapters V and VI). The lithic similarity was also noted by Watson (1947) and Neale (1959a), both of whom placed rocks of the complex in the archaic Baie Verte Group. Bursnall (1975) demonstrated the lithic correlation of parts of the complex with ophiolitic components of the Advocate Complex. Because of these similarities and correlations, most of the Birchy Complex is herein considered equivalent to the nearby ophiolites of the Baie Verte Belt, which are considered to be Late Cambrian to Early Ordovician in age.

Other lithic correlations have been made between the complex and rocks of the Fleur de Lys Belt. Kennedy et al. (1973) and Williams (1977a) correlated it with greenschist within the Fleur de Lys Supergroup on the Grey Islands. Similar greenschists also occur at the structural top of the Fleur de Lys sequence on Glover Island (Knapp, 1980). Rock types and relationships within the unit also closely resemble those of the Pelée Point schist of the Ming's Bight Group (see next unit), a correlation that was first suggested by Kennedy (1973, 1975a). Based on lithologic similarities, de Wit (1972) considered the Birchy Schist to be equivalent to amphibolite in the Rattling Brook Group along Wild Cove road and to the Garden Cove Formation of the White Bay Group. The geochemistry of these rocks (see Chapter VI) does not support de Wit's correlations; however, lateral equivalency of parts of the complex and the group is still possible. In addition to the possible interfingering relationship between the units (see above), the Birchy mélanges contain rock assemblages similar to Bursnall's (1975) Long Pond formation (marble - amphibolite - graphitic schist); thus, the mélanges may in part be disrupted equivalents of the Rattling Brook Group marble - amphibolite - graphitic schist assemblage.

MING'S BIGHT GROUP

Definition and Extent

The Ming's Bight Group comprises the dominantly metaclastic schists with subordinate greenschist and amphibolite that outcrop on the stubby peninsula between Ming's Bight and Pacquet Harbour. The greenschist and amphibolite, informally termed here the Pelée Point schist, form mappable divisions interlayered with the psammite. Both the metaclastics and the Pelée Point schist are well exposed in Ming's Bight. The thickness of the group is uncertain due to structural complexities; however, the maximum outcrop width is about 5 km in the area south of Cape Hat. Likewise, the top and bottom of the unit are unidentified because of complex structures in the group. The unit generally trends east, but dips are highly variable due to intense late folding.

The usage of the term Ming's Bight Group here closely coincides with the original definition of the Ming's Bight Formation by Watson (1947) and the Ming's Bight Group by Baird (1951). However, the definition of the group used here includes amphibolite in a fault-bounded block at Pelée Point. These rocks were traditionally assigned to the Pacquet Harbour Group, but are now considered an integral part of the Ming's Bight Group since they appear to be tectonically severed from the Pacquet Harbour Group, yet conformable with the Ming's Bight Group (see Contact Relationships, below).

The Ming's Bight Group was also mapped and described by Neale (1958a,b), Church (1969), Kennedy et al. (1972), Kennedy (1975a,b) and DeGrace et al. (1976). Kennedy et al. (1972) and Kennedy (1975a,b) included rocks of the Flat Point Formation and other easterly portions of the Rattling Brook Group in the Ming's Bight Group. This extension of the group has not been recognized by other workers, nor is it adhered to here.

Metaclastic Rocks

The metaclastic rocks form a monotonous sequence lacking persistent marker horizons. The medium to coarse grained psammite and semipelite are generally buff to gray and brown weathering and are layered over hundreds of metres; layers range up to 1 m thick (Plate 4-29). In many of the coarser grained schists, preserved pebbles define layering, indicating that it probably represents bedding. Church (1969) and DeGrace et al. (1976) reported clasts of white quartz, marble, calc-silicates, psammite, blue quartz, and possible mafic metavolcanics from these pebble metaconglomerate layers.

Plate 4-29: *Typical medium grained, thin to medium layered psammitic schist of Ming's Bight Group, Bois Island. (Field of view approximately 4 m across.)*

Semipelitic units are particularly abundant in the area between Cape Corbin and Grand Cove. These metaclastic rocks are typically composed of quartz, albite, biotite, muscovite and chlorite; near the contacts with the Pelée Point schist, chlorite concentration increases and amphibole and epidote appear in these metasediments. Baird (1951) summarized the petrography of typical psammites in the group as follows:

In thin section, the rocks are seen to be medium grained and foliated, and to have a granoblastic fabric. Quartz, which comprises from 50 to 80 percent of the rock, occurs as aggregates of interlocking serrate-edged grains that average between 0.1 and 0.5 mm in diameter in different fabrics. Biotite occurs in elongated wisps and shreds averaging in fine-grained varieties, 0.5 by 0.05 mm, and in coarse-grained facies as large as 2.5 by 1 mm. It constitutes up to 15 percent of the rock. Untwinned grains of albite (An_{6-10}) averaging about 0.5 mm in diameter occur in interlocking aggregates that comprise from 5 to 15 percent of the rock. Chlorite is a common secondary mineral, and occurs as patches of interlocking sheaves and as an alteration product along cleavage lines and on the edges of biotite crystals. Other accessory minerals include muscovite, magnetite, hematite, ilmenite, leucoxene, zircon, calcite, pyrite, apatite, zoisite, and, rarely, tourmaline.

Locally, as in Ming's Bight, satiny fine grained gray pelites are interlayered with the other metaclastics. Interlayers of garnetiferous quartz-muscovite semipelite are common on the north side of Pacquet Harbour, in Hardy Harbour, and sporadically throughout the group. Graphitic schist occurs as smears and thin zones (up to 10 m) in only a few places in the group; most notably, it is interlayered with the metaclastic rocks immediately south of Ming's Tickle and on South Brook, approximately 1.5 km south of Ming's Bight. The schists are generally fine grained, blackish brown and composed of biotite, muscovite, plagioclase and quartz. At South Brook, the graphitic schist contains large (1 by 2 m) pods of talc-carbonate and actinolite-carbonate schist as well as smaller pods of a metagabbro-like rock. The pods appear to contain all structural features of the surrounding rocks, and may mark a predeformational mélange zone similar to those of the Birchy Complex.

Larger actinolite schist pods (up to 4 by 10 m) occur within dominantly fine grained psammitic rocks on the north side of Pelée Point and near the mouth of the brook that empties into the Northwest Arm of Pacquet Harbour (Plate 4-30). The mafic pods on the north side of Pelée Point are dominantly feldspar-bearing actinolite schist pods. Those near the brook are composed dominantly of pale green actinolite and plagioclase, and resemble metagabbro. Two of the larger blocks contain layers up to 25 cm thick of pure actinolite schist along one margin. Smaller pods (up to 1.5 m long) of coarse actinolite-biotite schist also occur in the metaclastic rocks here. Based on their layered appearance and high content of nickel and chromium (sample 1340083, Appendix III), the rafts probably represent metacumulate rocks.

This zone of actinolitic pods appears to be unique in that it is coincidental with the occurrence of coarse kyanite and staurolite in the metaclastic rocks of the group. Approximately 0.5 km up the brook, along strike, a thick section of pelitic rocks contains an infolded layer approximately 30 cm thick of kyanite-bearing quartzite; the surrounding pelite is characterized by large blades of staurolite and kyanite (up to 4 cm long). Bright blue kyanite blades up to 10 cm long occur in quartz segregations within the pelite. Further up the brook, along strike, thin zones (up to 3 m) of flaggy quartzite as well as pelitic zones containing amphibolite pods (1 by 2 m) characterize the sequence. The flaggy quartzitic rocks are highly strained and reminiscent of slide zone mylonites.

Pelée Point Schist

Thick zones of greenschist and amphibolite are interlayered with psammite in four places within the group: inland southeast of Ming's Bight, at two localities on the east side of Ming's Bight, and at Pelée Point. The zones are informally termed the Pelée Point schist in this report. All of the units appear to be conformable with surrounding metaclastic rocks. The thickest unit on the east side of Ming's Bight has an outcrop width of approximately 500 m. In the Ming's Bight area, the schist is composed of foliated, fine to medium grained actinolite-chlorite-albite schists, satiny chlorite pelite, and gritty chloritic metasediments. Williams (personal communication, 1977) noted discrete pods of actinolite schist up to 20 cm long in a narrow zone of greenschist on the coast, immediately east of South Brook.

Fragmental rocks occur in the greenschists immediately to the northeast. Clasts here are composed of cream colored calc-silicates, which may represent felsic volcanic bombs, and amygdular actinolite-chlorite clasts. All the fragments are rounded, elongate and less than 20 cm long. DeGrace et al. (1976) reported amygdules from the northerly zone of greenschists in Ming's Bight.

At Pelée Point, the metamorphic grade is higher than at Ming's Bight and the schist is mainly fine grained, blackish green, hornblende-biotite-plagioclase rock. Generally, it is massive, though locally it displays a metagabbro-like texture and, in places, is feldspar porphyroblastic. Hornblende and biotite define the fabric in these schists and, in many places, random hornblende porphyroblasts (up to 2 cm) overprint the fabric. Locally, the amphibolite contains thin layers and clumped balls of fine grained psammitic schist, near the contact with metaclastic rocks of the group at Pelée Point.

Contact Relationships

The Ming's Bight Group is in tectonic contact with the Point Rousse Complex to the west. The contact is marked by a high angle fault in the area of South Brook that is intercepted to the south by an easterly directed thrust fault which thence marks the contact (Figure 1-1). The group is intruded by the Dunamagon Granite (DeGrace et al., 1976). The contact between the Ming's Bight and Pacquet Harbour Groups is controversial. Most previous workers considered the isolated Pelée Point schist at Pelée Point as part of the Pacquet Harbour Group and the contact between this schist and the metaclastic rocks of the Ming's Bight Group to be the contact between the Ming's Bight and Pacquet Harbour Groups. Baird (1951) considered this to be an unconformity but later workers (e.g. Neale, 1958b; Church, 1969; Kennedy, 1975a; DeGrace et al., 1976; Williams et al., 1977) considered the contact to be conformable.

In the present study, I have approached the relationship differently. Since the zones of actinolite schist pods along the southern boundary of the Ming's Bight Group strongly resemble those of the Birchy Complex, I interpret them to represent relict premetamorphic faults, analogous to those in the complex, along which ophiolitic rocks were transported over the Ming's Bight Group (see Chapters VII and IX). The actinolite schist pods most likely represent remnants of the transported ophiolite. As these distinctive blocks occur only near the contact of the Ming's Bight and Pacquet Harbour Groups, and because the basement to the Pacquet Harbour Group is most likely ophiolitic, I suggest that the Ming's Bight and Pacquet Harbour Groups were juxtaposed along the premetamorphic faults, which have largely been obliterated by

Plate 4-30: *Large actinolite-feldspar pod (to right of hammer) within thinly layered fine grained psammite of Ming's Bight Group at Northwest Arm, Pacquet Harbour.*

subsequent regional metamorphism and intrusion of the Dunamagon Granite. The Pelée Point schist at Pelée Point is consequently viewed as a conformable member of the Ming's Bight Group. It may well be significant that the trends of layering in the Pelée Point schist and Ming's Bight psammite are parallel to each other, yet oblique to that of the Pacquet Harbour Group on the south side of Pacquet Harbour. The contact between the two groups in the inland area south of Ming's Bight is most likely tectonic, as the groups also display divergent structural trends in this area.

Depositional Environment

The environment of deposition of the Ming's Bight Group is conjectural, though the Pelée Point schist indicates a pronounced local volcanic influence. The overall aspect of the metaclastics is most reminiscent of flysch deposits; thus, the group may represent a basinal turbidite sequence. The inclusion of probable ophiolitic metacumulates in the sequence is most likely related to premetamorphic faulting in the group and will be discussed with the regional structure.

Age and Correlation

The age of the group is uncertain. The Ming's Bight Group has been correlated with the Flat Point formation and other westerly metaclastics of the Rattling Brook Group. Most of the Pelée Point schist, with the exception of rocks at Pelée Point that were formerly considered part of the Pacquet Harbour Group, has been considered equivalent to the Birchy greenschists (Neale and Kennedy, 1967; Kennedy, 1971, 1973, 1975a,b; Kennedy et al., 1972). The occurrence of bright green, coarse grained actinolite schist and polydeformed mélange within the group provides a strong link between it and the Birchy Complex; it is only in these two units on the Baie Verte Peninsula that these distinctive rocks are found. The group also resembles portions of the Horse Islands Group, as discussed in the next section. These correlations suggest that the group ranges from Eocambrian to Early Ordovician in age, similar to the possible ranges of the Birchy Complex and the Rattling Brook Group.

HORSE ISLANDS GROUP

Definition and Extent

The name Horse Islands Group is herein proposed for the assemblage of metasedimentary rocks and greenschists confined to the Horse Islands. There are no previous descriptions of these rocks, though geological maps of Newfoundland (Baird, 1954; Williams, 1967) depict the rocks as part of the Fleur de Lys Supergroup.

The Horse Islands Group is separated into two formations, the Eastern Island Formation, composed mainly of semipelitic and mafic schist with minor metafelsite, and the Hit or Miss Point Formation, composed of alternating psammitic and semipelitic schist layers. The contact between the two sequences appears conformable and transitional on both the north and south coasts of Eastern Island. The stratigraphic sequence of the units is uncertain due to the structural complexity of the rocks; a few sparsely distributed younging directions based on grading within pebbly psammite in the Hit or Miss Point Formation suggest that it in part stratigraphically overlies the Eastern Island Formation. The original thickness of these formations is uncertain due to the intense deformation.

Eastern Island Formation

This formation is composed of steeply dipping, variably trending, semipelitic schist with psammitic and pelitic schist, greenschist, and minor metafelsite interlayers. It underlies approximately two-thirds of Eastern Island and outcrops on Black Rock, 2 km north of the island. Typical exposures of the unit are found in the cove directly east of the abandoned Horse Islands community and are here considered as the reference section for the formation.

The medium grained semipelite (O€he) is thin to medium layered, weathers gray to buff, and ranges from muscovite-chlorite-quartz-plagioclase to biotite-chlorite-plagioclase schist (Plate 4-31). The chlorite content increases significantly near greenschist layers. Locally, near the center of Eastern Island, a distinct garnetiferous muscovite-quartz semipelite occurs with fresh garnets up to 5 mm in diameter. Pebbly semipelite is common in the hills on the southeast portion of the island; clasts are commonly of feldspar and quartz, generally less than 1 cm in diameter.

Other metasedimentary components of the sequence include fine to medium grained, buff psammite and gray-green feldspathic pelite. The psammitic layers are generally less than 70 cm thick and increase in number and thickness toward the contact with the Hit or Miss Point Formation. The pelitic units range up to 15 m wide and are common at the

Plate 4-31: *Thinly layered, gray weathering semipelitic schist of the Eastern Island Formation affected by a late crenulation cleavage, along the northern coast of Eastern Island; Horse Islands Group.*

southeastern corner of Eastern Island. Tourmaline occurs locally within all of the metasediments.

Greenschist zones (O€hg), from less than 50 cm up to 300 m wide, occur throughout the metasedimentary sequence; the distribution of the thicker mafic schist units is shown on Figure 1-1. Their contact with the semipelitic schists is gradational in most places. The mafic units are dominantly composed of chlorite, epidote and plagioclase, and locally contain quartz-rich chlorite-epidote layers and garnetiferous greenschist. Knobby weathering epidote-chlorite nodules and rounded amphibole clasts are common in the mafic schists. Along the north coast of Eastern Island, marble bands up to 10 cm wide are interlayered with thinly (centimetre scale) banded mafic schists; elsewhere, small pods of carbonate are common in these units. In the cove immediately east of the Horse Island community, a black, fine grained, tourmaline-quartz vein, 50 cm wide, crosscuts banding in the mafic schist.

A few pink to light green weathering metafelsite layers up to 1 m wide occur within a thin but continuous mafic schist belt along the east coast of Eastern Island. They are quartz-muscovite-epidote schists that contain substantial amounts of sphene and large (up to 3 mm) clots of chlorite that appear to be pseudomorphs of garnet; one small unaltered garnet was noted in thin section. The pinkish cast of these rocks is produced by rose colored muscovite. Disseminated pyrite is common within the metafelsite. At one locality, a fine grained, angular epidote-chlorite-plagioclase clast occurs within a metafelsite layer that is bordered on one side by a greenschist band (Plate 4-32).

The general characteristics of the greenschist and metafelsite units and their interlayered nature and gradational contacts with the metasediments suggest a volcanic origin; the mafic schist may have largely been mafic fragmental rocks, tuff, and epiclastic volcanic sediments, and the metafelsite may have been silicic tuff.

Hit or Miss Point Formation

This is a moderately to steeply dipping monotonous sequence composed of alternating layers of medium to coarse grained psammitic and semipelitic schist with minor pelitic schist interlayers (Plate 4-33). It outcrops on the western portion of Eastern Island and on all of Western Island, where its outcrop width is approximately 3 km; it is named after typical exposures at its reference section, Hit or Miss Point. Individual layers are typically 20 to 60 cm thick and are generally persistent over hundreds of metres. Psammitic layers are thickest in the western parts of the formation; at Hit or Miss Point, psammite layers range up to 2 m thick. They are composed of quartz, plagioclase and muscovite, and are locally pebbly, containing clasts of blue quartz and feldspar. The semipelitic schists are similar to those in the Eastern Island Formation, though more commonly they are rich in epidote and chlorite and locally contain feldspar porphyroblasts up to 5 mm long. On the eastern portion of Western Island, chlorite-epidote pelitic schist zones up to 10 m wide are interlayered with the psammitic schist; the pelite contains local carbonate nodules up to 12 cm in diameter.

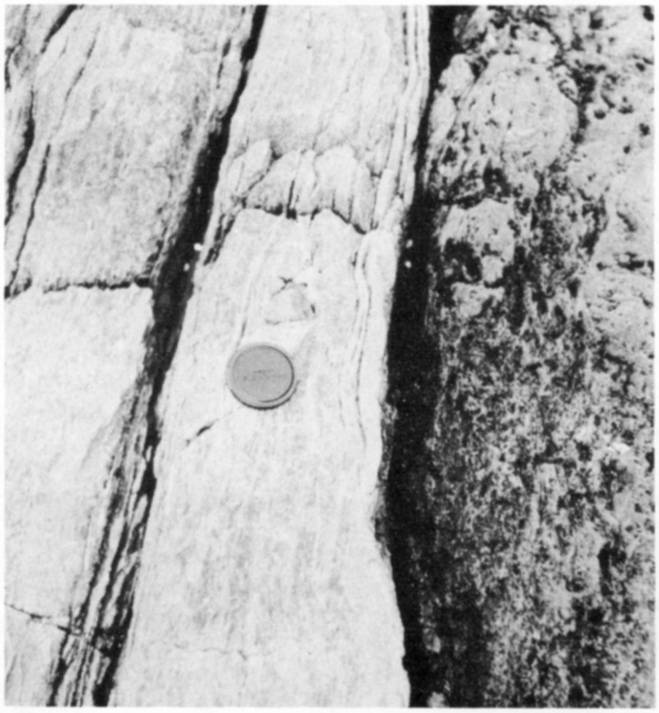


Plate 4-32: *Clast of fine grained epidote-chlorite-feldspar rock within metafelsite layer along the east coast of Eastern Island. Dark, vuggy rock type to the right is greenschist; Eastern Island Formation of the Horse Islands Group.*

Depositional Environment

The metaclastic portions of the Horse Islands Group were probably originally deposited as fine to medium grained graywacke, siltstone and mudstone. Their overall aspect is strikingly reminiscent of a flysch sequence in which primary turbidite structures have been tectonically obliterated. The Hit or Miss Point Formation may represent a proximal, partial equivalent to the more distal Eastern Island Formation. The greenschist and metafelsite members of the Eastern Island Formation indicate a volcanic influence in the basin where these metaclastics were deposited.

Age and Correlation

The age of this group is uncertain. Psammitic and semipelitic schist similar to the Horse Islands Group occurs within all parts of the Fleur de Lys Supergroup except the Birchy Complex. Lithically, the group appears to be most closely related to the Ming's Bight Group. The occurrence of metafelsite clasts associated with mafic rocks of the Pelée Point schist may be related to the metafelsite in the Eastern Island Formation of the Horse Islands Group; the metafelsics are unique to these two units within the Fleur de Lys Supergroup. The psammitic schists of the Horse Islands and Ming's Bight Group are also similar.

Blue quartz clasts occur in psammite of both the Ming's Bight Group and the Hit or Miss Point Formation. Thus, the Horse Islands Group may be, in part, correlative with the less pelitic Ming's Bight Group and may represent a more distal facies of the succession. Correlation of these two groups implies at least partial equivalency of the Horse Islands and Rattling Brook Groups. All of these stratigraphic links strongly bind the Horse Islands Groups to the Fleur de Lys Supergroup and, hence, indicate that the age of the group is within the age range of the supergroup, i.e. Late Hadrynian to Early Ordovician.

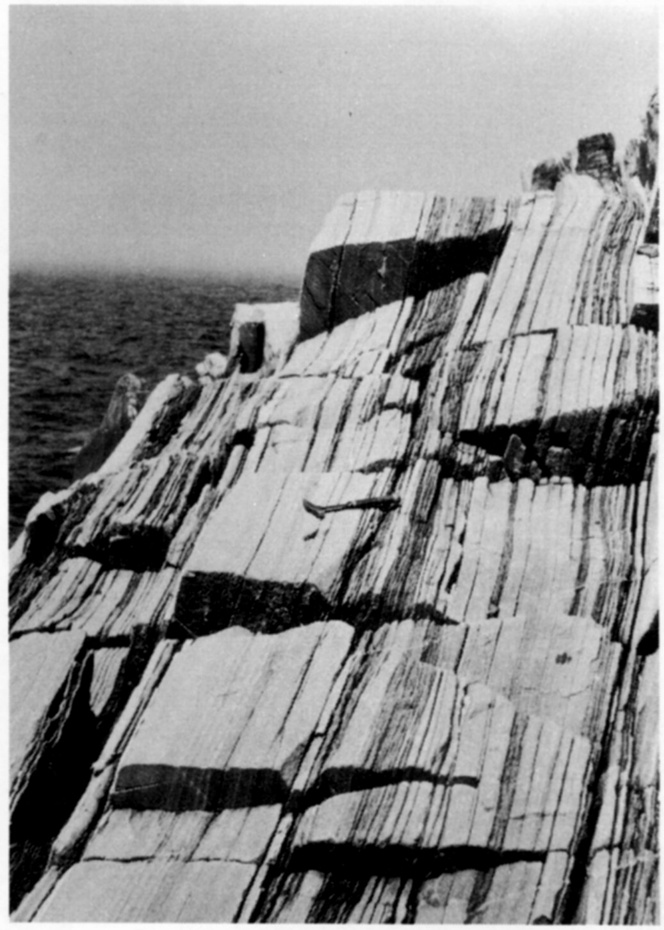


Plate 4-33: *Thin, alternating layers of psammitic and semipelitic schist of the Hit or Miss Point Formation, near its contact with the Eastern Island Formation, along the north coast of Eastern Island; Horse Islands Group.*

UNSEPARATED SCHISTS

Intensely deformed psammite, semipelite, garnetiferous semipelite, graphitic schist, greenschist and amphibolite outcrop between the Baie Verte Line (Figure 1-1) and the Wild Cove Pond Igneous Suite, and occur as large xenoliths within the suite. The combined effects of the paucity of outcrop, strong regional deformation, and tectonism caused by the intrusion of the Wild Cove Pond Igneous Suite prevent the separation of these rocks at the present scale of mapping.

Psammitic and semipelitic schists constitute most of the unseparated schists. Coarse grained, medium layered, buff weathering psammite, very similar to that of the Old House

Cove Group, is poorly exposed in the westernmost outcroppings of these rocks. It contains local massive amphibolite pods and layers identical to those of the Old House Cove Group. Gray to buff weathering, thinly layered semipelite is common throughout the schists. Typically, it is a fine grained quartz-feldspar-biotite-muscovite $\pm$chlorite schist. The semipelite contains local thin interlayers of graphitic schist and garnetiferous quartz-muscovite semipelite in the area of Black Lake. Greenschist and amphibolite layers also occur within the semipelites and are most common in the eastern portions of this belt. They are very similar to greenschist and amphibolite of the Birchy Complex, which occur along strike.

Rocks of these unseparated schists are probably correlative with the Old House Cove Group, Rattling Brook Group and Birchy Complex that occur immediately to the north along strike. Their environment of deposition and age were also probably the same.

PREKINEMATIC IGNEOUS ROCKS

There are several occurrences of deformed and metamorphosed igneous rocks in the Fleur de Lys Belt other than those described above with individual units. They range from ultramafic to granitic in composition. Almost all of these occurrences are considered to be of plutonic origin.

METAMORPHIC ULTRAMAFIC ROCK

Numerous bodies of metamorphosed ultramafic rock occur in the eastern, and in particular the northeastern, portions of the Fleur de Lys Belt. These bodies are most common in the Rattling Brook and Ming's Bight Groups and the Birchy Complex, but are rare in the Old House Cove Group. They range in size from small, centimetre scale fragments to large pods and sheetlike bodies up to 1.8 by 0.4 km. The smaller fragments and pods generally occur in the mélange zones of the Birchy Complex and the Ming's Bight Group, whereas larger pods and sheets commonly occur in high strain zones within all of the units. These bodies are typically associated with greenschist, amphibolite, metagabbro or graphitic schist, though less commonly they occur isolated in metasediments. In most cases, the bodies are concordant with the country rocks, but at Pacquet Harbour a layered meta-ultramafic/amphibolite body crosscuts the enveloping amphibolite and metasediments.

The mineralogy of the ultramafic rocks is dependent on the size of the body. Smaller occurrences as well as the sheared margins of larger bodies, such as those in the community of Fleur de Lys, are completely recrystallized to either coarse grained actinolite/tremolite $\pm$ talc $\pm$ carbonate $\pm$ fuchsite schist or, more rarely, talc-carbonate $\pm$ actinolite/tremolite schist. The common occurrence of the coarse bright green actinolite is the most distinguishing field marking of these rocks. The sheared margins, from 1 to 5 m wide, of the larger bodies grade into massive, rusty weathering blue-green serpentinite cores. The typical mineralogy is antigorite with subordinate talc and carbonate (most likely magnesite) (Kennedy, 1969, 1971; Bursnall, 1975). Primary igneous textures are commonly absent from these rocks, with the exception of rare bastites, i.e. antigorite pseudomorphs of pyroxene (Bursnall, 1975). Bursnall (1975) also noted small tension gashes filled with chrysotile in some of the metamorphosed ultramafic rocks north of Baie Verte.

Locally, the larger ultramafic bodies are internally brecciated, with rounded fragments of serpentinite in a network of talc-carbonate schist such as at Castor Pond. Williams (1977a) noted that the brecciation of serpentinite bodies in the mélange zones of the Birchy Complex predates their incorporation into the mélanges; thus, the brecciation of all of these bodies may be pretectonic with respect to the Fleur de Lys deformation. Some of the completely recrytallized bodies in the Ming's Bight Group are layered, indicating that some of them may have been cumulate in origin. Such an origin is supported by limited petrochemistry (see sample 1340083, Appendix III).

An unusual, medium to coarse grained, garnet-zoisite amphibolite occurs near the fault zone to the southwest of Wood Cove. The extent of the body is unknown due to poor exposure. The pale green amphibole in this rock is commonly twinned and displays the optical characteristics of cummingtonite. This rock may be an unusual manifestation (for the Fleur de Lys Belt) of an ultramafic rock, though the presence of zoisite indicates that the rock may originally have been metagabbro.

Where the contacts between the meta-ultramafic rocks and surrounding rocks were observed, the meta-ultramafic rocks are pretectonic with respect to the earliest regional fabric recognized in the country rocks (Kennedy, 1969, 1971; de Wit, 1972; Bursnall, 1975). In many cases, this fabric is the earliest regional fabric of the belt (Kennedy, 1969, 1971; de Wit, 1972; Bursnall, 1975). The one occurrence of meta-ultramafic rock in the Old House Cove Group is exposed in only one outcrop, and its contact relationships are unknown (Kennedy, 1969, 1971, personal communication, 1980), but it is most likely related to other ultramafic rocks in the belt.

The origin of these rocks is uncertain. Some workers have viewed them as pretectonic intrusions into the Fleur de Lys Supergroup (e.g. Kennedy, 1969, 1971), whereas others have interpreted them to be post-tectonic intrusions (Fuller, 1941). Bursnall (1975, 1979) indicated that they were tectonically emplaced into the belt. He cited the occurrence of ultramafic rocks in the Birchy Complex along a pretectonic fault, the Slaughter House Slide, as well as the occurrence of blastomylonitic and quartzitic slide zone rock types in the envelopes of some of these bodies. Bursnall (1975) concluded that they have the characteristics of Alpine-type ultramafics.

The meta-ultramafics in the Birchy Complex yield further evidence about the origin of these bodies. These ultramafic bodies are spatially associated with metagabbro and the South Cove schist. Considering other rock types in the complex, including mafic intrusives, trondhjemite, pillow lava, and manganiferous chert, the association has been interpreted as ophiolitic (Bursnall, 1975). Thus, the ultramafic bodies that occur throughout the supergroup may represent dismembered remnants of oceanic crust. I favor this interpretation based on the following evidence: (i) the obvious Alpine-type occurrence of many of the bodies, (ii) the ophiolitic association of those in the Birchy Complex, and (iii) the regional tectonic interpretation of the belt (see Chapters VII and IX).

Since the Birchy Complex abuts the Baie Verte Belt and is lithically and petrochemically correlative with portions of the belt, the meta-ultramafic bodies are most likely correlative with the ophiolitic basement of the Baie Verte Belt. These correlations suggest that the bodies are Cambrian to Lower Ordovician (see Chapter V).

The occurrence of these rocks in the Fleur de Lys Belt and their interpretation as remnants of oceanic crust raises a question about the nature of the basement to the Fleur de Lys Supergroup. In the south-central portion of the Fleur de Lys Belt, Grenvillian(?) gneisses appear to form basement to the supergroup (see East Pond Metamorphic Suite), but the easternmost representative of the supergroup, the Birchy Complex, appears to contain oceanic crust. Thus, it is conceivable that the northerly portion of the Rattling Brook Group and the Horse Islands and Ming's Bight Groups, as well as possibly the northeastern portion of the Old House Cove Group, were deposited on oceanic crust. Hence, the Fleur de Lys Supergroup may span a continental crust - oceanic crust juncture (also see Chapter VII).

AMPHIBOLITE

Amphibolite layers, pods and lenses occur randomly throughout the psammite, semipelite and metaconglomerate of the East Pond Metamorphic Suite and all of the Old House Cove Group. These are informally referred to here as Fleur de Lys amphibolites in order to distinguish them from the paleosome amphibolite of the migmatite at East Pond. De Wit (1972) identified three types of amphibolite in the metaclastic rocks designated here as part of the East Pond Metamorphic Suite, including: layered amphibolites with a metamorphic banding, mafic pods with a relict early fabric, and massive amphibolites which he correlated with amphibolites in the Old House Cove Group (de Wit, 1972; de Wit and Strong, 1975). This distinction of amphibolites is based on the presence or absence of structures that predate the main regional deformation. Such structures are rare (see Chapter VII) and, where present, are intensely overprinted by the main deformation. It is conceivable that they reflect an early phase of the main deformational event. Even at Southern Arm Head, where de Wit (1972) reported a layered amphibolite xenolith within amphibolite of the Old House Cove Group, the origin of the layered amphibolite can be interpreted in many ways. Therefore, the significance of these early structures is equivocal and requires further investigation. Since all of these amphibolites are grossly similar in mineralogy, they are herein described together, though similar geological histories are not implied.

Primary igneous textures are absent from the massive blackish green Fleur de Lys amphibolitic layers and pods. In most places, the amphibolite is concordant with the surrounding metasediments but, at Southern Arm Head and Pumbly Cove, near the mouth of Southern Arm, crosscutting relationships are clearly displayed. All variations in form, from distinct layers (up to 10 m wide) to isolated boudins and pods (Plate 4-34) were noted. In the East Pond Metamorphic Suite, the amphibolite is locally banded and commonly contains eclogitic cores. In contrast, amphibolite in the Old House Cove Group is characteristically spotted with plagioclase porphyroblasts (Plate 4-35), though layered varieties occur in places along the White Bay coast.

Banding (millimetre to centimetre scale) in the layered amphibolites is defined mainly by plagioclase-quartz-epidote

Plate 4-34: *Isolated boudin of Fleur de Lys amphibolite within psammitic schist of the Old House Cove Group at Lobster Harbour, White Bay.*

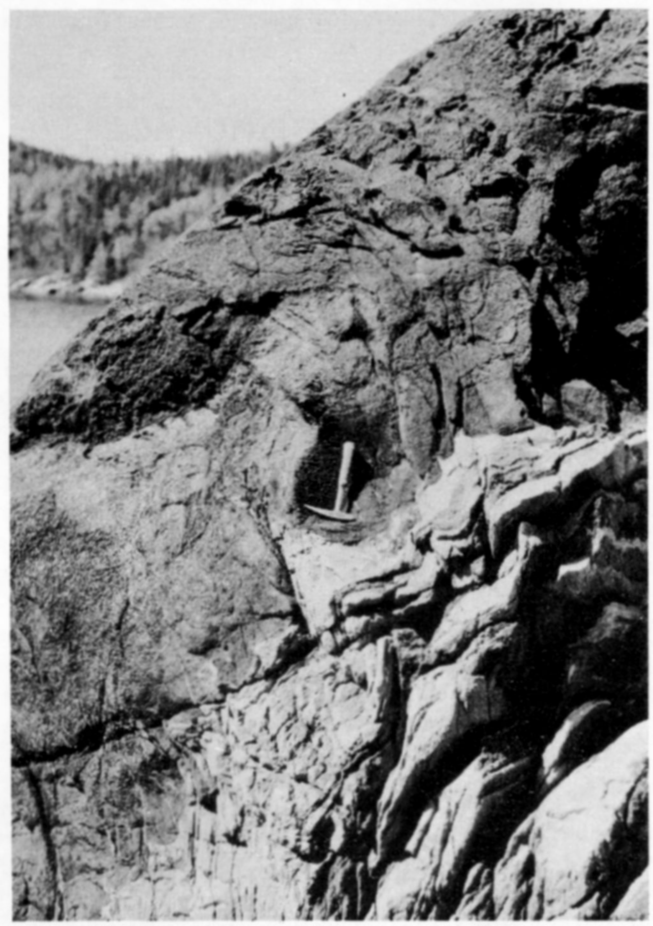


Plate 4-35: *Typical feldspar porphyroblastic aspect of Fleur de Lys amphibolites in the Old House Cove Group, near the mouth of Southern Arm, White Bay. Note structural modification of amphibolite-metaclastic contact.*

layers alternating with wider amphibole-rich layers; garnet defines thin bands in some cases. At East Pond, de Wit (1972) noted that the layered amphibolite is in transitional contact with epidosite and quartz-epidote layers. On the east shore of East Pond, plagioclase porphyroblast growth is more pronounced within the metaconglomerate country rock near the margins of a layered amphibolite. The significance of this association is uncertain and may be related to either original intrusion of the mafic sill or later metamorphic reactions.

The porphyroblastic amphibolite is characterized by plagioclase porphyroblasts up to 2 cm long that impart a distinct spotted texture to the rock. De Wit (1972) noted small clinozoisite ovoids within these bodies, and suggested that they represent relict amygdules. Large pods of rutile, hematite and ilmenite, up to 20 cm in diameter, occur sporadically within these layers.

The Fleur de Lys amphibolites are typically composed of hornblende, plagioclase, epidote and clinozoisite, quartz, biotite and garnet, with accessory sphene and opaques. Subidiomorphic hornblende forms up to 50% of the rock and generally defines the main fabric, where present. Plagioclase also forms up to 50% of the amphibolite; it occurs most commonly as idioblastic porphyroblasts of albite and oligoclase which are generally randomly distributed, but splotchy and linear aggregates also occur (de Wit, 1972). The porphyroblasts are commonly riddled with minute (<1 mm) garnets. In the layered amphibolites, plagioclase also occurs as xenoblastic poikiloblasts enclosing epidote and garnet. Yellow epidote and clinozoisite comprise up to 10% of the amphibolites, and quartz and biotite are generally less than 5% of the rock.

Eclogitic amphibolite occurs only in the East Pond Metamorphic Suite; it occurs mainly in the psammitic and semipelitic schists, but has also been found in the metaconglomerate (Neale and Kennedy, 1967; Church, 1969). The latter was confirmed during the present study. Eclogite also occurs in the paleosome of the migmatite at Pine Pond (see East Pond Metamorphic Suite). The eclogite always occurs in the core of amphibolite bodies, where pale green weathering, fine grained pyroxene distinguishes the eclogitic portions from the surrounding amphibolite. The eclogite is composed of omphacite (Church, 1966), garnet and quartz, with minor amphibole, biotite, rutile, and chlorite; sphene, epidote, calcite, apatite and opaques occur as accessories (de Wit, 1972; de Wit and Strong, 1975). The detailed petrography and geochemistry of these rocks was described by de Wit and Strong (1975). They showed that, geochemically, these rocks are identical to the amphibolites in the Old House Cove Group (see Chapter VI).

De Wit (1972) and de Wit and Strong (1975) indicated that the eclogitic amphibolites resulted from metamorphism of mafic dikes within a "dry" basement environment, whereas amphibolites in the Fleur de Lys Supergroup were subjected to "wet" metamorphic conditions which stymied the development of eclogites in the supergroup. De Wit and Strong (1975) used the presence of eclogites in rocks here assigned to the East Pond Metamorphic Suite as evidence that the suite represents a crystalline pre-Fleur de Lys basement. Reassessment of the stratigraphy of the suite in this report, as well as the occurrence of eclogitic amphibolites in definite supracrustal rocks of the Middle Arm metaconglomerate, invalidates the interpretation of de Wit (1972) and de Wit and Strong (1978).

Since the Fleur de Lys amphibolites crosscut layered metasediments and lack primary features, they probably originated largely as mafic dikes and sills. Local gradations with the country rock, as at East Pond, suggest that some were flows or volcaniclastic layers. The contrasts between amphibolites of the East Pond Metamorphic Suite and the Old House Cove Group, such as porphyroblast development, layering, and mineral assemblages, may reflect different conditions of deformation and metamorphism for the two units, but not necessarily different origins.

Based on geochemical similarities (de Wit and Strong, 1975) and stratigraphic considerations, the eclogitic amphibolite and other amphibolites in the Old House Cove Group have been correlated with the probably Eocambrian Light House Cove Formation and its feeders in the Strait of Belle Isle area (Williams and Stevens, 1969; de Wit, 1972; de Wit and Strong, 1975). Dikes near the Strait of Belle Isle are identical to mafic dike swarms throughout the Grenvillian Long Range Inlier of western Newfoundland (Strong, 1974). The latter dikes were dated at 605 Ma by Stukas and Reynolds (1974), supporting an Eocambrian age. In the present study, it is shown that the Fleur de Lys amphibolites are geochemically correlative with the Oody Mountain and Garden Cove amphibolites as well as those in the Rattling Brook Group (see Chapter VI). These correlations all suggest that the Fleur de Lys amphibolites are Eocambrian to Lower Ordovician.

METAGRANODIORITE

Neale and Kennedy (1967), Church (1969) and Kidd (1974) noted fine to medium grained felsic rocks that intrude the Birchy Complex between Kidney Pond and Red Cliff Pond. The intrusions occur as small bodies and numerous apophyses in surrounding rocks; only one body is mappable at the scale of Figure 1-1. Kidd (1974) reported that this main body is only weakly foliated and has relict plutonic textures; he described the petrography of the intrusion as follows:

> Aplitic veins are present cutting coarse grained granular rock, that when unfoliated consists of 60% andesine, 30% quartz, some of which is poikilitic, and interstitial large muscovite flakes, with accessory epidote, apatite, sphene and calcite. When the rock becomes foliated, and the original grains augened by a developing muscovite schistosity, the plagioclase is converted to albite, and a proportion of the albite grains contain myriads of aligned fine grained muscovite inclusions.

In the area of Kidney Pond, the felsite is strongly foliated and banded on a centimetre scale, with the concentration of epidote and/or chlorite as well as grain size defining the banding. Kidd (1974) demonstrated that these small intrusions are either synchronous with, or predate, the first regional deformation in this area.

Felsic intrusive rocks are very rare in the Fleur de Lys Belt. This metagranodiorite may be related to the only other known occurrences in the belt, including the felsic dikes that crosscut gneissic banding in the East Pond Metamorphic Suite (described with the suite) and the Crow Head Intrusion, described below. Kidd (1974) noted that the metagranodiorite occurs adjacent to the Burlington Granodiorite, but on the opposite side of the Baie Verte Line. On the basis of this relationship, he suggested that the metagranodiorite was a contiguous part of the Burlington Granodiorite. This correlation is plausible, since the Birchy Complex intruded by metagranodiorite is ophiolitic as is the Pacquet Harbour Group, host rock to the Burlington Granodiorite.

CROW HEAD INTRUSION

An unusual stock of fine grained, light green to cream colored, felsic rock crosscuts metasediments of the Old House Cove Group at Crow Head, on White Bay. The stock is small, outcropping for approximately 750 m along the coast. The contact of the body with the country rock varies from a sharply defined intrusive contact to a diffuse boundary. Psammite and amphibolite common to the Old House Cove Group are included as xenoliths in the felsite near the contact. The stock is homogeneous, but it is locally brecciated. De Wit (1972) first reported this intrusion and described it as follows:

> The composition is variable, but is essentially of quartz, actinolitic amphibole (very pale green - blue green; 2V 70; extinction angle 23°), epidote, zoisite and garnet, with sphene, biotite, chlorite and plagioclase as accessories. Quartz is dominant, never less than 50%, usually more, frequently up to 90%. Epidote and zoisite sometimes make up 40%. The rock is conspicuous because the actinolite forms bright green to black, randomly orientated poikiloblasts, generally 3-4 mm, but often reaching 2 cm in length.

De Wit (1972, 1980) considered the intrusion and adjacent country rocks to represent a Grenville basement inlier within the Old House Cove Group. In the present study, no difference was noted between the enveloping rocks and typical Old House Cove Group metaclastics; therefore, the intrusion is here considered to cut the Old House Cove Group. It is affected by all of the fabrics in the country rocks and, hence, is pretectonic. The origin of the intrusion is uncertain.

SUMMARY OF DEPOSITIONAL HISTORY AND PREKINEMATIC IGNEOUS ACTIVITY

Prekinematic rocks of the Fleur de Lys Belt on the peninsula appear to form a coherent depositional and magmatic package at the eastern margin of the Humber Zone. Even though the affects of intense tectonism are apparent in these rocks, i.e. transposition of original stratigraphy and obliteration of primary features, it is still possible to reconstruct the broad scheme of deposition and intrusion in the belt prior to the regional tectonism. Recognition that the belt intervenes between platformal deposits of the westerly Humber Zone and the oceanic rocks of the easterly Dunnage Zone is a decisive aid in reconstructing the prekinematic Fleur de Lys stratigraphy.

Cover rocks of the belt appear to have been deposited on both continental and oceanic crust. The White Bay Group overlies granitic gneiss of probable Grenvillian age, and the metasediments of the East Pond Metamorphic Suite appear to overlie reworked migmatite and gneiss, also probably of Grenvillian age. In contrast, the Birchy Complex is interpreted to be ophiolitic, and thus originally oceanic crust. However, the extent of this crust is uncertain; prekinematic ultramafic bodies within the Rattling Brook Group may indicate an original oceanic basement to this unit. The nature of the basement to the Ming's Bight and Horse Islands Group is uncertain.

Metasedimentary rocks of the cover sequence appear to represent both slope and submarine basinal deposits. The White Bay and Rattling Brook Groups appear to represent the slope deposits; they conformably interfinger with the basinal deposits of the Old House Cove Group. The Ming's Bight and Horse Islands Groups may represent distal basinal facies deposits.

Mafic meta-igneous rocks occur throughout the cover sequence and are both extrusive and intrusive. Based on geochemical data and geochemical correlation (see Chapter VI) most of the amphibolites in the main outcrop belt, except those of the Birchy Complex, appear to be rift facies igneous rocks. The Birchy Complex rocks appear to be ophiolitic and geochemically distinct from these rift-related rocks. Greenschist and amphibolite of the Pelée Point schist and in the Eastern Island Formation are of uncertain affinity. These units are the only ones in the Fleur de Lys Belt known to contain felsic volcanic fragments. This may indicate a link between these easterly units and the felsic volcanic - bearing Pacquet Harbour Group (see Pacquet Harbour Group). The metagranodiorite in the Birchy Complex may well represent a deformed and metamorphosed apophysis of the Burlington Granodiorite (see Chapter V), which also intrudes ophiolitic rocks. The Crow Head intrusion may represent a felsic differentiate of the rift facies mafic rocks in the belt.

Based on these observations, it appears that deposition and accompanying intrusion in the Fleur de Lys Belt took place along the interface between continental and oceanic crust. The metasedimentary rocks appear to have been derived largely from a continental source and deposited on a slope and in a basin. The meta-igneous rocks reflect a rift environment and the generation of oceanic crust. These characteristics in conjunction with the regional setting of the belt, between platformal rocks of the Humber Zone and oceanic rocks of the Dunnage Zone, indicate that the belt represents the ancient continental margin of eastern North America.

SYNKINEMATIC ROCK TYPES

Rocks of two separate intrusive events and a unit of highly strained schists originated during the regional deformation of the Fleur de Lys Belt. The intrusive rocks, the Dunamagon Granite and mafic dikes in the Ming's Bight Group, are also intrusive into the Baie Verte Belt; their origins appear to be closely associated with that belt and, hence, are described in Chapter V.

HIGH STRAIN SCHISTS

The contact between the Fleur de Lys Supergroup and the East Pond Metamorphic Suite is tectonic and is marked in many places by a steeply dipping zone, up to 750 m wide, of highly strained, plagioclase porphyroblastic mica schists. These were first recognized by de Wit (1972), who termed them "tectonic schists."

Characteristically, the high strain schists are brownish black, medium to coarse grained, plagioclase porphyroblastic muscovite-biotite schists (Plate 4-36). Layering in both the Fleur de Lys and East Pond rocks grades into these completely reconstituted, tectonic schists, and gradually disappears concomitant with plagioclase porphyroblast growth and intensification of the schistosity (de Wit, 1972). These schists are locally difficult to distinguish from the typical Fleur de Lys schists,

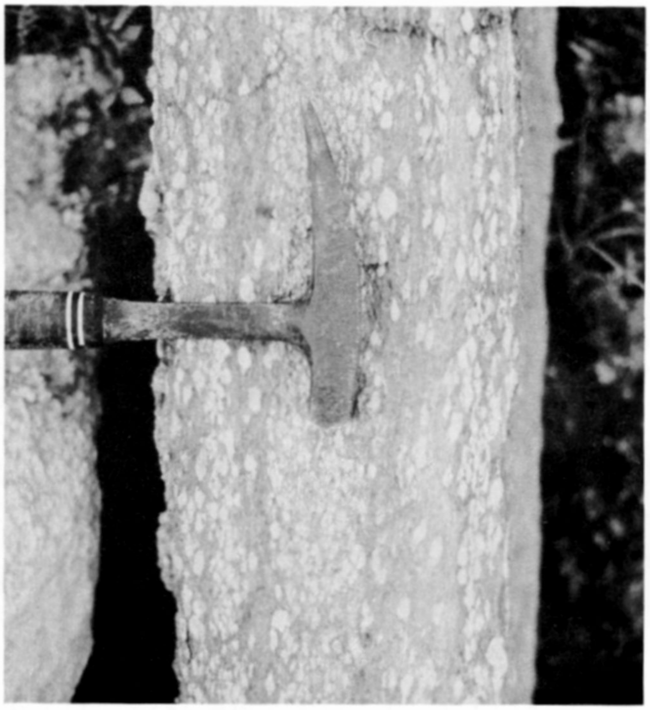


Plate 4-36: *Typical feldspar porphyroblastic, micaceous high strain schists that form the tectonic boundary between the East Pond Metamorphic Suite and the Old House Cove Group along the Westport Road.*

though the high strain schists (i) form thick zones at the boundary between the Old House Cove Group and East Pond Metamorphic Suite, (ii) are highly strained, with local zones of ribbon quartz, (iii) have either vaguely defined or no banding, and (iv) have local growth of large (>5 cm) feldspar-quartz segregations and feldspar porphyroblasts. Locally, highly strained amphibolite boudins, characteristic of the surrounding units, occur in these zones. De Wit (1972) reported boudins more than 50 cm long of East Pond psammitic rocks in which an internal banding is askew to the main fabric in the schists.

These high strain schists are apparently absent around large parts of the eastern outcrop area of the East Pond Metamorphic Suite. At Gull Pond, the contact with the Old House Cove Group is unexposed. Considering the poor exposure and polydeformation of these rocks, it is possible that unrecognized tectonic contacts occur where the high strain schists are absent.

De Wit (1972, 1980) considered these rocks to have formed during the deformation of the Fleur de Lys Supergroup, as a result of intense flattening; evidence of shearing is absent. The relationships of banding as described above support this concept.

POSTKINEMATIC INTRUSIVE ROCKS

Three types of intrusive rocks, two granitoid and one diabasic, crosscut the dominant structures in the Fleur de Lys Belt. The granitoids outcrop within the main portion of the belt on the peninsula, whereas postkinematic diabase dikes are apparently confined to the Horse Islands.

WILD COVE POND IGNEOUS SUITE

Kidd (1974) first termed the dioritic to granitic rocks outcropping between Wild Cove Pond and Black Lake, the Wild Cove Pond complex. These rocks form a small portion of a composite batholith that, in the present investigation, has been traced southward to Birchy and Sandy Lakes (Figure 1-1). This expanse of intrusive rock, including three small satellite bodies at the northern margin of the pluton, is herein termed the Wild Cove Pond Igneous Suite. The main body of the batholith is approximately 60 km long and reaches a maximum width of 23 km in the area north of Birchy Lake. It underlies two massive areas, the northerly of which is approximately one-third the size of the southerly portion and connected to it by an approximately 2 km wide "isthmus" of granitoid rocks northeast of Black Lake (Figure 1-1). At the eastern end of the "isthmus", the Black Lake Fault severs the northerly massif from the body to the south. The two portions of the main body are lithologically indistinguishable, though the proportion of different rock types in each region is not the same (Figure 1-1). However, a reconnaissance geochemical study of the whole batholith indicates that the two massifs are geochemically distinct and most likely had separate sources (see Chapter VI). Numerous xenoliths occur in both portions of the batholith, though those to the south are more diverse.

Field Aspect

Four broad phases of the batholith have been distinguished in the field, including diorite, granodiorite, biotite granite, and two-mica granite. The scattered distribution of these phases prevents their separation at the present scale of mapping, though biotite granite, the most common phase, apparently occupies a large area to the west and northwest of Upper Indian Pond (Figure 4-2). The two-mica granite is the only phase confined to fairly well defined areas, as shown in Figures 1-1 and 4-2.

DIORITE

Dioritic rocks comprise the oldest phases of the complex as they are either intruded by or included within all other phases. The larger outcrop areas of diorite tend to occur either near or at the margins of the pluton. The diorites are particularly common in the northerly portion of the batholith, near Wild Cove Pond and Black Lake; they are also common in the area east of Kitty Pond (Figure 4-2). In addition, the two small satellite bodies directly north of Wild Cove Pond are composed totally of diorite. There are four major varieties of diorite within the suite, including hornblende diorite, hornblende-biotite diorite, biotite-hornblende ± quartz diorite, and biotite-quartz diorite. Single outcrops of medium grained hornblende gabbro and coarse hornblende gabbro are associated with the hornblende-biotite ± quartz diorite near Lynx Pond and on the eastern shore of Black Lake, respectively. The hornblende diorite is generally gray-green, equigranular and fine to medium grained; locally, it is feldspar porphyritic, with phenocrysts up to 3 cm long. The biotite phases have a larger range in grain size than the hornblendic varieties and are commonly more feldspar porphyritic that the latter. Both the hornblende and biotite phases range from isotropic to foliated. Conspicuous, large (up to 5 mm) euhedral sphene is ubiquitous throughout all phases.

Field observations indicate an increase of quartz concomitant with the increase of biotite and decrease of hornblende. In the biotite-hornblende-quartz phases, quartz and hornblende form an unusual texture, informally termed here "hog-eye" texture after its glassy, vacuous appearance resembling the cold stare of a pig. Hog-eye texture is characterized by clear blebs of quartz, up to 5 mm in diameter, totally encased in black hornblende prisms, up to 5 mm long. The texture, apparently confined to the Wild Cove Pond area, imparts a distinct dark spotted aspect to the otherwise homogeneous gray diorite.

Primary layering within the diorite occurs locally in the area of Wild Cove Pond (Plate 4-37). On the west side of the pond, it is defined by rhythmic layers of biotite-hornblende ± quartz diorite that grade into a leucodiorite, which is in sharp contact with the darker part of the neighboring layers. Feldspar phenocrysts up to 2 cm long form nebulous zones throughout the layered sequence, but tend to cluster within the more leucocratic diorite. The phenocrysts commonly display primary flow alignment, particularly near the contacts of two layers. Flow alignment is also relatively common within the nonlayered diorite. Elsewhere, as on the east side of Black Lake, biotite-hornblende ± quartz diorite is regularly layered with biotite ± hornblende granodiorite. These layered sections all appear to be of very limited extent and may represent older portions of the batholith that have been totally riddled and isolated by younger phases of the pluton.

The contacts between the varieties of diorite vary from transitional to intrusive. Hornblende and hornblende-biotite diorite form the oldest varieties, as they are intruded by the quartz-bearing phases. Typically, the contacts between the quartz-free diorite phases are transitional, but at the north end of Wild Cove Pond hornblende-biotite diorite intrudes and brecciates the hornblende diorite. The contacts between the quartz-bearing phases are gradational. All other rock types of the suite clearly intrude the diorite, though, in places, the diorite is in gradational contact with the granodiorite and granite.

The contacts between the diorite and other phases of the suite are intrusive in most places; however, locally they are transitional. Sharp contacts between the diorite and the more leucocratic phases are the most common, and are of two forms. In many cases, the more leucocratic phases intrude, and are chilled against, the diorite. Intrusion breccias and agmatite (Plate 4-38) are formed typically at these contacts. Elsewhere, however, large patches of leucocratic phases occur, without chilled margins, in the diorite; unchilled apophyses of leucocratic rocks penetrate the dioritic body. This relationship suggests that the diorite was hot at the time of intrusion of the granodiorite and granite. Less commonly, the diorite is in gradational contact with the granodiorite and granite. On the east side of Black Lake, the diorite and granodiorite form gradational layers. Elsewhere, as on the southwest shore of Black Lake, at Wild Cove Pond, and in the area west of Barren Pond, the diorite grades into either granodiorite or granite over approximately 5 to 10 m.

GRANODIORITE

The granodiorite phase of the suite contains biotite ± hornblende, and is typically off-white to gray in color. The granodiorite ranges from fine to coarse grained, though medium grained granodiorite is most common. Texturally,

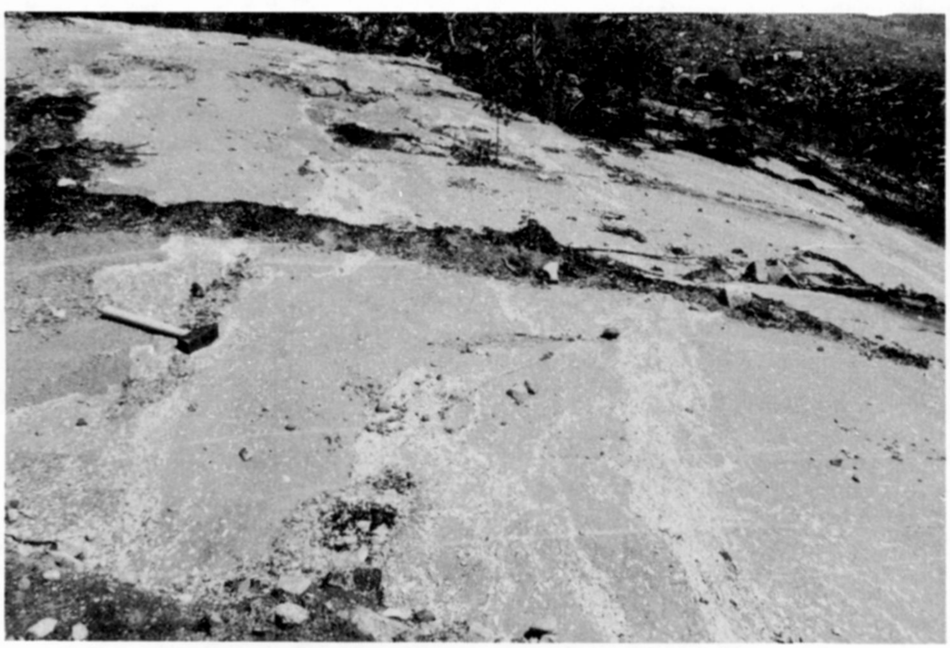


Plate 4-37: *Primary layering in diorite of the Wild Cove Pond Igneous Suite along the west side of Wild Cove Pond. Patchy white zones represent concentrations of plagioclase phenocrysts.*

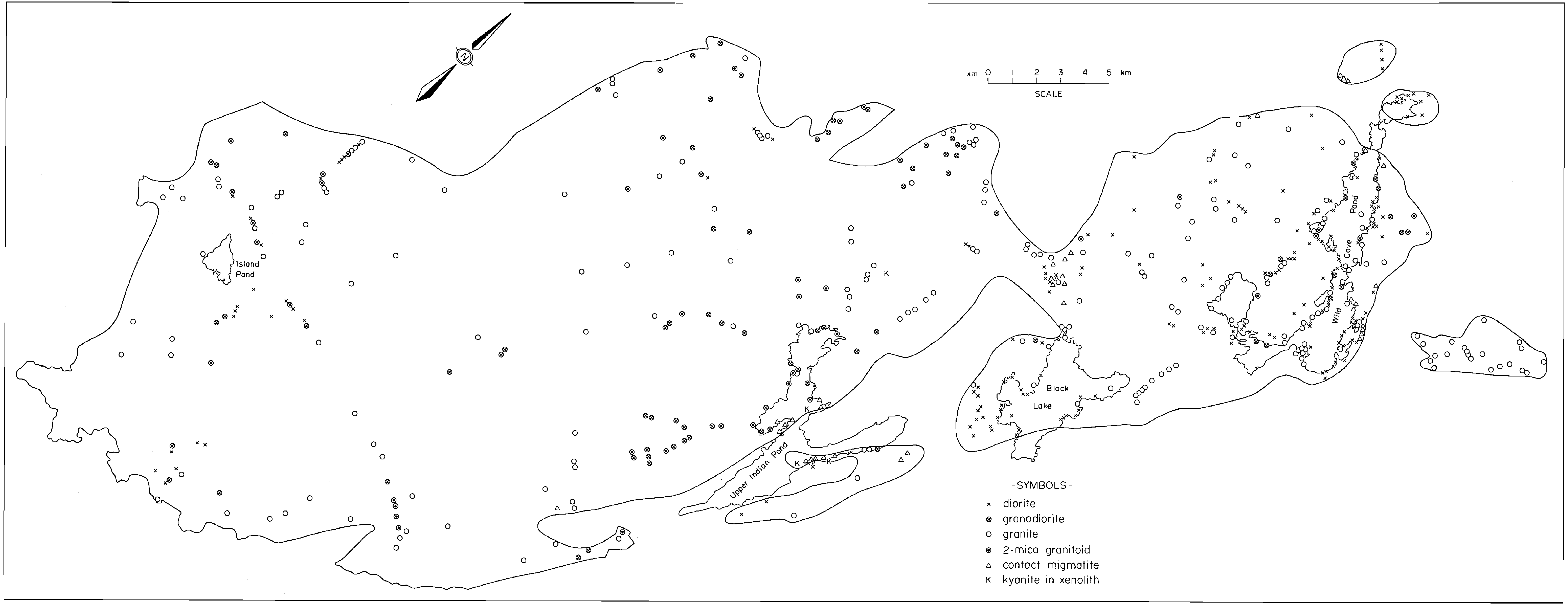


Figure 4-2: *Outcrop map of the Wild Cove Pond Igneous Suite showing distribution of phases.*

it ranges from equigranular to feldspar porphyritic and occurs in both foliated and isotropic forms. The granodiorite most commonly occurs in small patches associated with biotite-quartz diorite and biotite granite, though two large areas in the southern massif (Figure 1-1) are found near the margin of the batholith. Detailed petrography has not been undertaken in these larger areas; future work may indicate a substantial amount of granite rather than granodiorite here. In many places, the granodiorite is cut by the biotite granite, but locally, as on Wild Cove Pond, they are intermixed.

Plate 4-38: *Intrusion breccia between younger granite (light) and older diorite (dark) on the western shoreline of Upper Indian Pond; Wild Cove Pond Igneous Suite (photograph by J. Gagnon).*

BIOTITE GRANITE

Biotite granite is the most abundant rock type in the suite. It is common throughout the batholith, but forms extensive portions of the central and southern areas of the southerly massif. The biotite granite occurs as two distinct phases, including a megacrystic phase (Plate 4-39) and an equigranular phase. The pink to gray megacrystic variety is characterized by alkali feldspar megacrysts, generally less than 3 cm long, in a gray matrix of quartz (up to 3 mm), biotite, and subordinate euhedral sphene. The megacrystic granite is most common in the northerly massif, though it also occurs immediately east of Tea Bay on Birchy Lake; here, unusually large megacrysts, up to 10 cm long, are developed (Plate 4-40). In addition, it occurs sporadically elsewhere in the southerly massif. Flow alignment of the megacrysts is common throughout the suite.

The equigranular phase is generally pale gray to fine to medium grained. Locally, west of Upper Indian Pond and south of Barren Pond, it contains small (<2 mm) red garnets. The equigranular granite generally forms large areas in the more insular portions of the southerly massif. The relationship of the equigranular and megacrystic phases appears to be gradational. Usually, both phases are massive, though they are foliated locally.

In the southern, particularly the southwestern, portion of the batholith, the megacrystic granite contains wispy melanocratic layers that impart a gneissic aspect to the rock

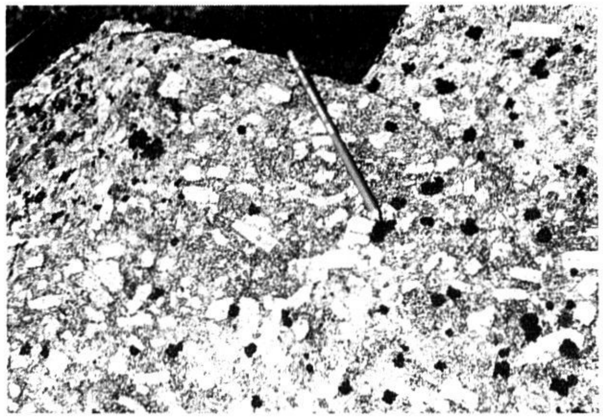


Plate 4-39: *Megacrystic phase of the biotite granite, inland west of Wild Cove Pond; Wild Cove Pond Igneous Suite (photograph by W. Muggridge).*

Plate 4-40: *Very coarse phase of megacrystic biotite granite at Tea Bay, Birchy Lake; Wild Cove Pond Igneous Suite.*

(Plate 4-41). The melanocratic layers are defined by the concentration of biotite; megacrysts are also present in these layers. The banding cannot be traced for more than a few outcrops in any one area, though it is parallel in separate outcrops over the large area just southeast of Kite Pond. The origin of this feature is uncertain; it may be related to flowage or it may represent relicts of highly assimilated inclusions.

The biotite granite contains subordinate phases of syenite, quartz syenite, monzonite and quartz monzonite, all of which

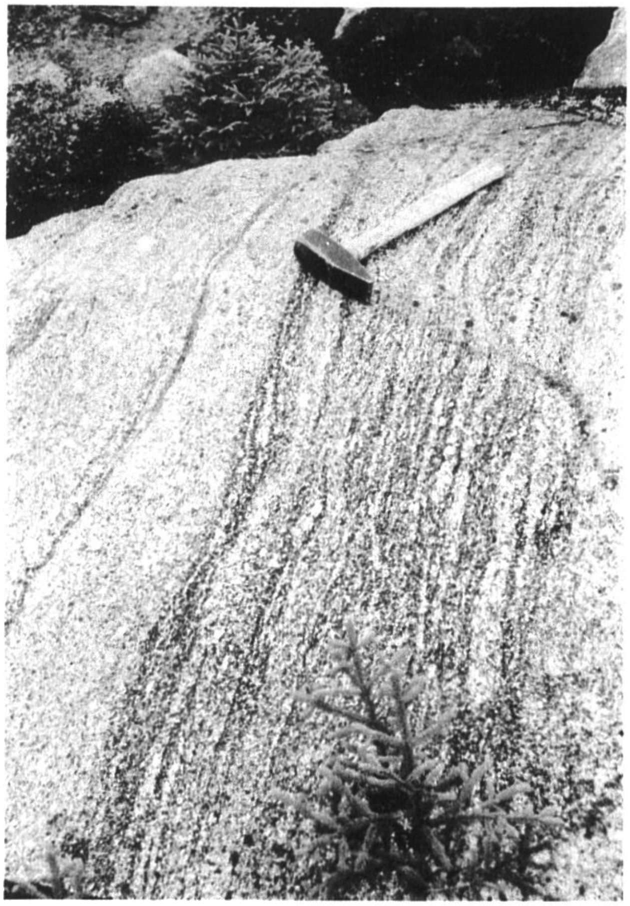


Plate 4-41: *Gneiss-like layering within the biotite granite, just southeast of Kite Pond; Wild Cove Pond Igneous Suite.*

are gradational with the granite. These phases have been identified microscopically. The satellite body immediately west of Kidney Pond is composed entirely of quartz monzonite.

The biotite granite intrudes and contains pods of most other phases of the granite, with the exception of the two-mica phases. Many biotite-bearing aplites that cut the diorite and granodiorite are probably related to the biotite granite. In addition, transitional contacts with other rock types are known, and have been described above.

TWO-MICA GRANITOIDS

The two-mica granitoids are the least abundant of the four major phases of the suite separated in the field. Biotite-muscovite granite is the most common of the two-mica granitoids, though two-mica granodiorite, quartz syenite, and quartz monzonite form subordinate phases. The granite composes most of the larger concentrations of the two-mica granitoids on the northwest side of Upper Indian Pond and on the abandoned section of the Trans Canada Highway north of Birchy Lake. Biotite-muscovite granodiorite occurs at both of these areas as well as at Wild Cove Pond, and biotite-muscovite quartz syenite and quartz monzonite also occur at Upper Indian Pond. The quartz syenite is also present in one exposure to the southeast of Hampden.

The two-mica granitoids are generally pinkish white to pale gray in color; they are fine to medium grained, equigranular and generally massive, though at the locality north of Birchy Lake they are clearly foliated locally. In places, particularly north of Birchy Lake, the two-mica granitoids are garnetiferous. In addition, aplite and pegmatite related to these rocks are muscovite- and garnet-bearing. Muscovite-biotite pegmatites around Wild Cove Pond are probably related to the two-mica granitoids.

The biotite-muscovite granitoids are the youngest phase of the suite; these granitoids form small stocks with radiating aplitic and pegmatitic dikes that cut all other phases in the batholith.

DIKES

In addition to the main phases described above, several dikes, most likely associated with the suite, intrude the country rocks around the periphery of the batholith. Two andesitic dikes, up to 2 m wide, have been noted cutting rocks of the Fleur de Lys Supergroup. One is located on the Baie Verte highway directly west of the north end of Micmac Lake, and the other occurs on the tributary that empties near the mouth of Big Chouse Brook. The dike on the highway is vesicular, containing free quartz and sparsely distributed large feldspar megacrysts (up to 3 cm) in a fine grained, brownish gray matrix. The dike on the brook is a brownish gray, fine grained diabase to andesite.

Kidd (1974) reported felsic dikes from the Wild Cove Pond area. He noted an 8 m wide, chalky weathering, gray microspherulitic rhyolite dike from the area near the intersection of the Baie Verte highway and the logging road that leads to Wild Cove Pond, as well as a gray microporphyritic rhyolite dike that is tectonically juxtaposed against the northernmost part of the Micmac Lake ultramafic body.

Petrography

Rock types identified from thin sections of the Wild Cove Pond Igneous Suite are outlined in Figure 4-3, the IUGS classification scheme (Streckeisen, 1973) for plutonic rocks. It should be emphasized that there may be some sampling bias in this figure since it represents only samples that were sectioned.

The main minerals of the suite include plagioclase, alkali feldspar, quartz, biotite and hornblende, with accessory sphene, primary muscovite, apatite, zircon, garnet, allanite, hematite and magnetite. Significant secondary alteration minerals include muscovite, epidote and chlorite. The occurrence of all of these minerals is generally similar throughout the suite, with only slight variations between the different rock types. Rocks of the suite typically display a xenomorphic to hypidiomorphic granular texture.

Plagioclase occurs as blocky to lath-shaped, generally subhedral grains that range in size from 1 to 4 mm; in the diorite, it is generally subhedral to anhedral. The most common plagioclase is oligoclase; however, andesine occurs in the more mafic diorites. Locally, the feldspar is zoned, with selective alteration of the core zones; the alteration is generally sericitization, but saussuritization has been noted. Commonly, the more mafic rocks are more highly altered; in the gabbro, plagioclase is completely sericitized. Carlsbad, albite,

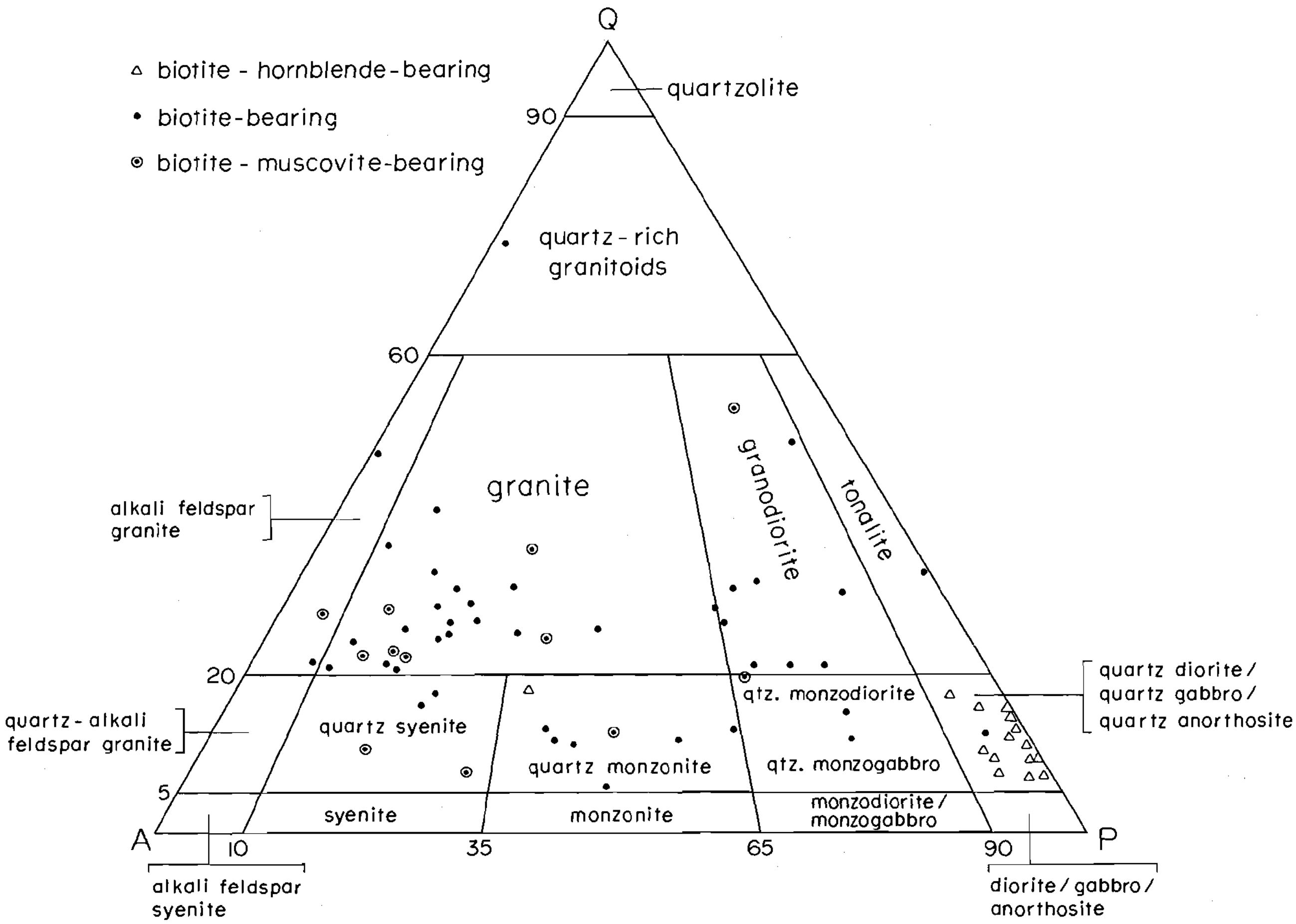


Figure 4-3: *Lithologic classification of samples from Wild Cove Pond Igneous Suite. Diagram after Streckeisen, 1973.*

and Carlsbad-albite twins are common, whereas pericline twinning is less common. In one section of the biotite granite, blebby antiperthite was noted, and in one section of the granodiorite a few plagioclase crystals exhibit mortar texture along their boundaries. Commonly, in the foliated portions of the pluton, the alignment of plagioclase helps define the foliation.

Alkali feldspar occurs as irregular, angular to ovoid, anhedral to subhedral grains up to 4 mm in size that are commonly zoned in the alkali granite. It also occurs as megacrysts up to 5 cm long. Microcline is the main alkali feldspar in the granodiorite, granite and alkali granite, whereas equal or greater amounts of orthoclase occur in the other rock types. Carlsbad twinning is rare. The alkali feldspars are only slightly altered to very fine grained epidote. Stringlet perthite is common, but not ubiquitous.

Quartz typically forms irregular anhedral grains and clusters of grains less than 4 mm in diameter. Locally, the quartz is highly serrated and stretched, and displays slightly undulose extinction, whereas surrounding components of the rock are virtually fresh. Locally, in the biotite and alkali granites, quartz is vermicular, forming re-entrant knobs and blebs (< 1 mm) in the feldspar and biotite. More commonly, however, thin fingerlike myrmekitic intergrowths occur in all rock types in the suite.

Biotite and hornblende typically form subhedral grains with a maximum dimension of 4 mm. The biotite is most commonly a dull, dark shade of either olive green or gray-brown, though, in almost all unaltered samples of granodiorite and granite, it is bright reddish brown. Locally, the biotite is altered to an anomalous purplish to brownish birefringent chlorite. The hornblende is also a dull brownish green to olive green, although bright green varieties occur locally. Some hornblende is twinned.

Sphene is the most common and conspicuous of the accessory minerals; it forms euhedral to subhedral crystals up to 5 mm across. Muscovite is less common, and never forms more than 5% of the rock. In many places, it is associated with biotite. Apatite, zircon, garnet, allanite, hematite and magnetite are minor constituents.

Xenoliths

The Wild Cove Pond Igneous Suite encloses numerous xenoliths that range in size from millimetre scale inclusions up to rafts resolvable at the present map scale (Figure 1-1). They commonly occur throughout the batholith, though they are generally larger and more numerous in the zones of contact migmatite at the batholith margins. Almost all of the inclusions are metasedimentary schist and amphibolite from the country rocks of the Fleur de Lys Supergroup and East Pond Metamorphic Suite. Xenoliths of ultramafic rocks, garnetiferous metagabbro and granitic gneiss are only found locally. Considering their size, abundance and origin, most of the xenoliths probably represent roof pendants within the pluton. In addition, inclusions of older portions of the suite occur within younger portions. In particular, dioritic inclusions are quite common in the granodiorite and biotite granite near their contacts with the diorite.

The metasedimentary and amphibolite inclusions are ubiquitous and range from well defined angular blocks and rafts (Plate 4-42) to almost totally assimilated schlieric and nebulitic inhomogeneities within the granite. The boundaries of many of the larger xenoliths display several features of the contact migmatites that occur at the margin of the batholith (see below). Generally, in the more highly assimilated xenoliths, alkali feldspar megacrysts are formed. The melanocratic layering described above in the biotite granite may represent an advanced stage of assimilation of a layered metasediment. Some of the semipelitic schist blocks in the contact migmatitic zone at Indian Pond and one xenolith at Island Pond contain kyanite and sillimanite (Figure 4-2).

Plate 4-42: *Well defined xenolith of semipelitic schist in biotite granite, Wild Cove Pond; Wild Cove Pond Igneous Suite (photograph by W. Muggridge).*

Meta-ultramafic and metagabbro rafts with dimensions up to 1 by 7 km are confined to the area north of Tea Bay, Birchy Lake. The two are closely associated, and in places compose part of the same xenolith. The xenoliths are all recrystallized and, thus, are not considered to be genetically related to the batholith. The largest of these xenoliths (Figure 1-1) is largely defined by aeromagnetic data (Geological Survey of Canada, 1970) due to poor exposure. The similarity of these inclusions, as well as their proximity to ultramafic rocks along the Baie Verte Line, implies that they may be remnants of the Advocate Complex. Their occurrence in the suite is unusual in that (i) they are confined to a small region, and (ii) the area to which they are confined does not abut ultramafic country rock. Although the batholith is in contact with ultramafic rocks further to the east, no such xenoliths occur there.

One large, angular, well defined raft of granitic gneiss occurs near the ultramafic xenoliths immediately north of Lidstones Brook. It is white weathering, with strongly deformed, discontinuous layers and lenses of alternating biotite-rich and felsic-rich compositions. This xenolith may represent a protoclastic remnant of an early phase of the suite, though its aspect is very similar to the granitic gneiss in the Oody Mountain Amphibolite and Grenvillian granitic gneiss that outcrops on the west side of White Bay. If this does represent a Grenvillian relict, its position in the pluton, amidst xenoliths of oceanic affinity, is intriguing.

Internal Structural Relationships

Flow alignment of minerals and primary layering were the only structures noted in the main portion of the batholith; migmatitic structures occur at the margins. In the dioritic and granodioritic phases, a penetrative foliation is locally developed, defined by flattened quartz and mortar-textured, aligned feldspars. This foliation is apparently axial planar to tight, local folds (Plate 4-43) which probably developed during emplacement of younger portions of the batholith.

Plate 4-43: *Infolding of granodiorite (light) and diorite (dark) with strong foliation axial planar to folds, southwest corner of Black Lake; Wild Cove Pond Igneous Suite.*

The orientation of some of these sparsely distributed planar features in the Black Lake - Wild Cove Pond area defines an ellipsoidal shape, with moderate dips toward a core located to the southwest of Wild Cove Pond. Thus, the northerly portion of the pluton may represent a body distinct from the southerly portion; Kidd (1974) also made this inference. Data from more southerly portions of the batholith are too meager for resolution of any such features. The two portions of the batholith are separated by the Black Lake Fault; because of its location at the isthmus between the two portions, it seems likely to be a juncture between two separate plutons. However, the fault is somewhat enigmatic. It is defined by cataclastic rocks within a zone containing a high concentration of xenoliths, similar to the contact migmatites at the boundaries of the batholith. It is uncertain though if the fault completely traverses Black Lake, since it is unrecognized in the poorly exposed schists on the east side of the lake, and since diorite, with apparently concordant structures, occurs on both sides of the lake. Thus, either (i) the fault is a hinged fault, anchored in Black Lake, that separates two plutons to the northwest of the lake, or (ii) the fault is a through-going feature that coincidentally juxtaposes dioritic portions of the batholith. Despite this structural contrast, the two portions of the batholith are lithically indistinguishable; however, they are geochemically distinct (see Chapter VI).

External Structural Relationships

The Wild Cove Pond Igneous Suite intrudes the East Pond Metamorphic Suite and the Fleur de Lys Supergroup; it is faulted against the Advocate Complex. The suite is demonstrably postkinematic with respect to all regional deformations in the Fleur de Lys Belt, though in many places dikes emanating from the batholith are folded with the country rock (Plate 4-44). In some cases, a coarse crenulation cleavage is associated with these folds. These structures are attributed to emplacement of the batholith, as the folds are only locally developed and irregular in form, and the axial planar cleavage is only loclly developed and not regionally correlative with any of the Fleur de Lys Belt fabrics. Structures in the country rock marginal to the pluton are generally parallel to the contact with the batholith.

Plate 4-44: *Granitoid dike of the suite folded with unseparated Fleur de Lys country rock, east side of Upper Indian Pond; Wild Cove Pond Igneous Suite (photograph by J. Gagnon).*

In most places, the contact of the igneous suite with the East Pond and Fleur de Lys rocks is marked by a zone of contact migmatite that ranges in width from a few metres up to at least 400 m, as at Upper Indian Pond; some zones may be wider than suspected, but because of poor exposure and the heterogeneous nature of the zone, they would be unrecognizable. Only locally is the contact with the country rocks relatively sharp, such as at Rocky Brook near Hampden and near Kidney Pond on the Baie Verte highway.

Migmatites occur at the boundaries between the different phases of the suite and the country rocks, though most commonly at the external contacts of the granodioritic and granitic phases. A myriad of structures is displayed in these zones at outcrop scale, including agmatite, schollen (Plate 4-45), phlebitic structure (Plate 4-46), stromatic structure, folds (Plate 4-46), ptygmatic folds, schlieren (Plate 4-45), and nebulitic structures; these are displayed best at Upper Indian Pond, immediately northwest of Black Lake, and at Wild Cove Pond. The components of the agmatites as well as the independent, unoriented schollen are generally blocks of the immediate country rock. Commonly, the foliation in these rafts is parallel to their long axis, which is either at a random orientation or parallel to country rock structures. Locally, some of the well defined rafts contain kyanite (Figure 4-2), which is not characteristic of the country rock. The less well defined schlieren and nebulitic wisps appear to be mostly of metasedimentary origin, although ghostly amphibolite relicts also occur. The presence of these schlieren and more highly assimilated rocks, as well as the kyanite-bearing rafts, strongly suggests mobility along the contact zones at the time of granite emplacement. The more highly assimilated rafts and kyanite-

Plate 4-45: *Schollen and schlieren of Fleur de Lys schists in contact migmatite on the east side of Upper Indian Pond; Wild Cove Pond Igneous Suite. Note contact of granite with schist in upper left corner (photograph by J. Gagnon).*

Plate 4-46: *Phlebitic and fold structures within contact migmatite on east side of Upper Indian Pond; Wild Cove Pond Igneous Suite (photograph by J. Gagnon).*

bearing inclusions were probably derived from the country rock at levels of the batholith deeper than those presently exposed.

Locally, fine grained banded mixtures of diorite and granitic rock are associated with the contact migmatites (Plate 4-47); at Wild Cove Pond and Black Lake, they form the contact migmatites as they contain irregularly shaped pods and rafts of the country rock. These hybrid gneisses are strongly foliated and commonly fine grained, suggesting that they are protoclastic in origin.

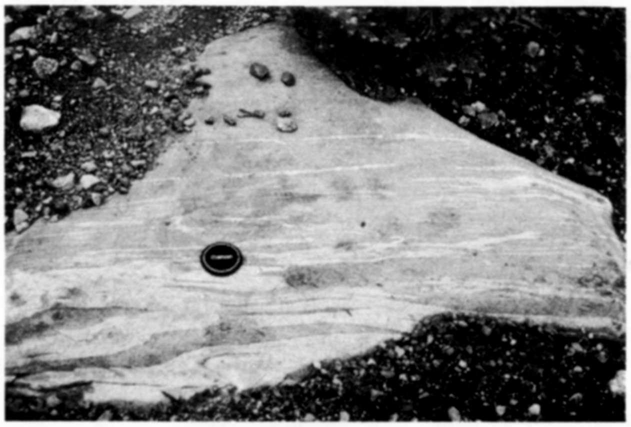


Plate 4-47: *Intermingled diorite and biotite granite form a gneiss-like zone near the northeast border of the batholith, inland northeast of Wild Cove Pond; Wild Cove Pond Igneous Suite (photograph by W. Muggridge).*

The contact between the main body of the Wild Cove Pond Igneous Suite and ultramafic rocks of the Advocate Complex along the eastern margin of the batholith is unexposed, but considered to be tectonic since most rocks near the contact are sheared. In addition, the granite is void of ultramafic inclusions in the area of the contact, thus supporting the idea of a tectonic junction between the units. A very small body of fine to medium grained granite to granitic porphyry that is probably related to the suite outcrops on Herman's Brook east of Upper Indian Pond, and is intrusive into the west side of an ultramafic body of the Advocate Complex. The contact on Herman's Brook between feldspar porphyritic felsite and the ultramafic mass is somewhat ambiguous, since it is highly sheared and altered, but flecks of serpentinite throughout the felsite indicate an intrusive relationship.

Age and Correlation

K/Ar isotopic dates of 392 ± 16 Ma for biotite from the granitic satellite near Kidney Pond (Wanless et al., 1972) and 365 Ma for biotite from the suite at the south end of Wild Cove Pond (Lowdon, 1961) indicate a Devonian age for the Wild Cove Pond Igneous Suite.

The Wild Cove Pond Igneous Suite is the largest of four known Paleozoic postkinematic granitoids that intrude the Humber Zone of Newfoundland. The suite, along with two other bodies, the Partridge Point Granite near Fleur de Lys and the Belle Island Granite (Kennedy et al., 1973) on the Grey Islands, intrudes rocks of the Fleur de Lys Belt, whereas the Gull Pond Granite (Lock, 1969a) intrudes fossiliferous Middle to Upper Silurian strata of western White Bay. The Partridge Point and Belle Island Granites, both of which are two-mica plutons, yield K/Ar biotite ages of 368 ± 16 Ma (Wanless et al., 1972) and 375 ± 16 Ma (recalculated from data reported in Kennedy et al., 1973), respectively. Thus, all of these granitoids are considered Devonian in age. Considering their unique location at the eastern edge of the Humber Zone and their common ages, these plutons are probably related to the same Acadian intrusive event.

PARTRIDGE POINT GRANITE

The Partridge Point Granite, originally termed the Partridge granite (Fuller, 1941) for exposures near Partridge Point, outcrops as a thin strip for 1.5 km along the northernmost portion of the Fleur de Lys Peninsula. It is a medium grained, garnetiferous, muscovite-bearing leucogranite. Fuller (1941) noted two variations besides the typical medium grained phase, including an aplitic phase (which occurs as rafts cut by, and included in, the medium grained granite) and a pegmatitic phase (which crosscuts the typical granite). Fuller (1941) also commented on the petrography of the typical form of the granite as follows:

> The Partridge Point granite is an even-grained, granular rock with an average grain size of about 1.1 mm. In hand specimen quartz, feldspar, muscovite, and tiny red garnets may be seen. The feldspar is usually milky white on a weathered surface; in places it shows a graphic intergrowth with quartz. Microcline is the most abundant feldspar, but albite-oligoclase, microcline-perthite and albite are also present. Of these feldspars, albite-oligoclase is the only one which shows much sericitization. Quartz with serrate borders, muscovite, chlorite, apatite, and small red garnets complete the minerals in the granite.

The granite locally encloses highly recrystallized xenoliths, up to 2 m square, of the bordering Fleur de Lys schists.

In addition to the main body of the granite, numerous aplites, pegmatites and felsic intrusive breccias occur in surrounding country rock; these rocks are probably related to the Partridge Point Granite. Sheetlike bodies of aplite and pegmatite occur in Hard Bay and Seal Cove, Baie Verte. Kennedy (1969) reported a small body of intrusive breccia between Barry's and Bishie Coves, and a second small body of breccia was found in the area southwest of Fleur de Lys Hill (C. MacKenzie, personal communication, 1980). The breccias generally contain angular fragments, up to 5 mm across, of the psammitic country rock as well as numerous cracked and fractured feldspar and quartz crystals set in a fine matrix of quartz-feldspar ± muscovite.

The granite and all of the presumably related dikes and smaller bodies are postkinematically intrusive into the Old House Cove and Rattling Brook Groups. The contacts are generally steeply dipping and sharp, and truncate all structures in the adjacent country rock.

Age and Correlation

A K/Ar isotopic age of 368 ± 16 Ma for muscovite from the Partridge Point Granite (recalculated from Wanless et al., 1972) indicates a probable Devonian age for the body. This age is in accordance with the postkinematic character of the pluton and the mid-Silurian ages (Dallmeyer, 1977) for regional metamorphic cooling of the country rocks.

The only other reported muscovite-garnet granites in the Humber Zone occur in the Wild Cove Pond Igneous Suite. The similar geologic setting and age of the Partridge Point Granite and the suite indicate that the granite may be related to the suite. A single chemical analysis of the granite, which is indistinguishable from analyses of the suite, supports this idea (see Sample 1340032, Appendix III).

GRANBY ISLAND FORMATION

The name Granby Island Formation is herein proposed for the sequence of black to gray slate, argillite and graywacke that is confined to Granby Island and Gull Rock in White Bay. The unit is homogeneous and somewhat monotonous, and lacks any continuous marker beds. It is also strongly deformed, locally exhibiting two fabrics, but only slightly recrystallized. In view of these characteristics, designation of a type section requires more detailed work; however, the section along the north coast of the island is considered here to be a reference section for the formation. The maximum outcrop width of the unit, along the north shore of Granby Island and including Gull Rock, is approximately 1 km.

Typically, the unit is composed of thin to medium bedded slate, argillite and graywacke, with minor quartzite and a few olistostromal horizons. Well defined graywacke beds range up to 1 m thick. Primary features are common in these rocks, particularly graded bedding in gritty graywackes. The graywacke beds typically grade from a gritty to pebbly base up into structureless massive argillite and slate; only locally is ripple-drift lamination evident in the finer grained fraction. Slump folds and convolute bedding were observed in graywacke beds at the south end of the island. Pebbles and grains in these grits are most commonly quartz and plagioclase, though de Wit (1972) reported blue quartz, carbonate and ultramafic detritus. Carbonate is common throughout the unit and locally forms isolated pods, stringers, or thin layers within the clastic rocks. Local buff to light green weathering of this rock type suggests a higher carbonate content than other layers in the formation. Thin (<5 cm) beds of buff weathering, fine grained quartzite outcrops on the northeastern coastline of Granby Island. A few layers of conglomeratic argillite have been noted; the most prominent of these is exposed in the southwestern corner of the island at the strand line. It is composed predominantly of chaotic black slate and argillite containing boulders and slabs more than a metre long of clastic rock indigenous to the formation. This member is more than 5 m thick. Similar though thinner members outcrop along the north coast of the island.

Systematic petrographic studies of the formation were not undertaken in this study, but a few thin sections of graywacke and argillite were briefly scanned. The sections revealed grains of mainly quartz and plagioclase, up to 2 mm in diameter, in a very fine grained matrix of muscovite $\pm$ chlorite $\pm$ biotite. The coarser grains are subangular to subrounded, but it is uncertain if the shape is of primary origin, since many grains display incipient recrystallization with sutured boundaries. Minor tourmaline was noted in one section; it occurs as irregular small masses in the matrix, indicating a possible detrital origin. Biotite occurs as large flakes that locally define the main fabric and in places appear to have overgrown the fabric.

The Granby Island Formation is substantially less deformed and metamorphosed than the Fleur de Lys Supergroup to the east. The primary features noted above, as well as the monotonous repetition of graywacke within fine grained clastics and the occurrence of local olistostromes all indicate that the formation probably represents a turbidite sequence.

The age of the formation is uncertain but is most likely older than Middle Silurian, since the rocks record a regional metamorphism similar to that of the Fleur de Lys Belt, though of a slightly lower grade. Lock (1969a) considered rocks on Granby Island to be equivalent to the base of the Lower Cambrian Coney Head Group in western White Bay. De Wit (1972) correlated the Granby Island clastics with more highly deformed and metamorphosed rocks of the Outboard sequence of the White Bay Group. R. Smyth and H. Williams (personal communications, 1979) each suggested that rocks here included in the Granby Island Formation are correlative with the Lower(?) to Middle Ordovician Goose Tickle flysch of the Hare Bay - Canada Bay region. The Goose Tickle flysch represents easterly derived flysch that marks the assembly and westerly emplacement of ophiolitic Taconic allochthons in western Newfoundland. As such, the unit is characterized and distinguished from older clastic units by detrital chromite and ultramafic clasts derived from the allochthonous ophiolites; de Wit's (1972) observation of ultramafic detritus in the Granby Island clastics strongly enhances this correlation. On the basis of these previous correlations as well as my own observations of all of the units mentioned above, I also suggest an Early(?) to Middle Ordovician age for the Granby Island Formation.

De Wit's (1972) correlation of the formation with the Outboard sequence suggests that the White Bay Group, in part, is Middle Ordovician. It should be noted, however, that even though the units are lithically similar, no ultramafic detritus has been reported from the White Bay Group.

CHAPTER V

STRATIGRAPHY OF THE BAIE VERTE BELT

The term Baie Verte Belt is informally established here to denote the assemblage of dominantly volcanic and intrusive rocks that outcrop on the east side of the Baie Verte Peninsula (Figure 1-1). These rocks form the western margin of the Appalachian Dunnage Zone (Williams, 1978a) and contrast sharply with rocks of the Fleur de Lys Belt to the west and north. The belt is distinguished from the remainder of the Dunnage Zone largely by its geographic isolation, its depth of exposure, and, in part, its Cambro-Ordovician stratigraphy. The belt is composed of three major stratigraphic elements, including (i) ophiolitic basement, (ii) volcanic cover sequences, and (iii) a variety of mafic to granitoid intrusions. In most of the ophiolitic units, the overlying cover rocks are difficult to distinguish from the upper members of the ophiolite. These cover sequences are described with their respective ophiolite. Where the volcanic cover rocks are easily distinguished from the ophiolites they are discussed under the heading "Other Volcanic Cover Sequences." In addition, a small, poorly exposed area of Carboniferous sedimentary rocks occurs in the southwest corner of the belt and is described with the cover sequences.

Rocks equivalent to those of the Baie Verte Belt also outcrop along the western portion of the Dunnage Zone at Glover Island on Grand Lake (Knapp, 1980) and in the Annieopsquotch Mountains (Herd and Dunning, 1979).

OPHIOLITIC BASEMENT

Neale (1959b) first noted the ophiolitic affinity of the mafic-ultramafic assemblages on the eastern portion of the Baie Verte Peninsula. Four geographically distinct ophiolitic assemblages are now recognized on the peninsula, including the Advocate, Point Rousse and Betts Cove Complexes, and the Pacquet Harbour Group.

The term ophiolite is defined as:

> ...a distinctive assemblage of mafic to ultramafic rocks.... In a completely developed ophiolite, the rock types occur in the following sequence, starting from the bottom and working up: ultramafic complex, consisting of variable proportions of harzburgite, lherzolite, and dunite, usually with a metamorphic tectonite fabric (more or less serpentinized), gabbroic complex ordinarily with cumulus textures, commonly containing cumulus peridotites and pyroxenites and usually less deformed than the ultramafic complex, mafic sheeted dike complex, and mafic volcanic complex, usually pillowed. Associated rock types include (1) an overlying sedimentary section typically including ribbon cherts, thin shale interbeds, and minor limestone, (2) podiform bodies of chromite generally associated with dunite, and (3) sodic felsic intrusive and extrusive rocks. Faulted contacts between mappable units are common. Whole sections may be missing. An ophiolite may be incomplete, dismembered or metamorphosed, in which case it should be called a partial, dismembered, or metamorphosed ophiolite.
>
> *Geotimes, Volume 17, Number 12, page 25*

According to this definition, the ophiolitic units of the Baie Verte Belt qualify as either partial or dismembered; the Advocate Complex is highly dismembered and appears to be incomplete, lacking a definite ophiolitic pillow lava member, the Point Rousse Complex is dismembered but complete, the Betts Cove Complex is intact but appears to be incomplete, since a noncumulate ultramafic member has not been recognized at its base, and the ophiolitic portion of the Pacquet Harbour Group is incomplete, consisting of only a mafic volcanic member.

Ophiolitic parts of the Advocate and Point Rousse Complexes and the Pacquet Harbour Group are structurally and stratigraphically intermingled with younger sequences of dominantly mafic volcanic and volcaniclastic rocks. Thus, in these units, the cover rocks are described with the ophiolitic assemblages in this section. The Betts Cove ophiolite forms a mappable unit (Figure 1-1), distinct from its cover, the Snooks Arm Group, which is described in the succeeding section of this report with other volcanic cover sequences.

Nonophiolitic components of the Advocate Complex are distinct from those of the other complexes, and appear to unconformably overlie the ophiolitic elements (Bursnall, 1975). The volcanic-volcaniclastic sequences above the Point Rousse ophiolite, the Pacquet Harbour Group and the Betts Cove Complex conformably overlie the oceanic basement and are considered to be mutually correlative.

Since 1971, all workers have considered the ophiolitic assemblages to be remnants of ancient oceanic crust (e.g. Church and Stevens, 1971; Dewey and Bird, 1971; Bird et al., 1971; Kennedy and Phillips, 1971; Upadhyay et al., 1971). However, their age has been a major topic of controversy among workers. Originally, the assemblages were viewed as representing structurally isolated remnants, along with allochthonous ophiolites of the Humber Zone, of a once continuous ophiolite sheet related to one cycle of ocean floor generation (Stevens et al., 1969; Church and Stevens, 1971). Subsequently, some workers viewed the ophiolite complexes as individual occurrences, of at least two distinct generations, that formed in separate marginal basins (Dewey and Bird, 1971; Kennedy and Phillips, 1971; Kennedy, 1973, 1975a; Kidd, 1974, 1977; Kidd et al., 1978). This interpretation was based on different structural relationships between the ophiolites and metaclastics of the archaic eastern and western divisions of the Fleur de Lys terrane (Church, 1969). Recent studies by Bursnall and de Wit (1975), DeGrace et al. (1976), Williams et al. (1977) and Hibbard (1982) have demonstrated that the concept of an eastern division of the Fleur de Lys terrane is largely invalid and that deformation of the ophiolitic rocks is inhomogeneous (see Chapter VII). Therefore, these workers have revived the interpretation of a single generation of oceanic crust for ophiolites of the Baie Verte Peninsula and the Humber Zone. This interpretation is fully supported by the stratigraphic, geochemical and structural data of this report.

The best age constraints on the ophiolitic sheet are from the Betts Cove Complex and from allochthonous Humber Zone ophiolites to the west of the peninsula. A sample of Betts Cove gabbro yielded a U/Pb zircon age of $488.6\ ^{+3.1}_{-1.2}$ Ma (Dunning and Krogh, 1983); the complex is conformably overlain by fossiliferous Arenig strata of the Snooks Arm Group (Snelgrove, 1931). Mattinson (1975) erroneously interpreted a 463 ± 6 Ma U/Pb zircon age from a plagiogranite dike in the complex as the age of formation of the suite; Church (1977) subsequently reported that the rock dated is from a dike of the Burlington Granodiorite. Therefore, the Betts Cove Complex is Early Ordovician in age.

Dating of the other ophiolites is less direct and depends, in part, upon correlation with ophiolites outside the map area. Numerous isotopic dates on various portions of the Humber Zone ophiolites (Mattinson, 1975, 1976; Jacobsen and Wasserberg, 1979) indicate a latest Cambrian age of crystallization for these rocks. The ophiolitic rocks between those of the Humber Zone and the Betts Cove Complex (including the Advocate and Point Rousse Complexes and the Pacquet Harbour Group) are undated. However, the Ming ore body in the Pacquet Harbour Group is dated by Pb isotope methods at 460 Ma (Sangster and Thorpe, 1975) and the group is intruded by the Burlington Granodiorite, which is approximately 460 Ma old (see Burlington Granodiorite); thus, the Pacquet Harbour Group is probably Early Ordovician or older. These age constraints indicate that the ophiolite sheet was Late Cambrian to Early Ordovician in age, and appears to be younger to the east. This suggests that the undated ophiolites are probably slightly older than the Betts Cove Complex, but younger than transported ophiolites of the Humber Zone.

It should be stressed here that, although the ophiolites of the peninsula are all considered to be of the same generation, they exhibit variable internal relationships and different cover sequences; these differences are all considered compatible with expected inhomogeneities in a single large slab of oceanic crust and are summarized at the end of this section.

The following descriptions of the ophiolitic complexes on the peninsula draw heavily upon the detailed work of Kidd (1974) and Bursnall (1975) for the Advocate Complex, Norman (1973), Norman and Strong (1975) and Kidd et al. (1978) for the Point Rousse Complex, and Upadhyay (1973), Schroeter (1971), Riccio (1972) and Coish (1977b) for the Betts Cove Complex. Data from the present study augment and tie together these detailed investigations.

ADVOCATE COMPLEX

Definition and Extent

The Advocate Complex is defined as the steeply dipping, northeasterly striking assemblage of intensely dismembered and deformed mafic and ultramafic plutonic rocks, mafic volcanic and volcaniclastic rocks, and dark gray to black slates that outcrop near the town of Baie Verte. Mylonitic rocks are common. Most of the unit extends from Marble Cove southward to the bottom of Baie Verte harbor; plutonic components of the complex form a discontinuous line south to Birchy Lake, approximately 70 km from Baie Verte, that serves to emphasize a major regional structure, the Baie Verte Line (Figure 1-1; also see Chapter VII). The thickness of the complex is inestimable due to intense deformation, though its maximum outcrop width is nearly 4 km in the area east of Advocate Mines. Rocks of the complex are well exposed on the lower reaches of Rattling Brook, at Advocate Mines, and on the coast from Shark Point to Marble Cove; these areas are here considered as reference sections for the complex. In addition, large xenoliths of dominantly ultramafic rocks in the Wild Cove Pond Igneous Suite are included in the Advocate Complex due to geographic proximity and broad lithologic similarity. The Advocate Complex is in tectonic contact with all bordering units.

The complex is a stratigraphic enigma; it comprises two broad divisions, including the ophiolitic plutonic rocks and a sequence of dominantly volcanic and volcaniclastic rocks. These components are extremely disrupted and structurally intertwined at different scales. In some places, such as west of Baie Verte townsite, tectonic dissection and interleaving is at outcrop scale, and the geology is a chaotic mosaic of ophiolitic and cover rocks, in which it appears impossible to determine original relationships. Elsewhere, such as west of Duck Island Cove, tectonic disruption appears to be on a broader, mappable scale and is less intense; here, apparently larger tectonic blocks are preserved in which original stratigraphic relationships are better preserved. It should be stressed that, even in these more intact sections, small scale structural interleaving of units is present. This tectonic confusion is further compounded by the presence of sedimentary megabreccias containing map scale blocks in a scant slaty matrix; hence, in some areas it is uncertain if the apparent chaos is tectonic or depositional. In addition, the ophiolitic rocks display intense alteration in many places, and metamorphic grade and intensity of deformation increase toward the northwest of the complex. The unit as a whole thus qualifies as a tectonic complex.

The complex can be viewed, on a broad scale, as three imbricate, southwest facing, incomplete and dismembered ophiolite sheets with their presumed cover sequences. To facilitate the following discussion of rock types in the complex, these sheets are informally referred to, from northwest to southeast, as the Marble Cove (Bursnall, 1975), Duck Island Cove, and Sisters Cove sequences (Figure 5-1). The Marble Cove sequence is more highly tectonized than the more southeasterly sequences. The relationship between these sequences and the line of predominantly ultramafic rock south of Baie Verte is uncertain due to complex faulting, though it appears from the map pattern (Figure 1-1) that these rocks are most likely an extension of the Sisters Cove sequence.

Mafic rocks in the vicinity of the Baie Verte townsite, herein considered as part of the Advocate Complex, were originally assigned to the Baie Verte formation by Watson (1947) and later the Baie Verte Group by Baird (1951). Mafic and ultramafic plutonic rocks that outcrop along the Baie Verte Line (Figure 1-1) were considered by early workers as intrusions into the Baie Verte Group; they were first included in the group by Kidd (1974). Kennedy (1975a) noted the structural complexity of the rocks in the Baie Verte area and referred to them as the Advocate Sequence, a division of the Fleur de Lys Supergroup; he also included the mafic-ultramafic plutonic rocks along the Baie Verte Line in the sequence. Bursnall (1975) informally assigned the altered gabbro and intensely deformed greenschist immediately north

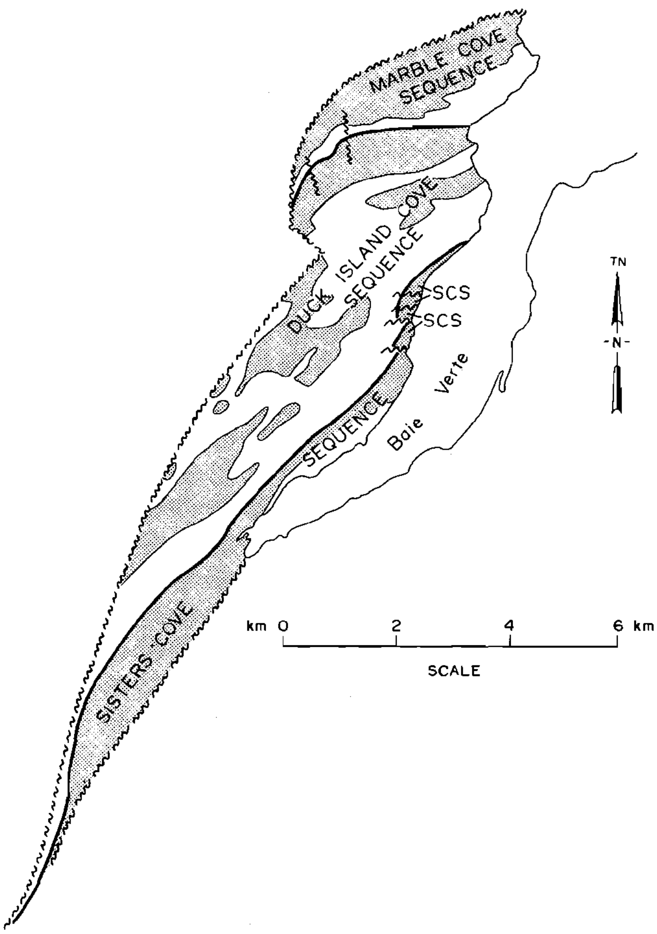


Figure 5-1: *Constituent sequences of the Advocate Complex; shaded areas represent ophiolitic plutonic rocks.*

of Duck Island Cove and Advocate Mine to the Marble Cove and Lodabatts formations, respectively. He considered the ultramafic and gabbroic bodies that occur between these units and Upper Duck Island Cove Brook as the Advocate Complex, and the volcanic and sedimentary rocks immediately to the southeast as the Shark Point Group. Williams et al. (1977) informally relegated the rocks outlined by Kennedy (1975a) as the Advocate Sequence, to the Advocate Complex, but, unlike Kennedy, they excluded the rocks immediately north of Lower Duck Island Cove from the unit and included them in the Birchy Complex.

Ophiolitic Plutonic Rocks

The ophiolitic plutonic rocks of the complex include serpentinized ultramafic rock, gabbro and metagabbro, sheeted diabase dikes, and altered and metamorphosed equivalents of these rocks. In addition, a single exposure of garnetiferous amphibolite is closely associated with ultramafic rocks at Advocate mines[1]; it has significant bearing on the interpretation of the history of the ophiolitic rocks and will be described with them.

SERPENTINIZED ULTRAMAFIC ROCK

Description: Serpentinite outcrops within each of the three ophiolitic sequences in the Advocate Complex, but is most common in the Duck Island Cove sequence; in addition, the largest ultramafic bodies form a discontinuous line; the Birchy Lake body, is the largest in the complex and reaches a maximum outcrop width of 3 km northeast of Gillard Lake. The large bodies adjacent to Micmac Lake and Flat Water Pond are herein called the Micmac and Flat Water plutons or bodies, respectively. The northernmost one is informally termed the Advocate pluton or body, after the mine it hosts. Most of the following descriptions pertain to the larger ultramafic bodies in the Duck Island Cove sequence.

Serpentinization of the ultramafic rocks is ubiquitous; only rarely are original mineral assemblages preserved (Kidd, 1974; Bursnall, 1975). The extent of serpentinization and other subordinate alterations, such as steatitization and carbonatization, is usually reflected in the weathering of these rocks. In cases of extreme alteration, the rocks are cream to white weathering, whereas the fresher rocks are characteristically orange to rusty weathering. All of the rocks range from blue-black and green-black to very dark gray on fresh surfaces. The alteration of these rocks will be considered following a general description of their field appearance and the petrography of fresher specimens.

The original plutonic textures of these rocks are present only in the interior of larger bodies, where bastites (antigorite pseudomorphs of pyroxene phenocrysts) up to 1 cm in long dimension are preserved (de Wit, 1972; Kidd, 1974; Bursnall, 1975). The bastites and, in rare cases, pyroxenes are generally more resistant to weathering than the surrounding, generally finer grained serpentine-olivine matrix and form a rough knobby weathered surface where present. This characteristic weathering pattern is useful in distinguishing pyroxene-bearing peridotites from generally smooth weathering equigranular dunite in the field. On the basis of this weathering pattern as well as thin section studies, Czamanske (1956), Kidd (1974), and Bursnall (1975) showed that most of the Advocate Complex ultramafic rocks were originally harzburgite, with minor dunite and rare lherzolite. In addition, coarse pyroxenite and chromitite have been recorded in the Advocate Complex ultramafics (Watson, 1942; Neale, 1959a; Kidd, 1974; Bursnall, 1975). Field evidence and thin section studies (Kidd, 1974; Bursnall, 1975) indicate that both noncumulate and cumulate ultramafic rocks are present, though the textures of many ultramafics in the complex cannot be determined due to extensive serpentinization.

Harzburgite forms most of the ultramafic rocks in the Advocate Complex, and generally, where serpentinization is not extensive, appears to be noncumulate in origin (Kidd, 1974; Bursnall, 1975). Czamanske (1956) noted that most of the Birchy Lake body is composed of massive, featureless harzburgite. An unusual occurrence of pegmatitic harzburgite with bastite crystals up to 5 cm long was reported from the area approximately 1 km south of the Flat Water Pond outflow (Kidd, 1974).

Kidd (1974) estimated that dunite forms approximately 5% of the Flat Water and Micmac bodies; it is typically inter-

[1] Advocate Mines Ltd. ceased production in 1981; mining was resumed in September, 1982, by a new company, Baie Verte Mines Inc..

layered on a centimetre to metre scale with the harzburgite. Rarely, dunite comprises the bulk of some larger bodies; Bursnall (1975) inferred that the extensively serpentinized Advocate body is dominantly dunitic, based on the meager amount of recognizable bastite. Locally in this body, vague zones of concentrated bastites suggest primary dunitic-harzburgitic layering (Bursnall, 1975). Kidd (1974) reported that, at the southeastern margin of the Flat Water pluton, interlayered noncumulate harzburgite and dunite contain numerous bands and veins of pegmatitic orthopyroxenite up to 10 cm wide; he inferred that this sequence represents part of the transition zone between noncumulate ultramafic rocks and cumulate ultramafic-gabbroic rocks because of the abundance of pyroxenitic pegmatites.

Pyroxenite, lherzolite and chromitite form a very minor part of the ultramafic rocks. In addition to the occurrence mentioned above, pyroxenite forms (i) large lensoidal areas in the serpentinite body at the base of the Marble Cove sequence (Bursnall, 1975), (ii) pods in serpentinite near the southern edge of the Advocate body (Bursnall, 1975), (iii) discrete areas in the ultramafic rocks northwest of Baie Verte townsite (Watson, 1942), and (iv) tectonic slivers along a fault zone, approximately 0.65 km north of the Burlington road - Baie Verte highway junction (Kidd, 1974). Lherzolite is rare in the complex; the only occurrence reported in outcrop is in the southeastern portion of the Flat Water pluton, where Kidd (1974) noted a vague xenolith of layered cumulate lherzolite and harzburgite within noncumulate harzburgite. Neale (1958a) reported a lens of chromitite approximately 1 by 3 m in size at the northern end of the Flat Water ultramafic body; Kidd (1974) reported thin chromitite seams, up to 2 cm wide, on Middle Arm Brook and along the Baie Verte highway approximately 1 km north of the Burlington road junction.

In addition to the primary layering, the ultramafics locally display fragmental textures. Bursnall (1975) recognized extensive brecciation of serpentinite from the northwestern portion of the Advocate ultramafic body; he noted that dark greenish gray serpentinite is veined and brecciated by a dark blue-gray variety. An unusual development of fragmental serpentinite occurs in an ultramafic block within slate on Baie Verte River at the abandoned Terra Nova Mine site. Here, well rounded, greenish black, egg-shaped serpentinite clasts up to 3 cm long occur in a greenish gray serpentinite matrix. Kidd (1974) reported a similar texture from the northern end of the Micmac ultramafic body. Here, the clasts range up to 20 cm across, and form an area approximately 80 m in diameter. Two dikes of similar texture radiate to the north from the main brecciated area. Kidd (1974) attributed this textural feature to gas brecciation and interpreted the Micmac breccia as an ultramafic tuffisite pipe.

A foliation is locally developed in the ultramafic rocks, particularly in the harzburgites. Based on petrographic evidence, Kidd (1974) interpreted this fabric as a high temperature foliation that predates any regional deformation in the complex.

Veins are common in the ultramafics of the complex and are variable in composition. They are generally composed of one or more of the following: picrolite, chrysotile, talc-carbonate, magnetite and nephrite (Kidd, 1974; Bursnall, 1975; J. Young, personal communication, 1978).

Petrography: Detailed thin section studies of the ultramafic rocks have been undertaken by numerous workers. Kidd (1974) summarized the petrography of the better preserved harzburgites and dunites in the Micmac and Flat Water bodies as follows:

> Harzburgite has anhedral forsterite of the same composition range as dunite (Fo_{85-95}), with 5 to 30% anhedral enstatite (about En_{90}) in a xenomorphic granular aggregate. The orthopyroxene is serpentinized in all outcrops except one and is termed bastite hereafter. Accessory anhedral chromite grains usually with intricate anticuspate margins are ubiquitous, but they never occur in seams or bands in the harzburgite. Olivine grain size is between 0.1 and 10 mm, commonly 1 to 4 mm; enstatite (bastite) between 0.1 and 10 mm, commonly 1 to 3 mm; chromite 0.1 to 2 mm, commonly 0.5 mm.
>
> Dunite consists of a xenomorphic granular aggregate of forsterite (Fo_{85-95}, probably mainly about Fo_{90}), 0.1 to 4 mm across, with accessory, sub to euhedral chromite grains 0.05 to 2 mm, commonly 0.5 mm across. Chromite grains often occur in seams one grain thick within and parallel to the dunite band.

Kidd (1974) and Bursnall (1975) described the clinopyroxenites in the complex; Kidd noted cumulus pyroxene, possibly augite, within the tectonic slivers along the Baie Verte highway, whereas Bursnall (1975) reported that the pyroxenite in the Advocate body is nearly 100% diallagic and probably cumulate in origin.

In the cumulate lherzolite in the Flat Water body, Kidd (1974) indicated that:

> ...the cumulus phases are olivine and (now serpentinised) orthopyroxene in the ratio about 4:1, with an accessory opaque ore mineral, possibly altered chromite. Very rare small clinopyroxene crystals are enclosed in cumulus orthopyroxene. Postcumulus processes were growth of poikilitic chrome diopside, probable slight resorption of orthopyroxene, and slight resorption of olivine where in contact with diopside, but slight overgrowth elsewhere. Olivine grains are 1 to 12 mm, usually 6-8 mm across, orthopyroxene are 0.4 to 4 mm, usually 1 to 2 mm across, and the chromite (?) 0.1 to 3 mm, typically 0.2 to 0.5 mm across. Olivine composition is in the range Fo_{80-90}, probably about Fo_{85}.

Alteration: Serpentinization is ubiquitous in ultramafic rocks of the complex and varies in extent in individual bodies; Czamanske (1956) reported 60 to 90% serpentinization in the Birchy Lake body, Kidd (1974) estimated 10 to 100% serpentinization for the Micmac and Flat Water bodies, though he noted that only in a few exposures is the alteration less than 50%, and Bursnall (1975) indicated that the Advocate body is almost totally serpentinized. Kidd (1974) reported that 95%+ serpentinized ultramafic rock occurs only at the margins of the Flat Water and Micmac bodies and that the width of intense serpentinization is greater on the east side of these bodies than on the west side. At least two generations of serpentinization are evident in some bodies; Bursnall (1975) presented evidence for this in the brecciated serpentinites of the Advocate body, wherein an early phase of serpentinized ultramafic forms blocks in a distinctly younger serpentinized matrix.

The mineralogy of the serpentinites displays a slight variation between individual bodies. Czamanske (1956) reported that the main constituent of the Birchy Lake body is lizardite with minor chrysotile, based on X-ray diffraction studies, whereas Kidd (1974) and Bursnall (1975) reported antigorite as the main constituent of larger bodies to the north. Kidd (1974) described the general petrography of the serpentinites:

> ...the serpentinite replacing olivine is almost always a randomly oriented rather coarse-grained form-antigorite... Two samples out of about

fifty sectioned show mesh-serpentine. Enstatite is always replaced by an ultrafine-grained fibrous serpentine, where the fibres are coaxial with the intersection of the two pyroxene cleavages... The initial replacement is a nephritic mineral, brown in transmitted light and a characteristic porcellanous green in white weathering rocks in outcrop. This is then replaced in the most serpentinised rocks by a colourless fibrous serpentine mineral with normal gray birefringence. Very fine grained magnetite dust is a ubiquitous product of serpentinisation, and tends to be concentrated along some of the original olivine grain boundaries, and in the bastites. Scattered crystals of Mg-rich carbonate are found in some thin sections of the highly serpentinised white-weathering rocks, and a small amount of brucite has also been identified in some samples. Talc has not been seen in the little to undeformed serpentinised ultramafic rocks.

In addition to serpentinization, portions of the ultramafic rocks have been carbonatized, steatitized, or chloritized. The most common alteration of serpentinite in the area is a carbonate-quartz-fuchsite rock, locally termed virginite (original term by Norman Peters, local prospector). The rock is typically streaked green and white on fresh surfaces, but otherwise weathers a distinct rusty color. It occurs along fault zones within ultramafic bodies of the Advocate Complex and in tectonic slivers along the Baie Verte Line. The virginite is best exposed along the Baie Verte highway at the outlet of Flat Water Pond. It has been attributed to metasomatism of serpentinite by the addition of CO_2 (Kidd, 1974) (see Virginite fault, Chapter VII).

Czamanske (1956) studied carbonate alteration in the Birchy Lake body. Here, all stages of the process are present, from 5% carbonatization to almost total carbonate rock. The most common carbonate mineral that he reported from this area is breunnerite, with subordinate magnesite and dolomite. Apparently, the carbonate preferentially replaces bastite. Czamanske (1956) also reported that talc and chlorite replace bastite in the Birchy Lake body; neither alteration mineral forms more than 2% of the rock. Farquhar (1959) reported an extensive zone of talc-carbonate alteration along the eastern margins of the ultramafic bodies in the Black Brook area.

GABBRO

Gabbro[1] is a major component of each of the sequences in the Advocate Complex and clearly defines the trend of each sequence (Figure 1-1). Southward from Baie Verte, along the Baie Verte Line, gabbro is not abundant; the largest area is associated with the Flat Water ultramafic body. The largest gabbro body in the complex is the tabular sheet, approximately 1 by 6 km, that comprises most of the Sisters Cove sequence. Commonly, gabbroic rocks are closely associated with ultramafic rocks of the complex and, in many places, are interlayered and comingled with them. Gabbro and metagabbro of the Duck Island Cove and Sisters Cove sequences, as well as gabbroic rocks at Flat Water Pond, are very similar in character and are described together below; locally, small quartz albitite bodies and trondhjemite stocks are associated with these rocks and are described with them. More highly tectonized, distinctive calc-silicate schists of the Marble Cove sequence display many textural similarities with other gabbros of the complex, and are considered equivalent to them (Watson, 1942; Bursnall, 1975).

Gabbro and metagabbro of the complex typically show primary igneous textures; Bursnall (1975) summarized the overall aspect of these rocks as follows:

Textural variation is large, varying from very fine-grained highly altered porcellanous types to a coarse-grained pegmatitic facies. Equally, compositional variation is large, falling between two extremes of 'mafic-gabbro', where mafic minerals (including serpentine) are dominant and enclose isolated altered plagioclase, to 'leuco-gabbro' where mafic minerals are subdominant and may even be megascopically absent over large areas. Undoubtedly much of the localised variation is due to the pervasive rodingitisation affecting much of the assemblage but the gross, large-scale, textural and compositional variation is thought to be of primary igneous origin.

Leucogabbro is the most common variety in the Advocate Complex, and is characteristic of the ophiolites in the map area. It comprises most of the gabbros in the Duck Island Cove sequence and a large portion of the eastern half of the Sisters Cove gabbro body. These rocks were described by Bursnall (1975) as follows:

In the field the rock varies from mesocratic to leucocratic with isolated light green mafic grains (pyroxene and/or amphibole) held in a whitish-cream base. The per cent density distribution of these elements varies from approximately 50:50 (particularly at the lower levels) to 25:75 respectively.... Pyroxene and/or amphibole grain size varies from a few millimetres up to 1 cm, frequently over a short distance, and occasionally may reach as much as 10 cm in maximum dimension in small (<1 m across) pegmatitic zones. Texturally the interstitial pyroxenes are commonly anhedral and have rounded grain boundaries against the enclosing altered plagioclase.

In thin-section plagioclase is normally highly saussuritised but may occasionally exhibit palimpsest polysynthetic twinning when replacement is incomplete.... The composition of plagioclases..., based on optical tests, falls within the andesine field. Relict pyroxene (a diopsidic augite) is normally rimmed with acicular tremolite/actinolite and the dense, altered, plagioclase areas are frequently rimmed and cut by thin zones of pellucid albite. Leucoxene occurs as a common accessory and small, fine-grained areas of polygonal granoblastic secondary albite and quartz are almost invariably present.

In addition, Bursnall (1975) noted that alteration of pyroxene and plagioclase in the gabbro associated with the ultramafic body at Advocate mines (the Advocate body) increases northward, toward the ultramafic body. He also reported zoisite, prehnite and carbonate as very common alteration phases throughout the leucogabbro. Brecciation is common within the gabbro associated with the Advocate body. The breccias consist of angular to subrounded gabbroic blocks ranging from metres to centimetres in diameter set in a fine grained anastomosing matrix that is either cream or brownish black in color (Bursnall, 1975).

In many places throughout the complex, the gabbro displays primary layering that is defined mainly by the concentration of mafic minerals; both ultramafic and anorthositic rocks are interlayered with the gabbro. Leucogabbro is interlayered with serpentinized ultramafic rock and mafic gabbro in the gabbroic rocks associated with the Advocate body; leucogabbro is also common along the western portion of the Sisters Cove sequence gabbro body. In the latter, large rounded inclusions (up to 5 m in diameter) of ultramafic rock occur in the layered sequence east of the Baie Verte highway - La Scie highway junction. Bursnall (1975) reported a transi-

[1] Gabbro is used in broad sense for a wide range of rocks that are interpreted as having been originally gabbro.

tional contact between massive leucogabbro and the layered gabbroic sequence in the Advocate Mines area. Bands of anorthosite occur within deformed gabbro on the northern side of the Flat Water Pond outlet; petrographic studies by Kidd (1974) indicate that these layers are cumulate in origin. In addition, he recognized massive, two-pyroxene gabbro southeast of the Flat Water ultramafic body (Figure 1-1) that displays cumulate textures. He noted that in thin section these rocks

> ...have somewhat variable proportions of both pyroxene and plagioclase as cumulus phases, with a very minor amount of cumulus olivine in addition in one specimen. Post-cumulus processes appear to involve overgrowth of clinopyroxene and especially orthopyroxene with either minor overgrowth or resorption of plagioclase.

Locally, the gabbros of the complex exhibit a foliation that ranges from a weak fabric over extensive areas to a more intense, zonal fabric and isolated, thin mylonitic zones. Detailed investigations (Kidd, 1974; Bursnall, 1975) demonstrated that these are high temperature fabrics that probably formed during ophiolite genesis and are not related to later regional deformation. The most compelling evidence presented by these workers for this interpretation is that, locally, the foliated gabbros are intruded by genetically related, isotropic diabase dikes and gabbro (Kidd, 1974; Bursnall, 1975). Where the foliation is seen with layering, it is parallel to the layering. The foliated gabbros appear to grade transitionally into isotropic gabbro (Kidd, 1974).

Locally within the Duck Island Cove and Sisters Cove sequences, the gabbros are extensively altered to rodingite (rock produced by metasomatism involving desilication and calcium addition; Coleman, 1977). The rodingite is all associated with serpentinite, and is most extensive near the Advocate ultramafic body. Bursnall (1975) described the rocks there as follows:

> Petrographically, the most common lithology is a prehnite-zoisite-grossular-diopside rock with very variable texture from fine-grained saccharoidal (almost aphanitic and occasionally porcellanous) to medium-grained 'porphyritic' types where diffusely bounded emerald green diopsides are set in a fine- to medium-grained creamish-white base.

Rodingite dikes also occur in the ultramafic body there (Bursnall, 1975). One dike near the southeastern boundary of the body extends for the length of the asbestos mine pit; the thickness of this dike is variable due to intense boudinage, but ranges up to 3 m. Complex veining patterns occur between the rodingites and serpentinites; Bursnall (1975) reported that thin veins of rodingite cut serpentinite and are, in turn, backveined by serpentinite. This supports evidence presented above for a multistage serpentinization history of the ultramafic bodies in the complex.

Kidd (1974) and Bursnall (1975) reported that trondhjemite and quartz albitite are locally associated with the gabbro of the complex. They form minor bodies within the gabbro, ranging from centimetre scale dikes to small stocks up to 10 m in diameter. These rocks are dominantly composed of medium grained albite with subordinate quartz and minor magnetite, sphene, zoisite and amphibole.

Gabbroic rocks that form the northerly portion of the Marble Cove sequence are more highly tectonized than those of the complex to the southeast and have been transformed into distinctive cream to pale green leucocratic schists (Watson, 1942; Bursnall, 1975). Bursnall (1975) summarized the salient features of this unit as follows:

> These texturally variable leucocratic rocks may be loosely termed 'calc-silicate' schists being essentially a zoisite-quartz-albite schistose rock with variable amounts of emerald-green fuchsite, clinozoisite and tremolite-actinolite. Calcite, magnesite, chlorite, epidote, sphene and, rarely, chromite and magnesite, all occur in variable amounts. Zoisite content varies from about 20% to as much as 90% in the mode. Microscopically, the proportion of constituent phases varies considerably but in the field the lithology appears compositionally monotonous over large areas.

Locally, the schists grade into massive metagabbroic rocks that are texturally identical to the rodingitized gabbro of the Duck Island Cove sequence. A single geochemical analysis of these schists (Watson, 1942) indicates that they lost silica and were enriched in calcium; hence, these rocks were probably rodingitized gabbro before being overprinted by intense regional tectonism. Where closely associated with ultramafic rocks, the Marble Cove schists contain fuchsite. Bursnall (personal communications, 1980) noted that the western portion of these schists is mylonitic.

SHEETED DIABASE DIKES

In the Sisters Cove sequence, sheeted diabase dikes outcrop in a thin belt between Upper Sisters Cove and Baie Verte community; in addition, Kidd (1974) and Bursnall (1975) identified local diabase dikes associated with gabbro of the complex. Those in the Sisters Cove sequence are sheared to massive, fine grained, gray-green dikes that comprise nearly 100% of the outcrop. In places, plagioclase phenocrysts up to 1 cm long are present; chilled dike margins are the only other observable field features. Locally, thin metagabbro screens up to 1 by 2 m are entrapped between dikes.

Bursnall (1975) reported diabase dikes that cut the Advocate leucogabbro body and noted a zone of brecciated dark green mafic schist along the southern boundary of this body. Brecciation is a common feature of the sheeted dike zone in other ophiolites (Williams and Malpas, 1972); considered in conjunction with the stratigraphic position of these schists, this implies that the mafic schist zone represents deformed diabase dikes. Because of the uncertainty of this assignment, these rocks are left unseparated from the cover rocks of the complex in Figure 1-1. Kidd (1974) also noted numerous diabase dikes (20 to 40 cm wide) cutting gabbro along the Baie Verte Line.

Sheeted diabase dikes may be more common in the inland portions of the complex than is now realized, but the paucity and general poor quality of outcrop prevents their identification.

GARNET AMPHIBOLITE

A very small tectonic sliver of garnet amphibolite, closely associated with ultramafic rocks of the complex, was first recognized by L. Riccio and W.R. Church in the southwestern corner of the Advocate open pit, close to where the Marble Cove sequence ophiolite and cover rocks meet the Advocate ultramafic body. The zone, which was recently removed by mining operations, was reported to be less then 30 m wide and consisted of "foliated mafic schists that exhibit a metamorphic zonation away from the ultramafic" (Bursnall, 1975). The zone was described by Bursnall (1975) as follows:

> Light pinkish-red anhedral garnet porphyroblasts, less than 5 mm in maximum dimension and commonly densely clustered and/or fragmented

into granular trains along the schistosity plane, occur in a medium- to fine-grained dark grey to bluish black amphibolite matrix. A thin, < 1 cm, compositional banding is sometimes present.

In thin-section the colourless garnets are much fractured and weakly deflect the main schistosity... Two distinctive amphiboles occur: light reddish-brown to colourless pleochroic hornblende with $\gamma > \beta > \alpha$ occurs in indistinct 5 mm folias where a preferred dimensional elongation is sometimes well-developed; light yellowish to bluish green amphibole sometimes rims the brown hornblende but more commonly occurs in sporadic concentrations throughout the rock. Sphene is a common accessory (sometimes >0.5 cm) but plagioclase is rare, occurring in small lenticular very fine-grained aggregates of anhedral lath-like grains.

In close association and apparently interbanded with this garnet-bearing amphibolite are pyroxene-amphibolites. A compositional banding between pyroxene (a colourless diopsidic augite) and reddish-brown hornblende is common in thin section. Green pleochroic amphiboles are slightly less common than in the garnet-amphibolite and sphene is again an ubiquitous accessory. Plagioclase and quartz are again rare or absent over large areas of a thin-section. Occasionally clear euhedral clinopyroxenes are enclosed within anhedral amphibole. Grain boundaries are frequently straight and polygonal grains are therefore common: elsewhere diffuse and ragged boundaries between amphiboles obtain — particularly within zones containing a strong planar fabric, and within augen around porphyroblasts.

CONTACT RELATIONSHIPS

From the foregoing descriptions, it is apparent that the contacts between the ultramafic rocks, gabbros, and sheeted dikes were all originally gradational. The contact between noncumulate ultramafic rocks and cumulate ultramafic-gabbroic rocks has been interpreted as transitional in the Flat Water body (Kidd, 1974). Likewise, in many places, layered gabbro appears to grade into leucogabbro, but mafic gabbro is cut locally by leucogabbro (Bursnall, 1975). The best examples of sheeted dikes are tectonically isolated from other ophiolitic rocks (Figure 1-1), but the occurrence of numerous dikes rooted in the southern portions of the Advocate body suggests that the dikes originated in gabbroic portions of the complex.

The pattern of the patchlike, incomplete ophiolite stratigraphy preserved in the complex strongly implies that ultramafic rocks form the base of the three sequences and are succeeded stratigraphically upward, and to the southeast, by gabbroic rocks and sheeted dikes. Subsequent to its generation, the ophiolite was subjected to various alterations, including serpentinization, steatitization, carbonatization, and rodingitization, all of which display polyphase histories.

The garnet amphibolite at the apparent base of the Duck Island Cove sequence has been interpreted as a metamorphic aureole indicating high temperature tectonic emplacement of the ophiolitic rocks, probably close to their present position. Thus, the Advocate Complex is considered as allochthonous. According to Bursnall (1975), the emplacement of the Advocate Complex along the major fault containing the garnet amphibolite sliver occurred very early in the structural history of the complex (see Chapter VII).

Advocate Cover Sequence

Mafic volcanic and sedimentary cover assemblages lie to the southeast of the ophiolitic plutonic rocks in each of the three sequences. Volcanics and sediments of the Sisters Cove sequence are poorly exposed, but are similar to volcaniclastics of the Duck Island Cove sequence. A stratigraphy for the cover rocks is recognizable only in the northern portion of the Duck Island Cove sequence. Also, within this sequence, one distinctive conglomerate member (O€ab) can be separated, locally, from the cover sequence at the present map scale (Figure 1-1); other similar members are preserved throughout the sequence, but are apparently limited in extent. Cover rocks in the Marble Cove sequence (O€am) are distinct from those of the other sequences (Figure 1-1) in that they are intensely tectonized and are best described as greenschists; there is no obvious stratigraphy preserved in this sequence.

Detailed work by Bursnall (1975) in the area immediately to the southeast of Advocate Mines indicates that, locally, a tentative stratigraphy can be inferred for the volcanics and sediments of the Duck Island Cove sequence. Bursnall's stratigraphic analysis indicates that the rocks between Lower Duck Island Cove and Big Head are disposed in a southerly plunging syncline; the lower portions of this cover sequence lie immediately to the southeast of Advocate Mines, whereas the upper portions are best exposed at the coast, between Shark Point and Big Head.

Most of the lower portion of the sequence is monotonous gray-green and buff to green volcaniclastic rocks with minor mafic lavas, but four members stand out in this sequence. The base of the sequence in the mine area is a black serpentinous slaty argillite that contains lenses and layers of pillow lavas and flow lavas. A lens of flattened pillow lavas completely enclosed in blackish gray slate at Schooner Cove (Plate 5-1) may represent this same horizon at the coast. There, the slaty argillite appears to be injected into and around nearby light greenish to buff weathering fragmental mafic volcanics and volcaniclastics (Plate 5-2). Three other distinct members occur in the lower portion of this sequence, including a conglomerate, a fragmental volcanic and a clastic ultramafic rock. The conglomeratic unit is separable at map scale (Figure 1-1) and was described in detail by Bursnall (1975) as follows:

Distinctive bedded conglomerate-breccia horizons and a megabreccia unit occurring within cleaved black argillite occurs at approximately 300 m from the observed base of the formation The member obtains a maximum thickness of 100 m north of Mine Pond but thins rapidly to the south into mafic lavas; to the north it is cut by late faults...

The conglomerate-breccia beds (normally < 30 cm) consist of both matrix and self-supporting clasts within a blue-black strongly cleaved matrix... The clasts are somewhat deformed...and vary from angular to well-rounded. Clast size is normally less than 12 cm in longest dimension. Compositional variation is apparently restricted to locally derived lithologies; rodingite, other altered gabbroic lithologies, rare serpentinite, (?)quartz-albitite, and relatively fresh dolerite can all be ascribed to the Advocate Complex; variably altered fine-grained basaltic clasts are probably derived from the... Advocate Complex. A qualitative analysis of clast angularity indicates that the detritus demonstrably from the gabbroic assemblage is, in general, more rounded than the basic volcanic lithologies — suggesting a more local derivation for the latter. None of the clasts apparently contains a pre-depositional tectonic fabric.

Lying above the conglomerate-breccias a zone of lenticular outcrop-brecciated leucogabbro bodies occurs... These may measure as much as 350 by 50 m and are completely surrounded by cleaved black and gray argillite and rare coarser grained volcaniclastics. It is suggested that they are of sedimentary origin, and are probably large rafts of Advocate Complex... gabbro.

In addition to this conglomeratic unit, a fragmental rock, composed of rounded, medium grained gabbroic clasts set in a medium to fine grained clastic matrix, outcrops north

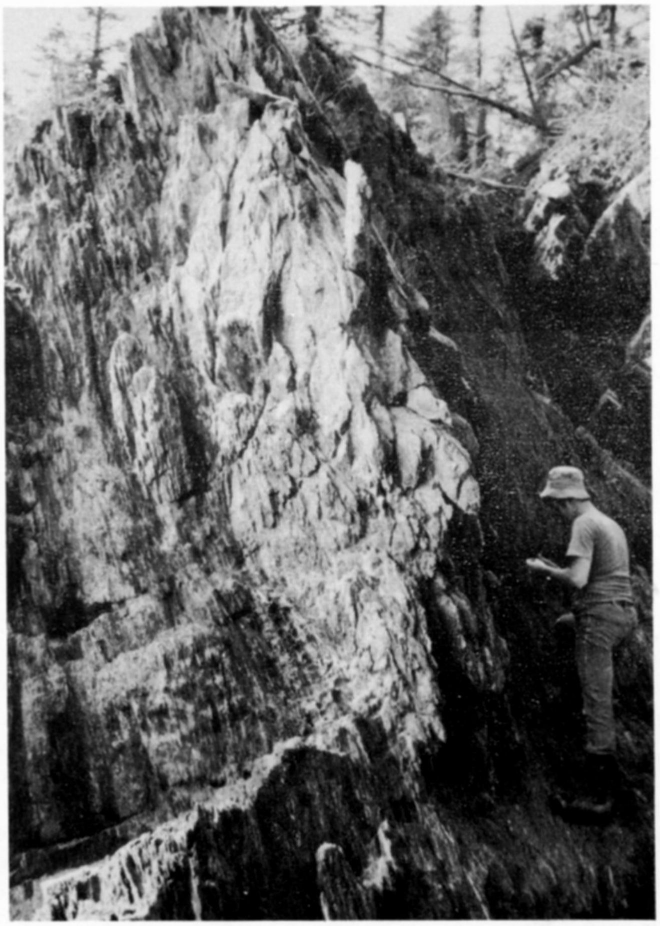


Plate 5-1: *A lens of flattened pillow lava within black slate at Schooner Cove, Baie Verte. These rocks mark the base of the Advocate Complex cover sequence at this locality. John Young discovers rock other than sheared serpentinite on the peninsula.*

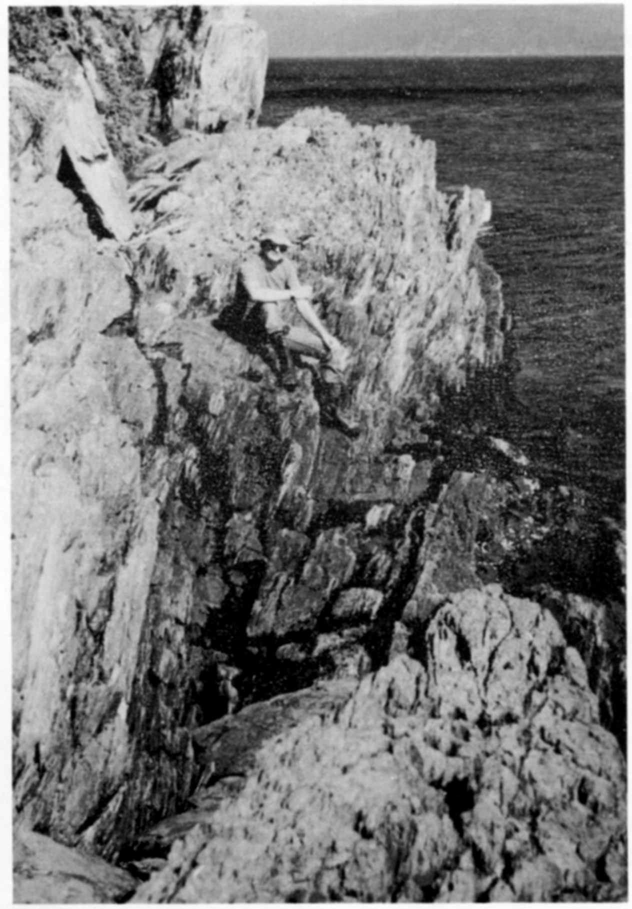


Plate 5-2: *Grayish black slaty argillite, outlined beneath J. Young, penetrating lighter colored volcanic fragmental rocks, Schooner Cove; Advocate Complex.*

of Mine Pond. The remaining distinctive member of the lower portion of the volcanic-sedimentary sequence was described by Bursnall (1975) as follows:

> Lying above and possibly in contact with the agglomerate unit is a distinctive blue-black coarse-grained ultramafic clastic horizon. No banding has been observed but the outcrop pattern and petrography of the unit suggests a pyroclastic origin. In thin-section angular poorly sorted clasts of altered mafic gabbro, clinopyroxene-bearing partially serpentinised ultramafic, highly altered sub-opaque plagioclase, banded serpentinite, and foliated altered gabbro occur in a medium- to fine-grained matrix of predominantly angular fragments of serpentinite. Magnetitic dusting within serpentinite may mark original olivine grain boundaries and internal cross-fractures: similarly, implicate boundaries between pyroxene and serpentinite may be relic of primary igneous textures.

The upper portion of the sequence in this area is dominated by north facing Carlisle sequences (Carlisle, 1963) of pillow lava, pillow breccia, green chert, and tuffaceous and fragmental volcanic rocks. In the area of Big Head, Bursnall (1975) noted a layer, up to 2 m thick, of quartzite with local fragments of black cherty argillite.

Very limited petrographic work has been carried out on these rocks, though Bursnall (1975) described the pillow and flow lavas in thin section as follows:

> In thin-section the mafic pillowed lavas and mafic sheets... are petrographically basaltic but exhibit a considerable variation in texture and relative distribution of pyroxene and plagioclase phenocrysts. An extremely fine-grained and hypocrystalline groundmass is almost invariably present: variolitic growths and a well developed hyalopilitic texture are occasionally present within the brownish 'palagonitic', partially recrystallised, glassy groundmass. Vesicular basaltic rocks are common, in which case chlorite and calcite are the most common secondary infill minerals: albite, quartz, prehnite, (?)pumpellyite, and unidentified zeolites are also variably, but less frequently, present—particularly within the less deformed material.
>
> Phenocrysts, when present, are mainly equant subhedral to euhedral pyroxene occurring as isolated crystals or in rounded aggregates resulting in a pronounced glomeroporphyritic texture and lath like polysynthetically twinned plagioclase. The commonly strained state, small grain size (generally <1 mm), and the partial saussuritisation of the majority of plagioclase phenocrysts has inhibited optical determination of compositions. A coarser grained holocrystalline and doleritic textured variety, possibly an intrusive sheet, contains, however, well-preserved but strained plagioclases which fall within the lower labradorite compositional field. The pyroxene is normally a colourless augite, may exhibit complex twinning, and frequently shows marginal uralitisation. Alteration is, however, noticeably much less advanced than in the... gabbroic assemblage of the

Advocate Complex. Sphene, magnetite, leucoxene, epidote/clinozoisite, chlorite, secondary albite, actinolitic amphibole, and very minor quartz, are the most common accessories.

Rocks similar to those of the Advocate - Big Head area occur in the highly dismembered terrane south of the Advocate Fault. Here, the cover rocks are much more intensely disrupted than in the Advocate - Big Head area, and are commonly tectonically intermingled with shreds of the ophiolitic rocks.

Rocks in the area east of Butler's Pond are strikingly similar to those at the base of the volcanic-sedimentary sequence at Advocate Mines. East of Butler's Pond, black slaty argillite is closely associated with mafic pillow lavas. Bursnall (1975) noted the similarity of these rocks and the possibility that the Butler's Pond sequence may be the equivalent of the Advocate basal unit and may occupy the southern limb of the syncline in the Advocate - Big Head area.

Conglomerate horizons, similar to those indicated in Figure 1-1 in the Mine Pond area, are locally prominent within the southerly dismembered sequence. These are particularly noticeable at the Terra Nova Mine site and on Rattling Brook. At the mine site, large blocks of serpentinite, brecciated serpentinite, and massive sulfide with metre scale dimensions are enclosed in a gray-green and brown slate. On Rattling Brook, thin black slate members intervene between many ophiolitic rock types; these locally contain pebbles and cobbles of ophiolitic and mafic volcanic origin. The disposal of the slate members around these rocks is thought-provoking; some of the larger tracts of ophiolitic rock sandwiched between the members may have been huge blocks deposited within a scant black slate matrix. Considering the size of the leucogabbro blocks near Mine Pond, this idea is highly plausible, particularly in view of the chaotic character of this more southerly sequence. The remainder of the cover rocks in the southerly area are mostly cleaved, monotonous, gray-green volcanic and volcaniclastic rocks.

Cover rocks of the Marble Cove sequence form a distinctive member of the Advocate Complex, as they are more highly deformed and recrystallized than the other cover rocks described above. These greenschists are commonly mylonitic and rarely exhibit primary features. They were described by Bursnall (1975) as:

> ... fine-grained, strongly schistose, green to greenish-gray schists that are occasionally rich in epidote, resulting in a lightly speckled appearance in outcrop. Clearly banded medium-grained buff (epidote-rich) to light green (epidote-poor) schists occur, however, in relatively less deformed areas. Individual layers may exceed 30 cm in thickness with sharp interlayer contacts and are likely to be of primary, sedimentary, origin. Elsewhere the relic banding has been partially destroyed or considerably modified by the subsequent deformations.....

Locally, Bursnall (1975) noted lenses and pods of fine grained metagabbro, massive mafic rocks, quartzites, and possible pillow lavas within the sequence. The questionable pillow lavas, first reported by Neale (1957) on the northwestern shore of Lower Duck Island Cove, are composed of bulbous elongate forms in massive mafic rock.

The portions of the cover adjacent to the altered metagabbroic rocks in the Marble Cove sequence have been altered (Bursnall, 1975). These metasomatites are generally massive, dull gray, fine grained rocks that are locally speckled with fine grained actinolite. The alteration is locally splotchy, indicating that it predates the regional deformation (Bursnall, 1975). Their microscopic characteristics were summarized by Bursnall (1975) as follows:

> In thin-section, porphyroblastic aggregates of greenish brown amphibole rimmed by colourless tremolite/actinolite occur in a matrix of predominantly finely granular clinozoisite-zoisite, quartz, and albite; sphene is a common accessory.

CONTACT RELATIONSHIPS

In the Duck Island Cove sequence, the contact between the lower and upper portions of the Advocate Mines - Big Head cover sequence is unexposed (Bursnall, 1975). Considering the extent of tectonic dissection in the complex, a tectonic contact is more likely than a conformable contact. This sequence is tectonically separated from the more highly dismembered sequence in the complex to the south.

The nature of the contact between the ophiolitic rocks and the cover rocks is varied. In most places, it is tectonic, such as in the Sisters Cove sequence and in many other portions of the complex where dismembered ophiolite slivers have been tectonically introduced into the cover sequence. Locally, map scale ophiolitic blocks may represent depositional blocks within this assemblage. The contact between the Advocate ultramafic body and overlying rocks has been interpreted as being originally unconformable (Bursnall, 1975), mainly because of the geometry of the contact and the nature of the overlying strata. The contact is highly strained and locally marked by reverse faults (see Chapter VII). It should be noted, though, that it is a contact of high lithological contrast and, thus, susceptible to movement during regional deformation. Most cogent for the concept of an unconformity at this location is that the same basal slaty argillite and pillow lava member of the cover sequence appears to transect the underlying ophiolite stratigraphy. Near the Advocate Mine, it rests on serpentinized ultramafic rocks whereas, to the east, the member overlies gabbro of the ophiolite sequence. In addition, the very coarse conglomerates containing abundant and large fragments of ophiolitic origin support the idea of an exposed ophiolite and, hence, an originally unconformable relationship at this highly strained contact. The interpretation of an originally unconformable relationship between the Duck Island Cove ophiolitic and cover sequence is adhered to in this report.

The contact between ophiolitic and overlying rocks in the Marble Cove sequence "is gradational and frequently poorly defined across a zone of presumed metasomatically altered lithologies" at Marble Cove (Bursnall, 1975).

DEPOSITIONAL ENVIRONMENT

Cover rocks within the Advocate Complex represent a submarine sequence that appears to have been deposited directly on oceanic crust. Portions of the more highly tectonized cover may include pillow lavas genetically related to the ophiolite. The apparent local unconformity between the cover and the ophiolite in the Duck Island Cove sequence indicates that oceanic basement was exposed during deposition of the cover, and the coarse metaconglomerates containing dominantly ophiolitic detritus demonstrate that the basement was undergoing active erosion. The huge depositional ophiolitic blocks in the cover suggest a proximal relationship of the cover sediments to the exposed basement. In addition, mafic volcanism was continuous within this environment, and was locally explosive as indicated by the ultramafic pyroclastic deposit near Mine Pond. All this evidence strongly indicates that the cover rocks were deposited along, or near, an ocean floor fault that exposed deeper portions of the oceanic crust.

Significantly, the garnet amphibolite aureole at the Advocate Pit and, thus, related faulting, formed very early in the structural history of the complex (Bursnall, 1975) (see Chapter VII). I suggest that this early faulting, related to the tectonic stacking of sequences in the complex, provided the source terrain for the cover clastics and interlayered volcanics (Figure 1-1).

AGE AND CORRELATION

There is no direct evidence for the age of the cover rocks on the Advocate Complex. Black slates along the highway north of Baie Verte have been checked for graptolites and acritarchs, but appear to lack identifiable fossils. Cover rocks of the Advocate Complex were traditionally considered as part of the archaic Baie Verte Group, which was considered equivalent to the fossiliferous Arenigian Snooks Arm Group. In the present study, it is apparent that there are major differences between the Advocate cover sequence and the Snooks Arm Group, and the relationships between these rocks and their ophiolitic foundation are considered fundamentally different. Locally, the Advocate cover sequence appears to unconformably overlie its basement, whereas the Snooks Arm Group conformably overlies the Betts Cove ophiolite.

The base of the cover sequence in the Duck Island Cove sequence, as well as the conglomeratic slates, are lithically similar to basal deposits of the Flat Water Pond Group immediately south of the complex, which are also juxtaposed against ophiolitic rocks of the Advocate Complex along the Baie Verte Line (see Flat Water Pond Group). In addition, the gabbroic conglomerate near Advocate Mine is identical to distinctive gabbroic boulder conglomerates that outcrop in many portions of the Flat Water Pond Group; this rock type is not recognized anywhere else on the peninsula. Thus, it is likely that part if not all of the Advocate cover sequence is equivalent to the Flat Water Pond Group.

Indirect paleontological evidence from strata above transported ophiolites in the Bay of Islands suggests that the Advocate cover sequence and correlative Flat Water Pond Group are Llanvirnian or younger in age. Casey and Kidd (1981) reported that fossiliferous Llanvirnian clastic rocks of the Crabb Brook Group unconformably overlie the Bay of Islands ophiolite and were derived from the ophiolite sheet during its emplacement. The basal rocks of the unit are distinctive coarse conglomerates containing abundant ophiolitic clasts. Thus, the tectonic setting and clast content of the lower Crabb Brook Group are similar to the Advocate cover rocks and they were most likely deposited in the same unique environment. Since the ophiolitic rocks underlying each sequence are considered as portions of the same sheet of oceanic crust, it is suggested that the distinctive Advocate cover sequence and correlative Flat Water Pond Group formed during the same event as the Crabb Brook Group. The age of the latter suggests that the Baie Verte rocks are likely Llanvirnian or possibly younger in age.

POINT ROUSSE COMPLEX

The Point Rousse Complex comprises the mafic and ultramafic rocks that outcrop on the Point Rousse Peninsula between Baie Verte and Ming's Bight. It is superficially similar to the Advocate Complex, as it is composed of a dismembered ophiolite overlain by a dominantly mafic volcanic-volcaniclastic cover sequence. The Point Rousse Complex is structurally more intact than the Advocate Complex; it also displays a different cover sequence that conformably overlies the ophiolitic components, in contrast to the apparently unconformable Advocate Complex cover sequence.

The Point Rousse Complex is disposed in a broad, generally east trending, structurally modified synclinorium (Church and Stevens, 1971; Kennedy, 1973) with ophiolitic plutonic components occupying zones to the north and south of the centrally located cover sequence. One small sliver of ultramafic rock, geographically isolated from the complex, outcrops on the east side of Ming's Bight and additional ultramafic members of the complex occur on Grassy Island, the Sisters and Tin Pot Islands in Baie Verte (Figure 1-1). The ophiolitic components are confined to structural blocks bounded by high angle and thrust faults; the thrust faults dip moderately to the northwest. One thrust slice of pillow lava and volcaniclastic rocks is structurally interleaved with the northerly ophiolitic plutonic rocks. The volcanic rocks that form the top of the ophiolite sequence are only locally distinguishable from the overlying cover sequence and, thus, are discussed with the Point Rousse cover sequence, below. It is uncertain if this cover sequence is as structurally dissected as the ophiolitic plutonic component because of poor inland exposure. The maximum outcrop width across the complex is about 7500 m. Late porphyritic diabase dikes crosscut both the ophiolitic and cover rocks of the complex. All components of the complex are well exposed along coastal sections, which are herein considered as reference sections for the complex.

The complex is tectonically bounded. The contacts range from thrust faults, such as the southerly Scrape Fault (Figure 1-1), to high angle faults, such as the one which separates the complex from the Ming's Bight Group at South Brook.

The complex has been affected by a single, moderately northwesterly dipping, penetrative cleavage that is only weakly developed, locally, in the plutonic ophiolite assemblage, except near major faults. The northerly volcanic thrust slice and the southerly portion of the cover rocks in Ming's Bight are moderately to intensely deformed, whereas the remainder of the complex is only slightly to moderately deformed. Thrust faults within the complex are generally either cogenetic with or later than the cleavage (Kidd et al., 1978) and appear to predate high angle faulting. The whole of the complex was metamorphosed to the greenschist facies.

Rocks of the complex were originally considered as part of the Baie Verte Formation (Watson, 1947), which was later raised to group status (Baird, 1951). Mafic and ultramafic plutonic rocks on the Point Rousse Peninsula were considered by most workers to be post-Early Ordovician intrusions into the mafic volcanic sequence (Watson, 1943; Baird, 1951; Neale, 1957), though Kennedy and Phillips (1971) interpreted ultramafic rocks on the Baie Verte Peninsula as pre-Early Ordovician intrusions that were unconformably overlain by the mafic volcanic sequence; the latter workers included mafic intrusions on the peninsula in the Baie Verte Group. Norman (1973) and Norman and Strong (1975) followed most workers of the time by interpreting the plutonic rocks as a portion of an ophiolite suite, and included them in the Baie Verte Group. Kidd (1974) and Kidd et al. (1978) also considered these rocks to be part of the Baie Verte Group, and separated the ophiolite, which they termed the Ming's Bight ophiolite complex, from the overlying volcanic-volcaniclastic sequence. The term "Ming's Bight Ophiolite Complex" is not recognized herein as it conflicts with the already established Ming's Bight Group (Baird, 1951; see Fleur de Lys Belt). Williams et al. (1977) abandoned the term Baie Verte from stratigraphic usage and informally assigned rocks on the Point Rousse Peninsula to the Point Rousse Complex. This latter name is retained in this report.

Ophiolitic rocks of the complex are petrographically similar to those of the Advocate Complex; hence, the following discussion generally highlights the macroscopic features of the Point Rousse ophiolitic components, and the reader is referred to the Advocate Complex for general descriptions.

Ophiolitic Plutonic Rocks

All of the components of an ophiolite suite are present in the Point Rousse Complex, including variably serpentinized ultramafic rock, gabbro and metagabbro, and a well developed sheeted dike section; the gabbro and diabase are locally altered in areas of high strain. These components are disposed on both the northwestern and southeastern side of the regional syncline.

SERPENTINIZED ULTRAMAFIC ROCK

These rocks outcrop on both the northwestern and southeastern sides of the complex; in addition, based on a positive aeromagnetic anomaly (Geological Survey of Canada, 1968a) and limited exposure on the Tin Pot Islands, the Sisters shoal, and Grassy Island, an extensive body of ultramafic rock is inferred to underlie the mouth of Baie Verte. Ultramafic rock was also noted during calm water on a submerged shoal (locally known as Whore's Knob) approximately 1.5 km west-northwest of the Sisters shoal (Figure 1-1). Another offshore positive aeromagnetic anomaly just north of Pacquet Harbour (Geological Survey of Canada, 1968b) may represent a detached portion of the Point Rousse Complex or an ultramafic slab within the Ming's Bight Group similar to smaller bodies in the group (see Ming's Bight Group).

The ultramafic rocks are serpentinized to various extents and are similar to those of the Advocate Complex. Field appearance and thin section studies by Watson (1943), Norman (1973) and Kidd et al. (1978) indicate that lherzolite, harzburgite, dunite, and pyroxenite are all present in the complex. Both noncumulate and cumulate ultramafic rocks occur locally but most of the ultramafics are serpentinized.

Harzburgite, with minor thin to medium interlayers of dunite, occurs on the islands in Baie Verte and is inferred to form most of the ultramafic body that probably underlies the mouth of Baie Verte (Norman and Strong, 1975; Kidd et al., 1978). Based on field distribution, harzburgite and subordinate dunite also seem to compose most of the ultramafic rocks to the south and east of Ming's Bight and at Red Point. All of these rocks have a xenomorphic granular texture and appear to be noncumulate in origin (Kidd et al., 1978). Locally, in these ultramafics, chromite forms very thin layers that rarely exceed a few millimetres in thickness. Watson (1943) noted minor lherzolite in association with harzburgite on Grassy Island. Kidd et al. (1978) estimated that the minimum thickness of these noncumulate ultramafic rocks is 1800 m, if their probable extension beneath the mouth of Baie Verte is considered.

Dominantly medium grained, layered cumulate harzburgite and gabbro and rare lherzolite are exposed best between Deer Cove and Western Point, but also occur south of Red Point, in Hammer Cove, and in the southernmost ultramafic body in the center of the Point Rousse Peninsula. The layered ultramafic rocks between Deer Cove and Western Point are surrounded by gabbro and constitute an area too small for resolution in Figure 1-1. In addition to these cumulate rocks, highly deformed and serpentinized layered ultramafics are associated with minor gabbro in the Trimms Brook - Ming's Bight road area as well as northeast of Kidney Pond; these occurrences are probably cumulate in origin. Layering in all of these rocks is generally less than 50 cm thick. At many of these localities, large (up to 10 m diameter) blocks of ultramafic rock outcrop in the layered sequence; these blocks may represent tectonic blocks, xenoliths or autoliths (Norman, 1973; Kidd et al., 1978). Kidd et al. (1978) described the sequence east of Western Point as follows:

> ...here, a sequence of altered (carbonate-talc) cumulate ultramafic rocks, originally mostly harzburgites, orthopyroxenites and websterites, are interlayered with some gabbroic rocks. Large 'sieve-poikilitic' oikocrysts of pseudomorphs after orthopyroxene up to 3 cm across occur in some of the ultramafic layers. The layers are typically 5-20 cm thick; some show normal size grading. Several other layers show a gradation from altered poikilitic harzburgite to gabbro over a thickness of about 30 cm. A few very fine grained, gray layers up to 50 m thick, now composed of talc with some magnesite, occur in the sequence; they are inferred to have been cumulate dunites. Although faulted, the section seems to first lose olivine and, soon after, orthopyroxene, and gain clinopyroxene and plagioclase upsection to the east.

Norman and Strong (1975) reported irregular, elongate patches of lherzolite within cumulate harzburgite layers in the same area.

The base of the layered ultramafic sequence is not exposed but its relationship to other ultramafics in the complex is everywhere tectonic (Kidd et al., 1978). Kidd et al. (1978) estimated this section with associated gabbro to be at least 300 m thick.

Locally, the serpentinized ultramafic rocks in the Devil's Cove and Hammer Cove area are brecciated. The breccia at Devil's Cove consists of a few metres of weakly cleaved granular serpentinite containing angular to subrounded serpentinite clasts up to 50 cm across within massive serpentinite. At Hammer Cove, the rocks are steatitized ultramafics in a talc-carbonate matrix. Norman and Strong (1975) noted layered talcose clasts and mafic clasts within this breccia. Kennedy and Phillips (1971) interpreted both breccias as conglomerates marking profound unconformities in the Point Rousse sequence. Norman and Strong (1975) indicated that there is no evidence for the interpretation of these rocks as sedimentary, and indicated that the breccias are related to thrust faulting at these localities. Kidd et al. (1978) mapped the Devil's Cove occurrence in detail and concluded that the breccia zone defines a half-cylinder form within the surrounding massive serpentinite; in view of this shape and the monomict character of the breccia, they suggested that it represents a gas-brecciation pipe. They indicated that the Hammer Cove breccia may be partially igneous in origin, as well as having a tectonic component. In this study, no new evidence has been gathered to resolve the differences between Norman and Strong (1975) and Kidd et al. (1978); both interpretations may be correct.

In places, a weakly to moderately developed foliation defined by mafic minerals occurs in both the noncumulate and cumulate ultramafic rocks. Since the alignment of mafic minerals could only have formed at high temperatures, Kidd et al. (1978) interpreted this fabric as a high temperature foliation that predates any regional deformation in the complex.

Overall, the Point Rousse ultramafics are less serpentinized than those of the Advocate Complex, though they display the same general features. Serpentinization is variable throughout the Point Rousse Complex rocks, ranging from 10% to 100%, though generally greater than 50% (Kidd et al., 1978). The most highly serpentinized rocks occur in the small sliver of ultramafics on the east side of Ming's Bight and in the bodies south of Ming's Bight; the least altered rocks outcrop on the islands in Baie Verte.

Steatitization and carbonatization of the ultramafic rocks are common in the complex. Talc-carbonate rocks form a discontinuous zone between Deer Cove and Red Point; Norman and Strong (1975) reported that talc predominates in the zone between Deer Cove and the pond immediately southwest of Devil's Cove Pond, whereas carbonate forms up to 80% of the rocks between this pond and Red Point. Using X-ray diffraction, they determined the carbonate to be magnesite with minor calcite and dolomite. In addition, they noted that talc is most abundant in highly sheared zones. Talc-carbonate rocks are also common in the area of Three Corner Pond south of Ming's Bight, and near the margins of the large ultramafic body south of Ming's Bight, particularly in the Trimms Brook area. Locally, such as near Trimms Brook and in small patches in the ultramafic body on the east side of Ming's Bight, talc occurs with only very minor carbonate and, thus, constitutes soapstone.

Locally, silicification accompanies carbonatization to form a bright greenish white altered rock called virginite (see Advocate Complex). This alteration is common in areas of high strain, particularly in Deer Cove, on the small pond directly south of Deer Cove, and in the small ultramafic occurrences along the Scrape Thrust. Intense chloritization of ultramafic alteration products was noted locally along the Scrape Thrust, in places along the Ming's Bight road, and in the ultramafic sliver east of Three Corner Pond.

GABBRO

Gabbros of this complex are similar to but generally better preserved than those of the Advocate Complex. Kidd et al. (1978) summarized the general features of the gabbroic rocks as follows:

> The generally medium-grained (1-2 mm) equigranular gabbros display well-preserved textures even though the plagioclase is ubiquitously altered, in places to albite but more commonly to an ultrafine-grained, turbid, zoisite-clinozoisite aggregate. The rare to subordinate orthopyroxene is almost always altered, whereas the clinopyroxene in places remains fresh. Opaque minerals are conspicuously absent.

These rocks range from layered to massive, with the layered varieties being gradational between layered cumulate ultramafics and massive leucocratic gabbro. Kidd et al. (1978) reported that cumulate textured gabbros have been identified at only two locales other than where they occur with the cumulate ultramafic rocks described above; these locales are northeast of Eastern Point and southeast of Point Rousse. The composition of the cumulate zones ranges from feldspathic pyroxenite to anorthositic gabbro. Layered gabbro also occurs along the coast south of Red Point and inland east of the ultramafic 'spine' of the Point Rousse Peninsula and northeast of Three Corner Pond. Along the coast south of Red Point, size graded layers were noted, as well as scour channels and three examples of crossbedding within the layers (Plate 5-3). Commonly, diabase dikes less than 1 m wide cut the layered sequences. A gneissic foliation identical to the high temperature foliation of the ultramafic rocks was identified in the layered gabbroic rocks (Kidd et al., 1978). Kidd et al. (1978) reported that irregular, podlike dikes of coarse pegmatitic gabbro crosscut both layering and foliation in most places, though they noted one occurrence of a foliated pegmatitic pod. Locally, the pegmatitic gabbro is gradational with surrounding medium grained gabbro, with no obvious boundary between them. Approximately halfway between Big Head and Red Point, the layered gabbros are cut by some aphanitic, cream colored felsic dikes up to 12 cm wide.

Massive, medium grained, generally leucocratic gabbro occurs along the Baie Verte coast south of Fox Gulch, west of Devil's Cove and, locally, north of Big Head, as well as at numerous localities inland. Norman (1973) reported a transitional contact between layered and massive gabbro south of Fox Gulch. Norman (1973) and Kidd et al. (1978) reported rare layering within the massive gabbro and local pegmatitic patches.

Kidd et al. (1978) estimated a minimum thickness of the layered gabbro sequence to be 300 m, based on the section at Point Rousse, and approximately 300 m for the homogeneous massive gabbro. They noted that the gabbroic member is likely to be greater than 600 m thick since the extent of gabbro south of Western Point and southwest of Devil's Cove suggests thicknesses of gabbro greater than 800 m.

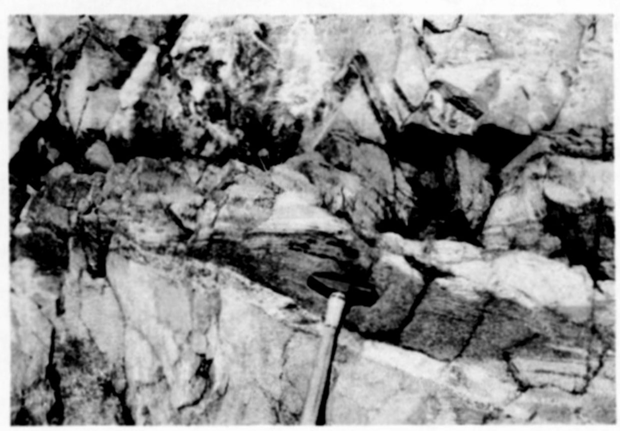


Plate 5-3: *Crossbedding within gabbro south of Red Point, Point Rousse Peninsula; Point Rousse Complex.*

SHEETED DIABASE DIKES

These are well exposed along the coast north of Big Head, south of Point Rousse, and between Lower Green Cove and Green Cove; inland, these dikes occur to the east of Three Corner Pond, though their extent there is poorly defined due to the lack of continuous outcrop. South of Point Rousse, the sheeted dikes are in transitional contact with gabbroic rocks continuous with layered gabbros at the point.

The dike terrane is composed of nearly 100% diabase dikes with well defined chilled margins; in places, screens of either gabbro or pillow lava occur in the unit, though these are usually confined to areas near the contacts of sheeted dikes with these other units. Norman (1973) noted that coarse grained, wide diabase dikes occur within the Green Cove section and south of Point Rousse, whereas thinner, finer grained dikes occur immediately north of Red Point and north of Big Head. He described each type as follows:

> The coarse diabase dikes... are composed of actinolite (replacing pyroxene) in a groundmass of fine grained epidote and sericite, with local secondary quartz. The chilled contacts of these dikes consist of minute fibrous grains of actinolite in a turbid groundmass of epidote and minor quartz.
>
> The fine grained diabase dikes... generally consist of about 40% metamorphic plagioclase (An_{10} - An_{15}), 40% actinolite, and 20% epidote, chlorite, calcite, sphene, zoisite and quartz; the chlorite commonly replaces actinolite. Secondary quartz is common locally.... Relict igneous textures vary from granular to intergranular, with plagioclase laths averaging 5 mm in length.

The sheeted dikes commonly display breccia zones similar to those in other sheeted dike units that have been attributed to gas brecciation (Williams and Malpas, 1972). South of Point Rousse and north of Big Head, fine grained, trondhjemite forms the matrix to net-vein breccias within the sheeted dikes, near their contact with the gabbro unit (Norman, 1973; Kidd et al., 1978).

Norman (1973) and Kidd et al. (1978) noted a difference in the trend between the sheeted dikes at Green Cove and Point Rousse and those at Big Head. The dikes at Big Head strike nearly east-west, whereas the others strike generally north-south; all are steeply dipping. This pattern led to the suggestion that the dikes either originated from two separate episodes of spreading, each with a different spreading direction, or formed together and subsequently were tectonically rotated with respect to each other (Kidd et al., 1978).

Norman (1973) estimated the sheeted dike section near Green Cove to be at least 900 m thick. In Green Cove, the sheeted dikes are in transitional contact with overlying pillow lavas; dikes in this transition zone generally contain screens up to 30 m long of pillow lava. One exposure here shows a diabase dike feeding the pillow lavas.

HIGH STRAIN ZONES

High strain zones occur in the complex, immediately east of Three Corner Pond and on the coast just north of Lower Green Cove. These rocks are generally fine grained, laminated greenschists in which the laminae are discontinuous (Plate 5-4). Locally, small (up to 20 cm) knots and pods of diabase and gabbro displaying primary textures are preserved in these schists. The mafic schists are locally gradational with gabbro around Three Corner Pond; thus, these rocks are herein considered to represent highly strained gabbro and diabase.

Plate 5-4: *High strain schists on coast just north of Lower Green Cove; note the discontinuous nature of the fine laminae and "knot" of diabase in upper center portion of photograph. Diabase "knot" approximately 10 cm long; Point Rousse Complex.*

CONTACT RELATIONSHIPS

Kidd et al. (1978) summarized the ophiolitic stratigraphy of the Point Rousse Complex as follows:

> The section consists of non-cumulate, residual, tectonite harzburgite containing minor dunite layers. Layered cumulate ultramafic rocks, consisting mainly of clinopyroxenite and websterite, with subordinate gabbro, dunite and harzburgite succeed the non-cumulate harzburgites. These grade up into layered, originally cumulate gabbros, which are affected in many places by a gneissic mineral foliation. Homogeneous, somewhat leucocratic gabbro overlies the layered, foliated gabbros; it is locally cut by net-vein breccias of trondhjemite that are also uncommonly found in the lowest part of the succeeding 100% sheeted diabase dike complex. Parallel diabase dikes cut the homogeneous gabbro and some cut the trondhjemite; the dikes increase in abundance upward at the expense of the gabbro until the rock consists entirely of parallel diabase dikes. Pillow lava screens occur above at least 350 m of sheeted dikes. These increase in abundance upward at the expense of the dikes, so that, at the top of the ophiolite complex, the rock consists entirely of pillow lava.

Point Rousse Cover Sequence

Moderately northerly dipping, generally east trending mafic volcanic and sedimentary rocks outcrop between the two ophiolitic terranes of the complex; this cover sequence has a maximum outcrop width of about 5 km. In addition, a small thrust sheet of deformed pillow lavas and greenschist lies between Deer Cove and Devil's Cove Pond. Continuous sections of the cover sequence are well exposed along the coast of Baie Verte and Ming's Bight, whereas small inland exposures are dominantly nondescript fine to medium grained greenschist.

South facing, overturned pillow lava sections that overlie the ophiolitic plutonic rocks at Green Cove and south of Big Head appear to represent the top of the ophiolitic assemblage. They appear to overlie the sheeted dike section transitionally, though at Big Head this original relationship is obscured by a fault near the contact zone. These ophiolitic pillow lavas are definitely identifiable only in the coastal sections of the complex and appear to be unmappable inland. Norman and Strong (1975) noted that the pillow lavas at Big Head are more mafic than those at Green Cove; they described the petrography of both types of lava as follows:

> [At Big Head] Typically, the pillows are composed of about 40% actinolite, 20% calcite, and 40% epidote, chlorite, sericite, quartz, zoisite and clinozoisite.
>
> [Green Cove lavas have]... assemblages of approximately 50% plagioclase (An_{10}-An_{15}), 30% chlorite, 10% epidote and 10% opaque minerals. The margins of the plagioclase laths, up to 0.5 mm in length, are corroded and replaced by groundmass chlorite.

The lavas at Big Head are associated with distinct layers (<1 m wide) of burgundy colored chert.

The remainder of the cover sequence exposed at the coast conformably overlies the ophiolitic volcanics. Norman (1973) and Kidd et al. (1978) described the section at Ming's Bight in detail. Here, the sequence is approximately 2 km thick and consists of a northern, south facing, overturned section of dominantly gray-green volcaniclastic rocks and a southern section of mixed pillow lava, mafic pyroclastic rocks, and minor volcaniclastics. Norman (1973) maintained that the southern section is north facing and in tectonic contact with the northern section, whereas Kidd et al. (1978) considered the whole sequence to be conformable and south facing; the latter authors recognized the tectonic disruption that Norman (1973) mapped as a fault, but suggested that major displacement has not occurred along it. Facing criteria observed inland during the present study indicate that portions of the southern section are north facing as indicated by Norman (1973). However, I agree with Kidd et al. (1978) that the high strain zone is probably not a zone of major displacement since no major displacement is apparent along the zone, nor are contrasting rock types juxtaposed along the zone.

Kidd et al. (1978) erected two formations representing the northern and southern sections of the sequence. These formations are not used in this report since they are defined only by coastal type sections and have not proven to be mappable inland.

The northern volcaniclastic section is composed of four fining upward sequences, each of which is initiated by a basal conglomerate (Kidd et al., 1978). The conglomerates are matrix-supported and generally contain cream colored cherty argillite and mafic volcanic and volcaniclastic rocks in a gray-green sandstone matrix. Just north of the Barry and Cunningham adit, the conglomerate contains clasts with large (up to 1 cm) distinct clinopyroxene phenocrysts that also occur independently in the matrix. In addition, the section contains sandy to silty epiclastic units, subordinate argillite, lapilli tuff, crystal and crystal-lithic tuffs, thin diabase sills, minor variolitic pillow flows and, locally, pods of marble up to 50 cm in diameter. Kidd et al. (1978) estimated this section to be at least 700 m thick. Inland, at the old Goldenville Mine site, burgundy colored chert is thinly interlayered with magnetite-rich layers (<2 cm wide), forming a typical iron formation. This facies is less than 5 m wide. It appears to be limited in distribution and its stratigraphic position with respect to the coastal section is uncertain, but it may be equivalent to the burgundy chert along strike at Big Head, which is associated with the ophiolitic pillow lavas.

The southern portion of the sequence is composed mainly of pillow lava and pillow breccia, with subordinate mafic pyroclastic rocks, tuff and volcaniclastics. The pillows are pale greenish to buff weathering, commonly amygdular and locally variolitic, with varioles generally <1 cm. Diabase dikes and sills up to 1 m wide are more common in this section of the sequence than in the northern section.

Strata similar to the sequence at Ming's Bight are exposed on the eastern shore of Baie Verte but, due to exposure gaps, the stratigraphy is not as well documented. This coastline section appears to display inverse structural-stratigraphic characteristics with respect to the Ming's Bight section; it consists of a northern, south-facing, overturned section dominated by pillow lava and associated pyroclastics that locally form Carlisle sequences (Carlisle, 1963) and a southern, north-facing, normal sequence of dominantly volcaniclastics with minor mafic pyroclastics. The change in facing directions immediately south of Pumbly Point is considered to indicate the location of the axis of the regional syncline in the cover rocks.

The pyroclastic and sedimentary rocks in the southern section of the sequence along Baie Verte are similar to those in the section at Ming's Bight, and include similar conglomerates and finer clastic rocks, lapilli tuff, and coarse pyroxene-crystal-bearing fragmental rocks. Abundant diabase dikes and sills occur south of Penny Cove.

The upper part of the pillow lava section at Green Cove may be equivalent to the volcaniclastic section south of Big Head; the southern volcaniclastic section on Baie Verte may, in turn, be time-correlative with the southern pillow unit in Ming's Bight. These relationships suggest major facies changes both laterally and vertically through the cover sequences, which is typical of volcanic terranes.

Limited exposures in the inland area between the two coastal sections are uninformative, though two unusual rock types for the complex have been reported from there. Drill core from Mud Pond, retrieved by M.J. Boylen Company and logged by A. Frew in 1968, indicates the presence of an aphanitic buff-gray keratophyre and a dark gray-green quartz-feldspar porphyry. The keratophyre occurs in only one hole, with a maximum intersection of approximately 4 m, whereas porphyry intersections occur in many holes, but are generally less than 2 m thick. These are the only felsic rocks reported from the Point Rousse cover sequence. The second unusual rock type for this cover sequence is a leucocratic gabbro that forms an isolated ridge on the western side of a small pond approximately 2.5 km east of Pine Cove and north of Scrape

Pond. The gabbro is highly fractured and weathered, and much like the ophiolitic gabbros of the complex. It may represent a faulted sliver of ophiolitic gabbro in the cover sequence.

Norman (1973) described a thrust sliver of generally easterly trending cover rocks between Deer Cove and Devil's Cove as composed of minor diabase dikes to the north, succeeded southward by pillow lava, which is succeeded near Deer Cove by volcaniclastic rocks and greenschist. All of these rocks are severely deformed and primary features are difficult to recognize in many places.

DEPOSITIONAL ENVIRONMENT

Point Rousse cover rocks represent a submarine sequence that is part of and conformably overlies the oceanic crust. The lowest pillow lava units at Green Cove and near Big Head are probably part of the ophiolite. The overlying volcaniclastic and volcanic sequence indicates a continuing submarine center for mafic volcanism, which may have been an island arc or an active ocean island volcano.

Characteristics of the cover sequence suggest deposition on the slopes flanking such a major volcanic center. Kidd et al. (1978) suggested that most of the sediments in the sequence were transported by grain flow, with subordinate turbidites present. Grain flow has been likened to sand avalanching and requires relatively steep slopes (18° to 37°) (Middleton and Hampton, 1973). If Kidd et al.'s (1978) interpretation is valid, it also suggests that the Point Rousse cover sequence was deposited proximal to a mafic volcanic center where either active construction of a volcanic edifice or faulting were taking place.

More evidence of a slope environment on the flanks of a volcanic center is provided by the iron formation in the Goldenville Mine area. This member exhibits the features of a deep water iron formation (Kimberley, 1978), in particular the thin bands of ferriferous chert. He noted that the generalized environment of these deposits is on a deep water slope, near a marine volcano.

AGE AND CORRELATION

There is no direct evidence for the age of the Point Rousse cover sequence. It conformably overlies ophiolitic rocks considered herein to be Late Cambrian to Early Ordovician in age (see Ophiolitic Basement). Traditionally, cover rocks of the complex have been correlated with those of the Advocate Complex, and the Snooks Arm Group, and with volcanic rocks of the Pacquet Harbour Group. The fossiliferous Arenigian Snooks Arm Group also conformably overlies ophiolitic rocks and is lithically similar to the Point Rousse rocks. Thus, the Point Rousse cover rocks are considered to be Arenigian.

Correlation of the Point Rousse cover sequence with rocks of the Pacquet Harbour Group is less certain because the northern portion of the group is highly deformed and metamorphosed. Similar volcanic facies with similar geochemical characteristics are present in both units (see Chapter VI), but the Pacquet Harbour Group contains small but significant quantities of felsic extrusive rocks. The Pacquet Harbour Group is in part ophiolitic; the balance of the group is considered partly correlative with the Point Rousse cover.

The Point Rousse rocks are apparently noncorrelative with the Advocate cover sequence. The Advocate rocks display a distinctly different cover sequence that unconformably overlies ophiolitic basement; hence, the Point Rousse cover may be slightly older than the Advocate sequence.

Late Dikes

Two large mappable (>20 m wide) dikes of plagioclase porphyritic diabase intrude the cover and ophiolitic portions of the Point Rousse Complex; one intrudes the cover immediately north of Ming's Bight community whereas the other intrudes harzburgite on the east side of Ming's Bight. Kidd et al. (1978) noted smaller porphyritic dikes north of Big Head, near Western Point, just west of Devil's Cove and on the east side of Devil's Cove. They also noted two nonporphyritic diabase dikes near Big Head and immediately southeast of Point Rousse that have the same trends as the porphyritic varieties. All of these dikes except the ones at Western Point and at Bear Pond in the cover sequence trend approximately 300° (Kidd et al., 1978).

The dikes predate most faulting in the complex since they are truncated by faults. They postdate the process responsible for the divergent orientation of the sheeted dike exposures described above, as they maintain a consistent orientation in every sheeted dike member.

Kidd et al. (1978) also noted rare, thin (< 50 cm), purplish buff diabase dikes containing titanaugite. They crosscut ophiolitic components of the complex, but are of uncertain age.

BETTS COVE COMPLEX

The Betts Cove Complex is herein defined as the sequence of ophiolitic rocks, including layered ultramafics, gabbro, diabase and associated pillow lava, that outcrops as a contiguous though tectonically disrupted unit from Tilt Cove in the northeast to Northwest Arm in the southwest. In addition, ultramafic and gabbroic xenoliths in the Cape Brulé porphyry, gabbro on Gull Island to the east of Cape St. John, isolated patches of pillow lava and greenschist in the Middle Arm and Gull Pond areas to the southwest, and mafic rocks on Bishops Rock to the northeast are all included in the complex. The ophiolitic sequence is incomplete, as it appears to lack an ultramafic tectonite member.

The complex forms a steeply southeasterly dipping arcuate outcrop area between Pittman Bight and Tilt Cove, whereas southeast of Nippers Harbour, it is "more or less flat lying", according to DeGrace et al. (1976).

Neale (1957) noted that the rocks in the northeastern area are disposed on the northwestern limb of a slightly overturned syncline, with its east trending, east plunging axis located in the vicinity of Bobby Cove and Wild Bight (Figure 1-1). The ophiolite suite is developed best in the area of Betts Cove, where it was estimated to be approximately 3000 m thick (Upadhyay et al., 1971; Upadhyay, 1973). Riccio (1972) and Upadhyay (1973) recommended the section along a line joining Betts Island, Betts Cove, and Kitty Pond as the type section of the complex. The complex changes character along strike to the northwest of Betts Cove, so that at Tilt Cove it is only 750 m wide and highly fault-modified and the dike zone is represented only by diabase breccia (Upadhyay, 1973).

The complex is metamorphosed in the lower greenschist facies except for the areas peripheral to the Cape Brulé porphyry and Burlington Granodiorite, where they locally attain amphibolite facies. Coish (1977a,b) documented greenschist grade ocean floor metamorphism within the complex. Structurally, the complex is relatively intact compared to the other complexes on the peninsula.

Snelgrove (1931) originally assigned the mafic plutonic rocks between Pittman Bight and Tilt Cove, here assigned to the Betts Cove Complex, as well as volcanic and volcaniclastic rocks to the southeast, to the Snooks Arm Group. Baird (1951) mapped similar rocks to the southwest as the Nippers Harbour Group, which he considered equivalent to the Snooks Arm Group. Snelgrove (1931) considered the ultramafic rocks of the area as Acadian intrusions whereas Baird assigned a post-Ordovician - pre-Late Devonian age to them. These ultramafics were later included in the Snooks Arm Group by Upadhyay et al. (1971) and the Nippers Harbour Group by Schroeter (1971). DeGrace et al. (1976) noted that Upadhyay (1973), Upadhyay et al. (1971), Schroeter (1971) and Baird (1951) all considered the Snooks Arm and Nippers Harbour Groups as equivalent; hence, DeGrace et al. (1976) included rocks formerly termed the Nippers Harbour Group in the Snooks Arm Group. Upadhyay and Neale (1976) noted that the ophiolitic rocks and overlying volcanic and volcaniclastic rocks of the Snooks Arm Group formed two distinct mappable units and commented that the group required redefinition. The ophiolitic rocks had been previously termed the Betts Cove Complex (Church and Stevens, 1971; Upadhyay et al., 1971), the Betts Cove ophiolite (Dewey and Bird, 1971), the Betts Cove subgroup (Riccio, 1972) and the Nippers Harbour Group (Baird, 1951; Schroeter, 1971). The term Betts Cove Complex is used herein for the ophiolitic rocks. The Betts Cove Complex is here removed from the Snooks Arm Group and formally defined; the Snooks Arm Group is redefined in the next section, and the term Nippers Harbour Group is abandoned.

The Betts Cove ophiolite contains lithologic units very similar to those of the Advocate and Point Rousse Complexes; hence, the following descriptions summarize only the more striking features of the ophiolitic members at Betts Cove. Rock descriptions are similar to those of the other complexes, to which the reader is referred for more details. I have not mapped in this area; thus, the following descriptions are taken from Riccio (1972) and Upadhyay (1973) for the Betts Cove area, with supplemental descriptions of the area southwest of Nippers Harbour from Schroeter (1971) and DeGrace et al. (1976).

Ultramafic Rocks

Ultramafic rocks of the complex have a maximum thickness of approximately 750 m in the area southwest of Betts Big Pond (Upadhyay, 1973). Upadhyay (1973) noted that this is the only area where the ultramafic member displays primary features and the least alteration. He described the extent and character of the ultramafics to the northeast as follows:

> Between Betts Big Pond and Red Cliff Pond, the Ultramafic Member has a consistent width of about 150 m. Layers, with a southeasterly dip of 45° to 65°, are seen immediately north of Betts Big Pond. The Ultramafic Member, between Red Cliff Pond and the central part of Long Pond, is somewhat narrow and discontinuous. It bulges out to a maximum width (?thickness) of about 400 m in the Tilt Cove area and then gradually thins out towards Beaver Cove Pond....
>
> The ultramafic rocks between Betts Pond and Beaver Cove show a partial to complete alteration to serpentinite and talc-carbonate rock. The margins of the ultramafic belt are generally sheared and consist of talc schist....
>
> Serpentinite occurs as greenish-black slickensided pods and lenses within the talc-carbonate rock. The only fresh ultramafic rock in the Tilt Cove area is a grey pyroxenite exposed near Supply Pond. This rock, occurring at the top of the Ultramafic Member, is a coarse-grained clinopyroxenite with minor patches of gabbroic material.

In addition, Schroeter (1971) noted minor occurrences of highly serpentinized ultramafics in the area of Green Head - Northwest Arm to the southwest of Betts Cove.

All of the fresher ultramafic rocks in the Betts Cove area appear to be cumulates, and include dunite, pyroxenites and wehrlite, with subordinate harzburgite and minor lherzolite (Upadhyay, 1973). Riccio (1972) recognized a layered zonation of these rock types in the Betts Cove area. He noted that, here, the lower, more northwesterly section of the ultramafic member is composed of repeated cycles of layered dunite-harzburgite ± orthopyroxenite, the middle section dominantly contains cycles of dunite-orthopyroxenite-websterite ± harzburgite, and the upper portion is formed of sequences of dunite - wehrlite - olivine clinopyroxenite - clinopyroxenite. These sequences of cyclic layers appear to overlap somewhat and are not strictly confined to any one portion of the ultramafic member.

Upadhyay (1973) described the physical characteristics of these layered sequences as follows:

> The thickness of layers varies from about one millimeter to several meters... The thick layers show further but less well-defined finer laminations within them. The layers are quite consistent in thickness except for the extremely fine laminae that gradually die out along strike. The layers are defined by texture, and also by the smoothness, depth, and colour of weathering. The contacts between any two adjacent layers are generally very sharp. In some cases, however, a gradually decreasing proportion of any particular mineral gives rise to another rock type with a gradational contact.
>
> On the upper level of the layered sequence the ultramafic rocks show superb cumulus structures and textures. Size-graded layers... occur to the south of Kitty Pond (Plate 5-5); pyroxene crystals get gradually smaller within a vertical distance of 0.5 m or less.... Reverse gradation, i.e. coarse crystals on top of fine at the bottom, also occurs. Cross lamination in the ultramafic rocks occurs at several places. The cross laminae truncate abruptly against the main layering. The laminae make an angle of up to 30° with the layers and, in some cases, are concave upwards.... Slump and primary pinch-and-swell structures are also seen in the layered sequence.... In these, the individual layers pinch and swell or abruptly merge into another rock type along strike.

DeGrace et al. (1976) also noted rhythmic layering, locally size graded, of pyroxenite with minor peridotite in the Rogues Harbour area to the southwest and in the ultramafic xenolith in the Cape Brulé porphyry at South Yak Lake.

In addition to layered ultramafics, Upadhyay (1973) described irregular zones of ultramafic breccia, up to 150 m across, just west of Kitty Pond as follows:

> ... (The breccia) shows gradational contacts with the surrounding ultramafic rocks. Layers in the latter can be seen merging transitionally into the breccia along strike. Where giant fragments of serpentinite are enclosed within the breccia, the groundmass assumes a ramifying vein-like pattern. The breccia is variable in appearance and texture. The fragments, which are mostly angular to subangular, range in size from a few millimeters to over half a meter; sorting is very poor. The coarser types are

Plate 5-5: *Size-graded layers of websterite from the area south of Kitty Pond; Betts Cove Complex.*

poorly cemented and have a rubbly appearance. The fragments consist mostly of smooth, saffron to reddish-brown weathering serpentinite but some lumps of rather fresh cumulate peridotite also occur. The fragments are embedded in a finer material of similar composition. No non-ultramafic material was observed in the breccia. It lacks any banding or planar structures.

Extensive serpentinization has largely obscured any primary textures of the rock....

Its occurrence as isolated pods, its gradational contacts with layered ultramafic rocks, virtual absence of any non-ultramafic material, and absence of direct spatial relation with any fault zone suggest that the breccia is of igneous origin.

Upadhyay (1973) also reported that veins and dikes of pegmatitic pyroxenite constitute a very minor portion of the ultramafic member.

Serpentinization appears to be pervasive in the Betts Cove ultramafic member (Upadhyay, 1973), though the area around Betts Cove is less altered than the area between Betts Big Pond and Tilt Cove. Upadhyay (1973) noted that serpentinization is more intense in orthopyroxene than in clinopyroxene. Riccio (1972) carried out X-ray diffraction studies and determined that the serpentine mineral is dominantly antigorite, though lizardite is ubiquitous, but forms less than 30% of the serpentine. Both workers as well as Neale (1957) reported that steatitization and carbonatization are common in the ultramafic rocks northeast of Betts Big Pond. Schroeter (1971) reported extensive serpentinization as well as quartz ± fuchsite ± chlorite-carbonate alterations of the ultramafics in the Northwest Arm area.

Riccio (1972) also noted that serpentinization of the ultramafic rocks in the Betts Cove area is fairly uniform, and does not appear to be affected by nearby intrusive rocks (Cape Brulé porphyry); thus, he suggested that the alteration was an autometamorphic process.

Gabbro

The gabbroic member consists of two units, a basal zone and an upper zone, that are in gradational contact. Upadhyay (1973) noted a combined maximum thickness of 330 m for these units. The basal zone typically consists of pyroxenite (Plate 5-6) and gabbro that grades vertically into the upper zone of predominantly leucogabbro. Upadhyay (1973) described the components of the basal zone as follows:

The two [pyroxenite and gabbro] are either interlayered or, less commonly, occur as diffuse pods of one into the other.... The layers range in thickness from a few centimeters to over a meter. A gradually decreasing concentration of clinopyroxene in gabbro gives rise to a rock consisting of nearly 100 percent plagioclase. The transitional zone is much narrower (30-40 m) in the vicinity of Kitty Pond than in the west.... Layering in the transitional zone is parallel to that in the underlying Ultramafic Member.

Isolated, generally lens-shaped, brown-weathering ultramafic pods occur in the transitional zone. They range in size from less than a meter to tens of meters. The contact between these pods and the surrounding gabbro/clinopyroxenite, where exposed, is generally sharp with no evidence of contact metamorphism or chilling on either side. Some of these, particularly those to the west of Joey's Pond, have their longer dimensions parallel to the layering in the associated rocks and a number of them, having nearly the same thickness, seem to be confined to a particular ultramafic layer. There are two possible explanations for the origin of these pods: (1) they may represent xenoliths brought up from the underlying Ultramafic Member, or (2) they were formed either by pinching-and-swelling in the pre-consolidated magma or by the aggregation and subsequent crystallization of ultramafic material as pods within the magma chamber. It is quite possible that pods of both origins exist.

Upadhyay (1973) further discussed the pods and favored the latter interpretation because they occur within a layered sequence and he found it difficult otherwise to explain their

Plate 5-6: *Pyroxenite that forms the basal zone of the gabbro member at Candlemass Head, Betts Cove Complex.*

emplacement in the sequence. Riccio (1972) and Church and Riccio (1974) interpreted these pods as xenoliths.

The upper leucogabbro portion (containing minor diorite) is very similar to that of the Advocate and Point Rousse Complexes, though available descriptions indicate that it is less altered in the Betts Cove Complex. Upadhyay (1973) also noted the local presence of gabbroic breccias. The gabbro is locally crosscut by komatiitic diabase dikes, and DeGrace et al. (1976) reported pyroxenite dikes cutting the gabbro at South Yak Lake. Where the gabbro is in close association with ultramafic rocks, it is locally rodingitized (Riccio, 1972).

Sheeted Dikes

The sheeted dike member consists of diabasic, picritic, perknitic, and silicic dikes, dike breccia, and screens of granodiorite, gabbro and ultramafics. Diabase dikes form most of this member. Upadhyay (1973) described the general features of the sheeted dikes as follows:

> The exposed maximum width of the sheeted complex, measured across the trend of the dykes, is about 4 kilometers. Its thickness within the ophiolitic suite varies from 400 meters to about 1.6 kilometers. The trend of dykes, in a broad way, is nearly at right angles to the layering in the underlying Ultramafic and Gabbroic Members, although there are local exceptions. The dip of the dykes is vertical or subvertical but shallower dips (ca. 50°) have been observed. At their base the sheeted dykes locally have a sharp contact with the Gabbroic Member but swarms of dykes also continue down into the gabbro and ultramafic rocks. Dykes within the Ultramafic Members are best seen to the south of Kitty Pond and Burtons Pond.
>
> ... The contact between the sheeted complex and the overlying pillow lava does not follow a generally north-south stratigraphic trend in the Betts Cove area but shows "indentation" of one into another. This is particularly obvious in the Betts Cove - Betts Cove Mine - Fault Cove area. The sheeted nature of the dykes ceases at the contact with the Pillow Lava Member. In the contact zone, the proportion of dykes varies from about 5 to 40 percent. The transition between the two is best seen along the coast between Betts Head and Betts Cove where the dykes stand out as vertical stripes in a pillow lava background...
>
> The average thickness of dykes within the sheeted complex is about 50 centimeters, although single dykes up to 6 meters thick have been observed in some places. Except where sheared, chilled margins are shown by nearly all the dykes.

He also noted a change in trend of the dikes in the Kitty Pond - Betts Cove area (Upadhyay, 1973); dikes to the southwest of this area have an easterly trend, whereas those to the northeast trend approximately 120°. Local, abrupt changes in orientation have also been reported (Upadhyay, 1973).

Upadhyay (1973) and Coish (1977b) both recognized diabase dikes and perknitic rocks in the sheeted section and Coish (1977b) identified picritic dikes in the member. Coish (1977b) described all of these rock types as follows:

> Diabase dikes [Plate 5-7] are the most common and the latest set of intrusions in the sheeted dyke member. They cut all other types of dykes and often branch and vein older dykes and gabbro in random directions...They range from 15 cm to 0.5 cm in width... The diabase dykes comprise nearly equal proportions of clinopyroxene/actinolite and albite.

These dikes are brecciated in many places, both along and across dike trends; the breccias constitute both veins and irregular patches (Upadhyay, 1973). Coish (1977b) continued as follows:

> Porphyritic [picrite] dykes, which weather to a rusty red colour, occur predominantly in the lower parts of the sheeted dyke member and often cut the gabbro member.... They are wider than the diabase dykes, usually averaging 1 to 1.5 meters. Two phenocryst phases, which may occupy up to 50% by volume of some specimens, are present in the porphyritic dykes: (1) Clinopyroxene, partially replaced by actinolite, either along or in clusters with individual crystals up to 1 mm long; (2) Chlorite-serpentine patches with a typical olivine crystal form.
>
> Perknite rocks... are very coarse grained and consist of actinolite, chlorite and clinopyroxene with less than 5% felsic minerals. At Betts Cove, the green-grey weathering perknite rocks are usually found near the base of the sheeted dyke member. In some instances, the 'dykes' may be discontinuous along their strike whereas, in other cases, they may be traced, and retain a uniform width, over considerable distances. This irregularity in outcrop pattern and the lack of chilled margins make it difficult in the field to determine whether the perknite material is intrusive like the other dykes or represents country rock which has been thoroughly invaded by diabase and picritic dykes.
>
> Actinolite and chlorite are the principal minerals present. Clinopyroxene occurs as unreplaced relicts as cores to actinolite crystals..

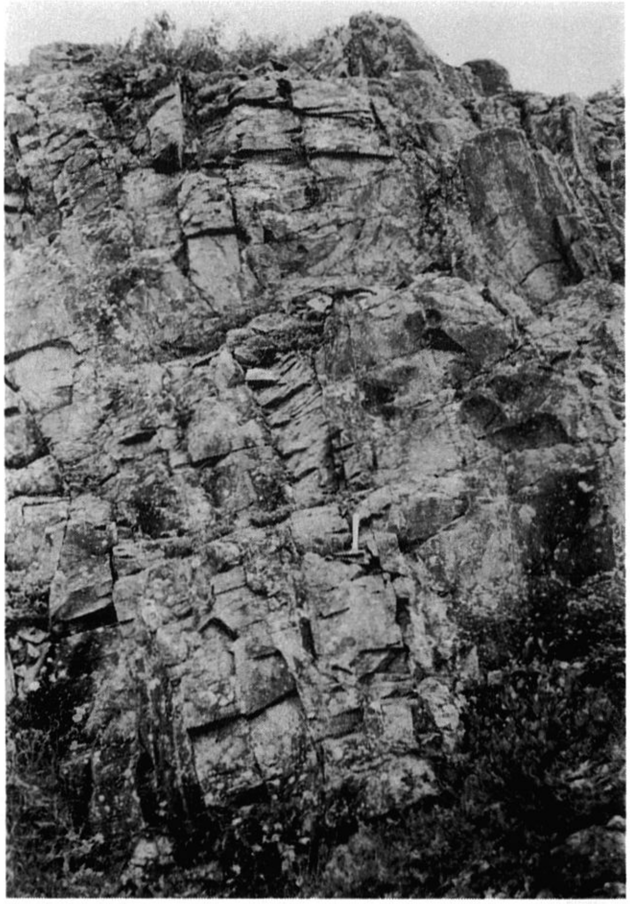


Plate 5-7: *Sheeted diabase dikes in the area south of Kitty Pond; Betts Cove Complex.*

Inland, just west of Betts Cove, a perknite dike is intruded and chilled against a diabase dike (Plate 5-8), indicating that at least some perknites are intrusive. Coish's (1977a) geochemical investigation suggested that the perknite rocks most likely represent screens of an older layered sequence that was subsequently intruded by the gabbro member and sheeted dikes.

Plate 5-8: *Perknite-like dike (coarser grained) chilled against diabase dike to left; note chill margin of perknite-like dike just to left of pocketknife. From the area just west of Betts Cove; Betts Cove Complex.*

Upadhyay (1973) reported dikes of albite aplite that, in places, grade into diabase along the length of the dikes. He also noted numerous gabbro and ultramafic screens in the sheeted dike member; at one locality east of Burtons Pond, Upadhyay (1973) reported a screen of granodiorite more than 30 m long, apparently associated with gabbro.

Pillow Lava

The pillow lava member was examined by both Upadhyay (1973) and Coish (1977b); Upadhyay described the general distribution of the unit as follows:

> The Pillow Lava Member stretches from Betts Cove to Tilt Cove... [and] has a maximum thickness of 1500 meters in the Fly Pond (Betts Cove) area. It shows a slight thinning towards the east and is abruptly reduced to a thickness of merely 40 meters between Long Pond and Tilt Cove... Farther east, in the vicinity of Tilt Cove mines, it again attains a thickness of about 300 meters.

Mafic lavas compose more than 95% of the pillow section; the remaining lavas are ultramafic (Upadhyay, 1973). Coish (1977b) recognized and described two apparently unmappable divisions of mafic lava in the Betts Cove as follows:

> The lower pillow lava member... has pillows which are variolitic and only occasionally vesicular. Very little interstitial material occurs. Clinopyroxene is found as a phenocryst phase, as laths in the groundmass of the cores of pillows... and commonly as radiating spherulitic crystals. Olivine, pseudomorphed by chlorite, and commonly containing minute chromite inclusions, is usually present as a phenocryst phase. Plagioclase, invariably replaced by albite, is always interstitial to clinopyroxene laths. Pyrite is the only opaque mineral recognized.
>
> The upper pillow lava member comprises nonvariolitic, vesicular pillows averaging 0.75 m by 0.25 m in size. Phenocryst phases include clinopyroxene and plagioclase, but not olivine. Plagioclase and clinopyroxene also occur as laths in the groundmass where they are partially or totally replaced by albite and actinolite/chlorite, respectively. Often plagioclase and clinopyroxene form spherulitic fan structures. Magnetite, usually replaced by hematite, and sphene are accessory minerals. Altered interstitial mud and chert occur between pillows.

Coish (1977a) also reported that these units are geochemically distinct. Coish and Church (1978) recognized a narrow (150 m) zone of intermediary pillow lavas between the two pillow zones in Betts Cove, with a mineral content similar to the upper lavas, but physically and geochemically similar to the lower lavas.

The ultramafic pillow lavas occur in close association with the mafic lavas in the area north-northeast of Betts Big Pond (Upadhyay, 1973). Upadhyay (1973) described the relation of these lavas to surrounding rocks and their character at a small pond in this area as follows:

> At Partridge Point a sheeted sill complex is transitionally overlain by an assemblage of mafic-ultramafic pillows. The ultramafic pillow unit, about 12 meters thick, is gradationally overlain as well as underlain by amphibole-rich mafic pillows. Along strike the ultramafic pillow unit also shows a gradation into the massive reddish brown ultramafic sill... The pillows range in size from a few centimeters to about half meter. On fresh surfaces they have a creamy white colour. They consist chiefly of tremolitic amphibole with some talc and are classified as perknite. Reddish brown chromite occurs as an accessory. The rock contains scattered phenocrysts of amphibole, 2 to 3 millimeters across, that have very diffuse and irregular outlines. These ultramafic pillows, with gradually increasing content of felsic minerals, grade first into the "related" felsic perknite pillows and then into the mafic pillows.

In addition to pillow lavas, Upadhyay (1973) described sediments, pillow breccia, and sills associated with the pillowed member. The sedimentary unit extends from Long Pond southwestward to the pond immediately southwest of West Pond; it is composed of approximately 60 m of thinly bedded argillite. Upadhyay (1973) reported that the pillow breccia unit occurs near the base of the pillowed member and is very irregularly developed, varying from a few metres to >50 m in thickness. It appears to be more abundant in the Tilt Cove area than the area to the southwest. Upadhyay (1973) described the sills as follows:

> The Pillow Lava Member contains innumerable sills of variable thicknesses throughout its length between Betts Cove and Tilt Cove. These sills, as observed in the Betts Cove area, represent the continuation of the sheeted dykes into the pillow lava and differ from them only in being somewhat thicker and parallel to the stratigraphic trend of the Pillow Lava Member....
>
> The sills are primarily diabasic and ultramafic (perknitic) in composition, with a mineralogy and texture quite similar to those of the corresponding rocks in the Sheeted Dyke Member.

DeGrace et al. (1976) mapped the pillowed member to the southwest of Betts Cove, but they noted that:

> Pillow lavas are rare and were only noted in an area about two miles north of Stocking Harbour. Interbedded mafic volcanic sediments (metamorphosed by the intrusion of the nearby Burlington Granodiorite) are exposed along a ridge just north of Rix Cove.

Contact Relationships

Contacts within the complex are all reported to be, at least locally, transitional except for those between the ultramafic and gabbroic members. The nature of this latter contact is in dispute; Upadhyay et al. (1971) and Upadhyay (1973) reported that it is transitional whereas Riccio (1972) and Church and Riccio (1974) considered it to be a major discontinuity whereby the gabbro intruded the ultramafic assemblage. The latter workers cite (i) injection features of pyroxenite into the ultramafics, (ii) xenoliths of ultramafic rock within the gabbroic member, and (iii) the disparate compositions of clinopyroxenes from the gabbroic and ultramafic members as evidence for their interpretation. Quite likely, as proponents of both interpretations have suggested, this contact varies from transitional to intrusive along strike (Upadhyay, 1978b; Church, 1977); the local intrusive contacts would indicate slightly later intrusion of gabbroic magma than that which formed the immediate ultramafic sequence.

The contacts of the complex with surrounding units are variable. The complex is conformably overlain by the Snooks Arm Group to the southeast, unconformably overlain by the Cape St. John Group at Pinnacle Point and at Rogues Harbour (Neale, 1957; Schroeter, 1971; Neale et al., 1975; DeGrace et al., 1976; Williams et al., 1977), and intruded and hornfelsed by the Cape Brulé porphyry and the Burlington Granodiorite southwest of Pittman Bight. Along the arcuate outcrop belt between Burton's Pond and Tilt Cove, the complex is faulted against the Cape Brulé porphyry and the Cape St. John Group.

PACQUET HARBOUR GROUP

The Pacquet Harbour Group is here defined as the moderately to steeply northerly dipping sequence of variably deformed and metamorphosed mafic volcanic and volcaniclastic rocks, felsic volcaniclastic rocks, and mafic dikes that outcrops in the north-central portion of the Baie Verte Peninsula (Figure 1-1). The name of the group is inappropriate in some respects, since the rocks at Pacquet Harbour are not representative of the whole group, but the term is maintained because of its common usage since 1969. The unit roughly encompasses the Rambler Mine - Pacquet Harbour area originally outlined by Church (1969), but also includes the lobe of mainly mafic rocks that extends to the southeast of this area, into the region of Northeast Pond and South Yak Lake. The base and the top of the group are not exposed and the thickness is unknown due to a combination of poor exposure, complex structure, and complex stratigraphy; the maximum outcrop width across the group is approximately 15 km, though undoubtedly this does not represent the true thickness of the sequence. A type section for the unit has never been erected, and its choice is herein deferred pending future detailed stratigraphic studies of the group. Instead, the section through the group along the La Scie highway is chosen as a reference section since, here, most constituent rock types are exposed and easily accessible. The unit is host to four massive sulfide bodies that have been mined.

The Pacquet Harbour Group is inhomogeneously deformed and metamorphosed. North of the La Scie highway, the group is polydeformed and polymetamorphosed, displaying as many as three fabrics and reaching lower amphibolite grade metamorphism; south of the highway, this tectonic style gradually changes; here the group generally displays a single penetrative fabric and greenschist grade metamorphism. Portions of the group peripheral to the Burlington Granodiorite have been thermally metamorphosed to amphibolite grade for up to 0.5 km from the contact. One striking structural feature of the group is a strong mineral, clast and pillow lineation (see Chapter VII) displayed by all its northerly members.

Church (1969) first recognized the dominantly mafic succession of rocks between the Burlington Granodiorite and Pacquet Harbour and termed it the Pacquet Harbour Group. He separated it from Baird's (1951) original Baie Verte Group. He considered these rocks to be an easterly portion of his Fleur de Lys Supergroup (Church, 1969). DeGrace et al. (1976) noted that the group is lithically distinct from large portions of the supergroup, and that it has a structural history different from that of the western Fleur de Lys Supergroup. Hence, they suggested that the definition of the supergroup as used by Church (1969) was not valid. Williams et al. (1977) agreed, and also separated the Pacquet Harbour Group from the Fleur de Lys Supergroup. Gale (1971, 1973) was the first to consider the group as partly ophiolitic, based on geochemical studies. This, in conjunction with rock types present and regional correlation, is the main criterion for considering these rocks as ophiolitic in the present report.

Some of the rocks included in the Pacquet Harbour Group of this report have been considered as part of the Lush's Bight Group (Neale and Kennedy, 1967) and the Rambler Group (Bird and Dewey, 1971). In addition, several writers previously included some of the mafic rocks of the Cape St. John Group (Neale, 1958b; DeGrace et al., 1976) within the Pacquet Harbour Group (Neale and Kennedy, 1967; Phillips et al., 1969; Church, 1969; Coates, 1970; Dewey and Bird, 1971; Kennedy, 1973, 1975a; Tuach and Kennedy, 1978). Traditionally, metamorphosed mafic rocks at Pelée Point were included with the Pacquet Harbour Group; in this report, these rocks are included in the Ming's Bight Group (see Ming's Bight Group, Chapter IV, for discussion).

The internal stratigraphy of the Pacquet Harbour Group is uncertain due to poor exposure and complex structural and stratigraphic relationships. Geochemical studies by Gale (1971, 1973) indicate that the group is composed of two lava types, tholeiitic and komatiitic basalt. Initially, Gale (1971, 1973) interpreted the tholeiites as typical mid-ocean ridge basalts; but later he noted that the exact tectonic environment of the tholeiites could not be deduced from the geochemical data, although he still supported an ocean ridge origin for the rocks (Gale, personal communication, 1982). In the present study, the basaltic komatiites, i.e. boninites (see Chapter VI), are considered correlative with similar ophiolitic lavas of the Betts Cove and Point Rousse Complexes; hence, the Pacquet Harbour lavas most likely represent, at least in part, the top of an ophiolite sequence.

Tuach and Kennedy (1978) attempted to synthesize the stratigraphic succession of the group near Rambler Mines. Based on local north younging strata in the area, they suggested that the northerly portion of the group is a continuous, north facing sequence. However, they noted the major uncertainties of their model, including poor exposure, sparse younging direction indicators, and the occurrence of medium to large scale tight to isoclinal folds in the area. No reliable marker units have been identified in the group. In addition, I have noted local small scale examples of extreme layer-parallel transposition in northerly portions of the group, which suggest that a reliable stratigraphy cannot be established in this area. Hence, in this study, the stratigraphy of the group is considered to be unknown and the rocks are described. as strictly lithic units, including (i) pillow lava, (ii) felsic volcaniclastics, (iii) mafic epiclastics and volcanics, (iv) unseparated volcanics and volcaniclastics, and (v) gabbro and diabase. A sixth division, composed of ultramafic rocks, is areally limited and has been seen only in drill core; it is described with the unseparated rocks, below.

Pillow Lava

Large areas of pillow lava are common south of Rambler Pond, along the La Scie highway, and west of South Yak Lake (Figure 1-1); smaller areas of pillow lava also occur in the unseparated rocks of the group. The pillows appear to be interlayered with all other members of the group.

The two geochemical types, *viz.* tholeiitic and high magnesian (see Chapter VI) are indistinguishable in the field[1]. Based on geochemical analyses, the tholeiitic lavas appear to be confined to the area of the La Scie highway, and the high magnesian lavas appear to compose most of the pillow lava zones south of the Rambler Main Mine. The nature of the contact between the two types in the area of the Main Mine is uncertain. One mafic crystal tuff north of the La Scie highway has high magnesian series geochemical attributes (sample 1340007, Appendix III), which suggests that local pillow lava lenses in this area are highly magnesian.

All of the lavas are typically pale green weathering, blackish green basalts that are locally amygdular with pillows up to 2 m in diameter, but generally less than 1 m across (Plate 5-9). Locally, either carbonate or chert forms interpillow material. In some places, such as west of South Yak Lake, pillow breccia is intimately associated with the pillow lavas. Commonly, pillow lavas in the southern portion of the group are variolitic; in one place, they contain varioles up to 4 cm in diameter.

The lavas are composed predominantly of actinolite, albite and epidote with subordinate chlorite and quartz. Chlorite is a significant component in lavas to the north. Generally, the only original textures preserved in these rocks are plagioclase phenocrysts and amygdules. Tuach (1976) summarized the character of the major minerals in the Rambler area as follows:

The actinolite is pale green and slightly pleochroic; it occurs as the dominant mineral in the fine grained matrix (0.1 mm) and as porphyroblasts (0.5 mm). The actinolite crystals are rarely euhedral and usually exhibit fibrous terminations, and porphyroblasts are commonly twinned. Actinolite also occurs as a minor constituent in veins and amygdules. Blue-green hornblende was recorded in several pillows by Gale (1971).

Plagioclase (An_2 - An_{10}) (Gale, 1971) occurs as relict phenocrysts (0.5 mm), as polycrystalline aggregates in amygdules, and in veins; the crystals are partially or completely altered to epidote, and commonly show polysynthetic twinning; partial replacement is controlled by composition planes which result in straight trails of granular epidote within one plagioclase crystal.

Epidote occurs as an alteration product in feldspar crystals and as individual crystals or aggregates of crystals (0.1 mm) in the matrix; it also occurs in veins and amygdules. The minerals epidote and clinozoisite are most common; minor zoisite is present. Epidote minerals average less than 10% in the rock, although locally 75% epidote is present.

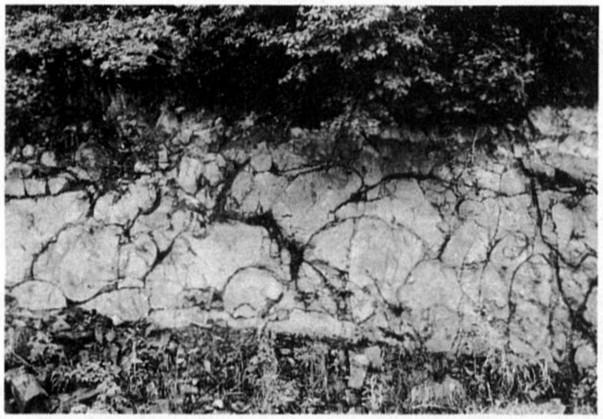


Plate 5-9: *Typical pillow lavas of the Pacquet Harbour Group along La Scie highway directly northeast of Side Pond.*

Felsic Volcaniclastic Rocks

Gray to cream weathering felsic volcaniclastic rocks form a conspicuous bull's-eye shaped map pattern north of Rambler Pond (Figure 1-1); sporadic thin belts of felsic rocks occur both eastward along strike and in the southerly portion of the group, west of South Yak Lake (Figure 1-1). The rocks appear to be dacitic, but are best termed keratophyres because of their tectonized nature. They vary in color on fresh surfaces from gray-green to reddish purple and, locally, black. Characteristically, the only primary features other than layering are fragment outlines; the fragments range in size from lapilli to blocks averaging 15 cm or less in diameter, though locally blocks up to 60 cm in diameter have been reported (Tuach and Kennedy, 1978). Tuach (1976) reported local mafic blocks from within the felsic fragmental rocks. The coarsest deposits occur in the Rambler area, whereas lapilli tuff predominates in other areas. Gale (1971) reported minor flows from the Rambler area; however, Tuach and Kennedy (1978) proposed that the massive appearance of these rocks is due to tectonism, and that vague fragment outlines are visible on close inspection.

[1] Since the writing of this section, A. Kerr has noted that, in extensive new outcrops in the southern portion of the group, the high magnesian lavas weather bright emerald green in contrast to the pale green weathering tholeiites (A. Kerr, personal communication, 1982).

Some massive layers west of South Yak Lake and south of Gull Pond may represent flows, though alternatively they could have originated as felsic sediment or tuff. These layers, up to 3 m thick, contain small globules, up to 2 cm across, of diabase and many fragments of amphibole, apparently pseudomorphic after pyroxene crystals. One of these massive layers is associated with a thin bed (up to 10 cm thick) of epidote-quartz-magnetite rock that probably represents a chemical sedimentary deposit. South of Gull Pond, another massive felsic layer, approximately 2 m thick, is cut by a bright emerald green mafic dike, approximately 6 to 8 cm wide. Based on its distinct weathering color, the dike is probably related to high magnesian lavas in the area.

Both the matrix and larger fragments of the felsic volcaniclastics commonly display quartz "eyes" up to 5 mm across, and small lenses and stringers of actinolite give the rock a mottled appearance in many places. Tuach and Kennedy (1978) suggested the quartz eyes in the Rambler area represent porphyroblasts; the prevalence of this feature in lower grade rocks of the southern portion of the group suggests to me that it represents original phenocrysts. In addition to the fragmental felsic rocks, thin (< 6 cm) felsic layers, probably either tuff or reworked tuff, are commonly interlayered with the mafic rocks in areas peripheral to major felsic outcrop areas. These tuffaceous rocks are most common along South Brook, north of South Brook Pond. At one locality, southeast of Gull Pond, gray quartz "eye" felsic rock forms the matrix to a mafic fragmental rock containing angular clasts up to 4 cm across.

The main mineral constituents are quartz, plagioclase, and epidote with subordinate biotite and minor chlorite, clinozoisite, muscovite and amphibole. Quartz and feldspar phenocrysts locally form up to 25% of the rock, and locally the feldspar is altered to granular epidote and clinozoisite. The feldspar appears to be mostly plagioclase in the albite-oligoclase range, based on optical methods. Biotite generally forms less than 5% of the rock and defines a planar fabric. The amphibole associated with the felsics appears to be actinolite.

Mafic Volcaniclastic and Epiclastic Rocks

These rocks were distinguished as a member of the group in the area immediately west of South Yak Lake by DeGrace et al. (1976) (Figure 1-1). They occur throughout the unit, though more detailed mapping is necessary to separate them. DeGrace et al. (1976) described the rocks in the South Yak Lake area as thickly bedded, reworked green tuff with minor chert. A similar belt of volcaniclastics, unseparated on Figure 1-1, outcrops parallel to and north of the La Scie highway and east of Rambler mine site. In this broad belt, medium to thickly bedded, reworked tuff and volcanic graywacke locally display spectacular flame structures (as on the Woodstock road), crossbeds (Plate 5-10), and sedimentary loading features. Some of the medium bedded rocks are reminiscent of flyschoid deposits. Immediately west of Side Pond, a pebbly conglomerate with rounded graywacke, tuff and chert clasts is interlayered with these epiclastics. Thin members of finely laminated mafic and felsic tuffs, possibly reworked, are interspersed throughout this sequence. Locally, gray to black chert is associated with these rocks. In addition, mafic lapilli tuff and crystal tuff occur inland, north of the road; these strata appear to be pyroclastic. The crystal tuff is of particular interest, as it is distinct and conspicuous and may be useful as a marker horizon throughout this area. It is typically medium bedded and pale green to gray-green with prominent knobby weathering mafic crystals up to 5 mm across; these crystals appear to be amphibole pseudomorphs after pyroxene. Reconnaissance mapping indicates that the tuff forms at least two horizons in the group and occurs at Ming's South Brook, northwest of Side Pond, south of Three Corner Pond, and south of Scrape Pond. Detailed mapping will be necessary to determine the inter-relationships of these units.

Unseparated Rocks

Much of the Pacquet Harbour Group remains unseparated mainly because of abrupt facies changes and poor exposure. The unseparated rocks in Figure 1-1 include all the facies described above as well as massive mafic flows, mafic agglomerate, volcanic conglomerate, and mafic schist. The massive flows are petrographically similar to the pillow lavas and are also amygdular in places. The mafic agglomerate (Plate 5-11) is generally composed of angular to subangular basaltic clasts that range in size up to 60 cm across, set in a fine grained mafic matrix; clasts commonly compose more than 25% of the rock. The mafic clasts are petrographically identical to the pillowed and massive lavas of the group. Locally, felsic clasts are intermixed with the mafic clasts. The agglomerate is most commonly associated with lavas. In places, volcanic conglomerate is associated with other epiclastic and volcaniclastic rocks of the unit (Plate 5-12); the conglomerates are composed mainly of rounded to subangular mafic and felsic clasts, up to 20 cm across, in a predominantly chloritic matrix. These deposits are most common in the South Brook area near Rambler Pond and in the area south of Rambler Pond. The mafic schists, or greenschists, are chloritic rocks in which primary features have been obliterated by a pervasive schistosity. They are common throughout the central and northerly portions of the group and have been attributed to intense deformation of a mafic protolith (Tuach and Kennedy, 1978).

In addition to surface exposures, Tuach and Kennedy (1978) reported ultramafic lenses in the hanging wall of the Ming Mine. The lenses, which range up to 3 m across, occur approximately 250 m above the Ming ore horizon. They are composed of talc, serpentine, carbonate, chlorite and magnetite (Tuach and Kennedy, 1978). Tuach and Kennedy (1978) interpreted these ultramafics as intrusive dikes. Although boninitic basalts do occur, neither ultramafic dikes nor ultramafic lavas have been reported from elsewhere in the Pacquet Harbour Group. In contrast to Tuach and Kennedy (1978), I am impressed by the rareness and tectonic setting of these bodies, and believe that they represent tectonic slivers along layer-parallel and schistosity-parallel faults. They appear to be confined to the hanging wall and do not reach surface. Significantly, the nearest other ultramafic bodies are localized tectonic slivers along the Scrape Thrust, approximately 3 km to the north. Thus, the ultramafic pods may be localized along previously unrecognized subsurface thrust planes, possibly related to the Scrape Thrust.

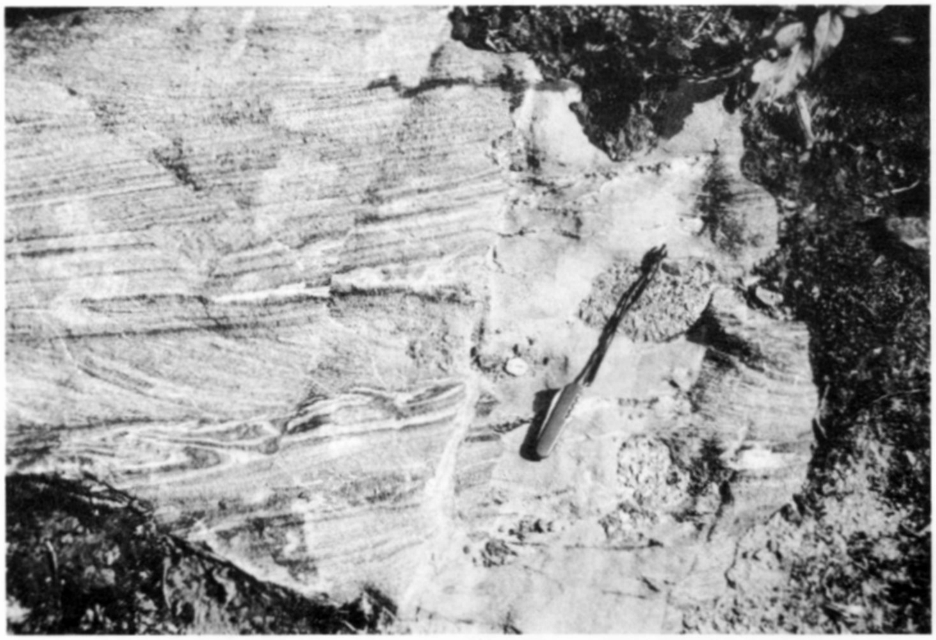


Plate 5-10: *Crossbedding and probable slump folds in upper greenschist grade epiclastic rocks, near the Woodstock road - La Scie highway junction. Note prekinematic dike that crosscuts the primary structures; Pacquet Harbour Group.*

Plate 5-11: *Mafic agglomerate with subangular vesicular fragments, from the inland area due south of Side Pond; Pacquet Harbour Group.*

Mafic Intrusive Rocks

Diabase and gabbro, and equivalent metadiabase and metagabbro in northerly portions of the group, constitute approximately 40% of the exposed rocks of the group and intrude all other constituent members. This apparently high percentage of intrusive mafic rocks probably reflects their resistant nature and, hence, susceptibility to exposure. Large intrusive mafic bodies occur immediately to the northeast and due south of the Rambler Mines property and are separated on Figure 1-1. The margins of the intrusives are commonly unexposed but, in some areas, these intrusives form a swarmlike group, with mafic dikes chilled against each other. This swarming of dikes is prevalent in the area south of Scrape Pond.

The mafic intrusive rocks display a wide variety of textures, including coarse pegmatite (feldspar and amphibole up to 2.5 cm), medium grained phases, fine grained feldspar porphyritic phases, fine grained diabase, brecciated diabase, and large dendritic clusters of amphibole (up to 7 cm long) in medium grained metagabbro. These textural varieties appear to reflect intrusive cooling histories rather than sequential phases of intrusion, since each kind of intrusive is seen to cut the others over the full extent of the group; the only exception is the pegmatitic phase, which I have observed only in isolated occurrences near the base of the Scrape Thrust. These pegmatitic gabbros are locally protomylonitic and occur as pods and lenses up to 5 by 2 m; they strongly resemble the pegmatitic gabbro of the Point Rousse ophiolite to the north. It is possible that these rocks are ophiolite slivers, tectonically localized along the Scrape Thrust zone.

Geochemically, most of the dikes analyzed are tholeiite (Gale, 1971, 1973; this report) but one dike in the Main Mine area is boninitic (Gale, 1971).

In thin section, the mafic intrusives are composed of amphibole, plagioclase and epidote minerals, with subordinate biotite, chlorite, sphene, quartz and opaques. The amphibole ranges from hornblende in the north to actinolite-tremolite in the central and southern portions of the area; it ranges in size from 0.1 to 2 mm in length. The plagioclase ranges in size up to 4 mm long and is variably altered to fine grained epidote; relatively clear plagioclase from the southern portion of the group was determined, by optical methods, to be

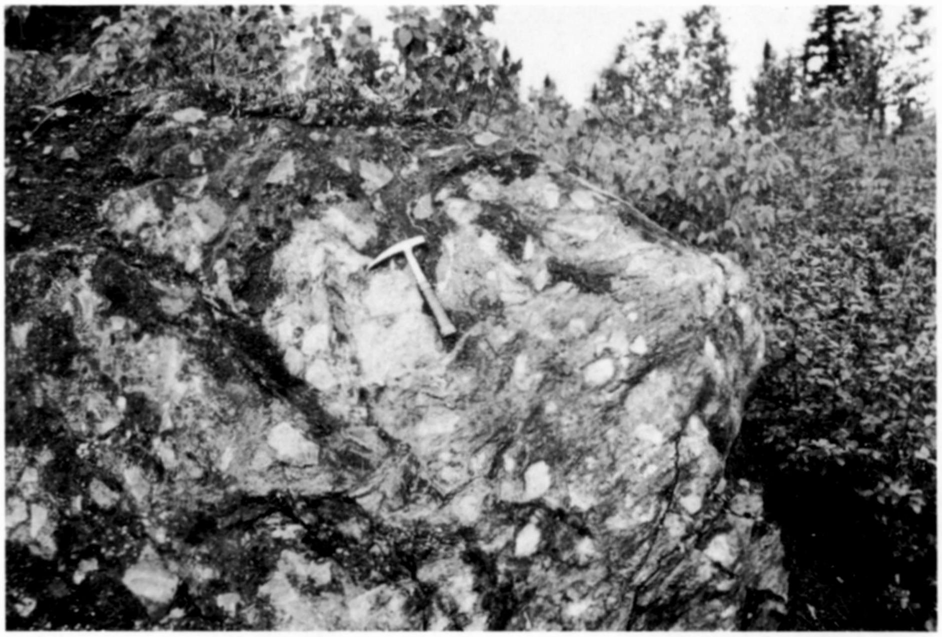


Plate 5-12: *Pacquet Harbour Group volcanic conglomerate composed of angular to subrounded dacitic, dioritic and mafic clasts within a dark green chloritic matrix; from the La Scie highway directly east of the contact of the group and the Burlington Granodiorite.*

andesine. The other minerals are minor constituents, though biotite and chlorite are locally abundant in the northern portion of the group, replacing amphibole.

Contact Relationships

The Pacquet Harbour Group is intruded by the Burlington Granodiorite, the Dunamagon Granite and the Cape Brulé porphyry. Locally, as on the shore opposite Pacquet Harbour, the porphyry is faulted against the Pacquet Harbour Group. The Pacquet Harbour Group is in tectonic contact with the Point Rousse Complex and the Ming's Bight Group. The contact with the Point Rousse Complex is largely defined by the Scrape Thrust and a northeasterly trending fault of uncertain displacement. The tectonic juncture between the Ming's Bight Group and the Pacquet Harbour Group appears to be a zone of faulting that predates regional polytectonism (see Ming's Bight Group and Chapter VII). The contact between the Pacquet Harbour and Cape St. John Groups appears to be unexposed in the area southeast of Rambler Brook (see Cape St. John Group).

Depositional Environment

Based on rock associations and their geochemistry in the Rambler area, Gale (1971) interpreted the group to have been deposited on oceanic crust. The occurrence of dike swarms and the geochemistry of the pillow lavas (see Chapter VI) indicate that portions of the group represent the uppermost part of an ophiolite. In this report, the Pacquet Harbour Group is considered to be geochemically equivalent to the Betts Cove ophiolitic lavas. The disposal of tholeiitic lavas atop high magnesian lavas in the Betts Cove Complex suggests an analogous setting for the Pacquet Harbour Group. Thus, the southern high magnesian parts of the group may represent lower levels of the ophiolitic lava member whereas the tholeiitic lavas along the La Scie highway may represent the upper part of this member. The abundance of both mafic and felsic volcaniclastic rocks in the northern area supports this notion (Tuach and Kennedy, 1978). However, it is noteworthy that a high-magnesian-like dike crosscuts felsic rocks in the southerly part of the group (see Felsic Volcaniclastic Rocks above). This suggests that the felsic rocks either predate or are interlayered with the high magnesian lavas in the area.

Hutchinson (1973) and Tuach and Kennedy (1978) drew attention to the unusual geochemistry and ore mineralogy of the Pacquet Harbour Group rocks and likened them to "primitive" crustal conditions, similar to those of the Archean. The remainder of the group represents continued mafic volcanism on this oceanic crust, which may indicate either nearby island arc or oceanic island volcanism.

DeGrace et al. (1976) considered that portions of the unit west of Woodstock might be subaerial in origin based on fiamme-like features in these rocks. However, metamorphic recrystallization in these rocks is pervasive and deformation is intense; thus, I doubt if the fiamme-like features are of primary origin. In addition, pillow lavas occur directly along strike immediately to the west, indicating a probable submarine depositional environment for all of these rocks.

Age and Correlation

Sangster and Thorpe (1975) reported a tentative lead isotope age of 460 Ma (Middle Ordovician) for the Pacquet Harbour Group, based on galena from the Ming Mine. The ore body lies in mainly felsic and mafic volcaniclastic rocks which probably represent rocks above the ophiolitic component of the group. In addition, all of the group is intruded by the Burlington and Dunamagon plutons, which are approximately 460 Ma in age. The age constraint of the group is similar to that of lithically and geochemically similar rocks of the Betts Cove Complex and the Snooks Arm Group; thus, the ophiolitic rocks of the Pacquet Harbour Group are considered correlative with the Betts Cove lava member and the remainder of the group is probably equivalent to the Snooks Arm Group. Felsic rocks have not been reported from the Snooks Arm Group, although felsic detritus is present (Church, 1969; Upadhyay, 1973).

Neale and Kennedy (1967) noted the similarity between the Birchy Schist and the Pacquet Harbour Group, and DeGrace et al. (1976) considered the group to be partly equivalent to the Cape St. John Group. Correlation of the Pacquet Harbour Group with the Cape St. John Group is herein considered dubious; significantly, the two units appear to be of different ages (see Cape St. John Group), represent totally different depositional environments, and are geochemically unrelated. Both the Pacquet Harbour Group and the Birchy Complex are in part ophiolitic and pre-Middle Ordovician in age; tentatively, they are considered to be at least partial correlatives. The Birchy Schist is considered equivalent to the Pelée Point schist and, hence, the latter unit is also tentatively correlated with the Pacquet Harbour Group.

SUMMARY OF OPHIOLITIC BASEMENT

The ophiolitic rocks of the peninsula are summarized and compared in Figure 5-2. It should be noted that these are composite sections and faulting may have significantly altered original thickness of the various ophiolite members.

The composite sections in the figure conveniently display the distribution of features within the ophiolitic basement. Strikingly, three unusual features occur almost throughout plutonic parts of the basement; namely, (i) ultramafic blocks and pods occur within the cumulate portions of all the complexes, (ii) the sheeted dike member appears to cut down into the cumulate section in both the Point Rousse and Betts Cove Complexes, and (iii) gabbro appears to intrude, at least locally, ultramafic rocks of the Advocate and Betts Cove Complexes. Also, the Betts Cove Complex plutonic rocks appear to lack the high temperature foliation that is found in the lower portions of the Advocate and Point Rousse Complexes, possibly due to an oversight in field work. These features are anomalous in light of the generally held notion of ophiolite stratigraphy (Coleman, 1977).

The geochemistry of the lava members of the Point Rousse and Betts Cove Complexes and the Pacquet Harbour Group is also unusual (see Chapter VI). Essentially, all contain high magnesian lavas that are progressively more magnesian from the westerly Point Rousse Complex to the easterly Betts Cove Complex.

The Pacquet Harbour Group high magnesian lavas appear to be unique among those of the peninsula, as they may either overlie or interdigitate with felsic rocks of the group. These felsic rocks could plausibly represent extrusive trondhjemite in this ophiolitic unit.

The Advocate Complex is anomalous with respect to the other complexes in that it is unconformably overlain by a volcanic-sedimentary sequence. The Betts Cove and Point Rousse Complexes are conformable with overlying strata, and these cover rocks are different from those overlying the Advocate Complex. In addition, the Advocate Complex contains a single sliver of a basal dynamothermal aureole in the Advocate Mines area; this feature is absent from the other ophiolite suites on the peninsula.

The significance of these uncommon features in the ophiolites may be related to their environment of formation. The age of genesis of these suites appears to be approximately synchronous with the Early Ordovician regional obduction of ophiolites onto the ancient North American continental margin (see Chapters VII and IX). Such a process could cause the development of the unusual plutonic stratigraphy of these complexes; and the admittance of large quantities of sea water to the mantle via obduction faults could account for the unusual geochemistry of the lava members (see Chapter VI). The aureole at the base of the Advocate Complex ultramafic body indicates that at least a part of this unit was involved in obduction processes (Bursnall, 1975). Such an event could have led to stripping of the original ophiolitic cover and eventual deposition of a younger cover sequence. The other ophiolites lack an aureole and appear to have retained their original, conformable cover sequences. These suites may have been either just forming or else situated away from the Advocate Complex during obduction processes.

The coincidence of many anomalous features in the ophiolites of the Baie Verte Belt appears to be consistent with the interpretation of these units being derived from a single generation of oceanic crust. The unusual nature of this crust may be due to its genesis near a dynamic zone of obduction.

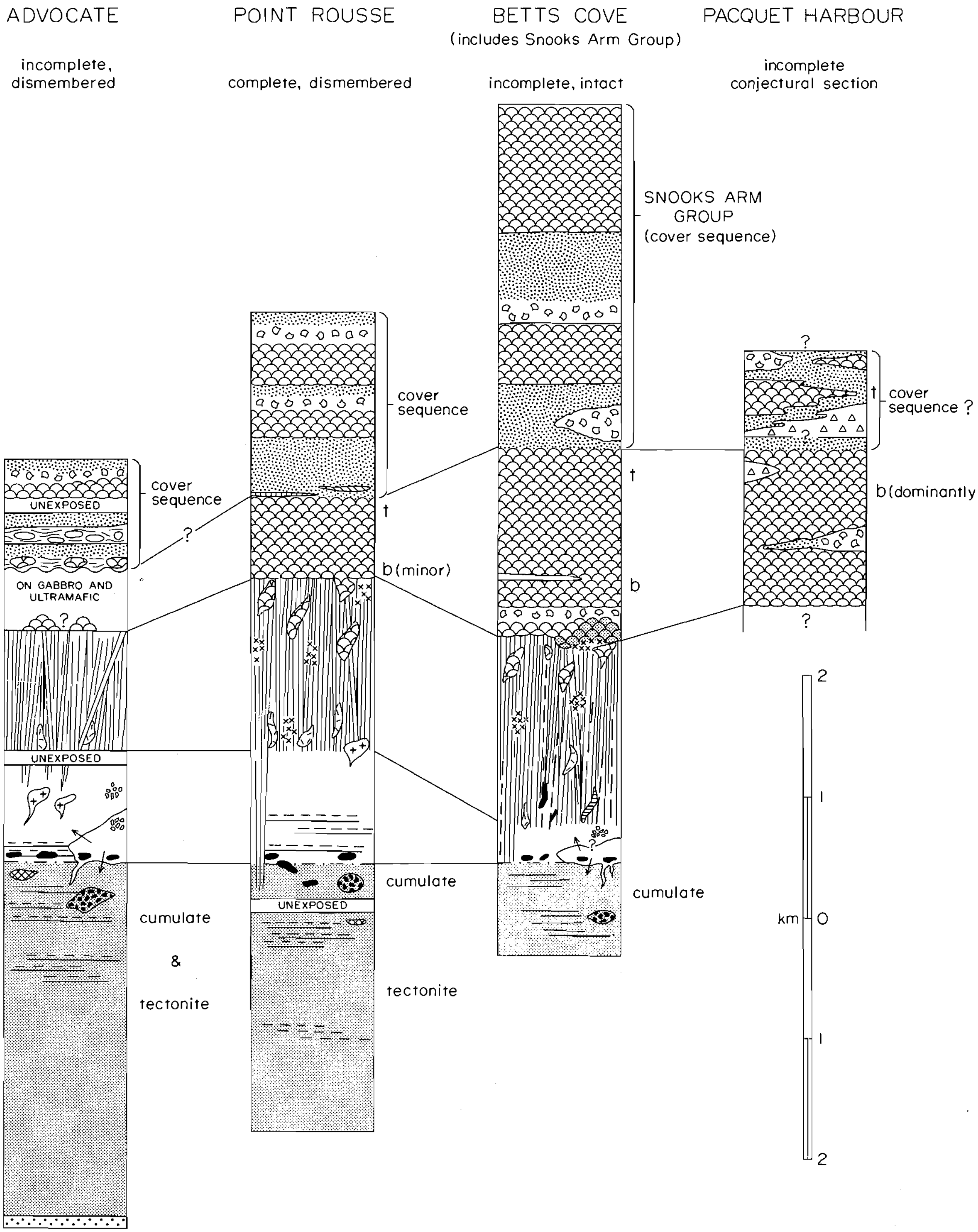


Figure 5-2: *Comparison of sections through ophiolitic units of the Baie Verte Peninsula.*

SYMBOLS

- Boulder conglomerate
- Felsic pyroclastics
- Mainly volcaniclastics
- Argillite
- Iron formation
- Chert horizon
- Ultramafic pillow lava
- Mafic pillow breccia
- Mafic pillow lava
- Screens in dikes (pillow lava, gabbro, ultramafic, granodiorite)
- Ultramafic dikes
- Brecciated diabase dikes
- Sheeted diabase dikes
- Quartz albitite, trondhjemite
- Gabbro breccia
- Mainly gabbro
- Ultramafic pods and lenses
- Chromite pod
- Ultramafic breccia
- Ultramafic
- Garnetiferous amphibolite

b = Boninitic lava

t = Tholeiitic lava

- High temperture foliation
- Compositional layering
- Unconformity
- Transitional contact
- Conformable contact
- Intrusive contact

Late dikes excluded from sections

Thicknesses are based on either maximum preserved thicknesses or outcrop width; original thicknesses were probably greater.

OTHER VOLCANIC COVER SEQUENCES

Volcanic, volcaniclastic and subordinate sedimentary rocks that are separate from and overlie the ophiolitic basement and associated rocks form the remaining volcanic cover sequences in the Baie Verte Belt.

These cover rocks comprise two major divisions that are separated from each other by an unconformity at two locales. The lower division is dominated by the products of mafic submarine volcanism and includes the Snooks Arm and Flat Water Pond Groups; these units are largely equivalent to the cover sequences that have been described with the Advocate and Point Rousse Complexes and parts of the Pacquet Harbour Group. The upper division is characterized by mainly subaerial felsic volcanic and associated rocks, and is composed of the Micmac Lake and Cape St. John Groups. The divisions appear to be separated in time; the lower division is considered to be largely older than Middle Ordovician, with the possible exception of the Flat Water Pond Group, whereas the upper division is probably Siluro-Devonian in age, though it should be noted that the age of the Cape St. John Group has been a major topic of controversy. More detail of the ages of these units is presented below with the description of each group.

The two major divisions of the cover sequence in the Baie Verte Belt reflect a much broader, fundamental division of the Dunnage Zone, as recognized by Dean (1978). The informal lower and upper divisions recognized here are equivalent to Dean's pre-Caradocian and post-Caradocian island arc sequences. Dean (1978) indicated that, in most of the Dunnage Zone, these divisions are separated by an intervening Caradocian argillite and chert unit; this unit is apparently absent from the Baie Verte Belt and the time interval it represents is probably manifested in the unconformity that separates the lower and upper divisions in the belt.

Excluding rocks associated with the ophiolites, I have mapped portions of only two groups in the cover sequence, namely the Flat Water Pond and Micmac Lake Groups. Consequently, the following descriptions rely largely upon the work of Neale (1957, 1962), Upadhyay (1973) and Jenner (1977) for the Snooks Arm Group, Kidd (1974) for detailed work on the Flat Water Pond Group, Kennedy and Neale (1967) and Kidd (1974) for the Micmac Lake Group, and DeGrace et al. (1976) for the Cape St. John Group.

SNOOKS ARM GROUP

The traditional Snooks Arm Group is herein redefined as the four formations of mafic volcanic and volcaniclastic rocks that directly overlie the Betts Cove Ophiolite; the latter, formerly included in the group, is given the status of a complex in this report. The group comprises two mainly volcaniclastic formations, the Bobby Cove and Balsam Bud Cove Formations, and two dominantly pillow lava formations, the Venams Bight and Round Harbour Formations (Upadhyay, 1973). The group is disposed in an easterly trending and plunging syncline (Neale, 1957; Upadhyay, 1973), exhibits slaty cleavage, and is metamorphosed to the lower greenschist facies. The unit conformably overlies the Betts Cove Ophiolite Complex and is unconformably overlain by the Cape St. John Group. The descriptions of rock types in the group are largely from Upadhyay (1973).

Snelgrove (1931) first termed the dominantly volcanic and volcaniclastic rocks between Candlemass Head and Beaver Cove, Notre Dame Bay, the Snooks Arm Series. Douglas (Douglas et al., 1940) followed Snelgrove's terminology, though he noted an unconformity within the group in the Tilt Cove area. This discordancy of lava overlying tilted red slates has not since been recognized by workers mapping the area in detail. Baird (1951) raised the series to group status, and subsequently Neale (1957) informally divided the group into three pillow lava units and two volcaniclastic units. Church and Stevens (1971) and Upadhyay et al. (1971) recognized that ultramafic and gabbroic rocks bordering the group to the north are part of an ophiolite suite, the top of which is the Snooks Arm Group; consequently, they included the ophiolite in the group. DeGrace et al. (1976) noted:

> In the vicinity of Beaver Cove, the rocks immediately underlying the unconformity at the base of the Cape St. John were originally named the Snooks Arm Group by Snelgrove (1931). They were renamed the Beaver Cove Group by Dewey and Bird (1971), a term also adopted by Upadhyay (1973). One of the many minor faults in the area separates rocks of this "group" from rocks of undisputed Snooks Arm Group, and there are no significant differences between them and rocks of the Bobby Cove Formation (Upadhyay, 1973) as regards lithology, chemistry... or structural history (DeGrace et al., 1975; Neale et al., 1975). Because of their contiguity and similarity to Snooks Arm Group rocks, we see no reason whatsoever to assign them to any other rock group, and here restore them to their original place as part of the Snooks Arm.

Upadhyay (1973) thoroughly described the group and divided it into the Betts Cove Ophiolite plus four overlying formations. DeGrace et al. (1976) followed Upadhyay's (1973) nomenclature.

Bobby Cove and Balsam Bud Cove Formations

The Bobby Cove and Balsam Bud Cove Formations were described by Upadhyay (1973) as follows:

> A sedimentary/pyroclastic unit that overlies the Pillow Lava Member of the Betts Cove Ophiolite is here called the Bobby Cove Formation. It has a thickness of about 500 m and extends over the entire distance between Betts Cove and Beaver Cove. Faulting has caused a complex outcrop pattern of this formation in the Tilt Cove area. The best sections occur along the Betts Island - Bobby Cove coastal exposure and along the East Pond - Snooks Arm trail, the latter being more easily accessible.
>
> The Balsam Bud Cove Formation, another predominantly sedimentary unit, is separated from the Bobby Cove Formation by the intervening Venams Bight Basalt. The total thickness of the Balsam Bud Cove Formation is about 750 m. Owing to shallower dips in the west the outcrop is wider there than in the east.
>
> The sediments, pyroclastic and associated igneous rocks are quite identical in both of them.

Upadhyay (1973) also noted that andesitic agglomerate, tuff and flysch are the most common rock types, though graywacke, argillite, chert and conglomerate are also present. He described all of these as follows:

> The agglomerate consists of subangular, rounded or ellipsoidal fragments ranging in size from a few centimeters to as much as half meter. The fragments consist of light gray andesitic rock with pyroxene, and less commonly amphibole phenocrysts up to 1.5 centimeters across.... Bedding within the agglomerate is generally either lacking or poorly developed; however, there are scattered thin units of well bedded graywacke and argillite. At Button Hole Cove, the agglomerate contains large lumps of porphyritic andesite, gabbro, foliated amphibolite, pyroxenite and hornblendite.

...Andesitic tuff is perhaps the most common pyroclastic rock in the Snooks Arm Group.... Bedding in the tuff, although not perfect, is better developed than in the agglomerate ....

Andesitic flysch is intimately associated with, and grades into, the andesitic tuff. It is generally medium-grained and, where well-sorted, resembles a diabase or gabbro.

Graywacke is generally associated with argillite. Thickness of the graywacke units varies from a few centimeters to several meters. Size-graded beds are common [Plate 5-13]. The average grain size is about 1 to 2 millimeters although much larger clasts are relatively common. The grains are generally angular to subangular....

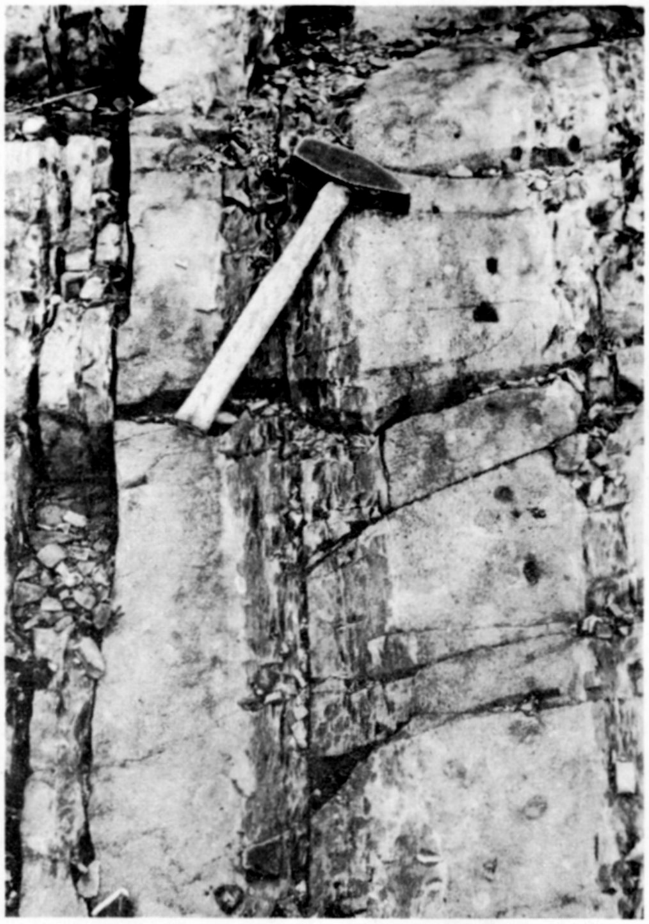


Plate 5-13: *Graded graywacke and argillite beds within the Bobby Cove Formation, Snooks Arm Group, just northwest of Snooks Arm.*

Chert and argillite invariably occur as thinly bedded units. In some parts such beds are only a few centimeters thick and, being of different colors, they produce a ribbon pattern. Cherty argillite and argillaceous chert also occur. The chert is red, maroon, light green, gray or buff....

[Conglomerate and breccia]...constitute a minor part of the Snooks Arm Group. Thickness of conglomerate and breccia units may lie anywhere between one and fifteen meters. They do not show bedding and are poorly sorted. The clasts are generally angular and subangular. Their size ranges from less than a centimeter to as much as one meter. The composition of the clasts and their proportion varies from place to place. A 6-meter conglomerate unit on the Round Harbour road, for instance, contains 80 percent clasts of rhyolitic rocks. Clasts of similar composition, although in lesser amount, also occur in a conglomerate on the northern shore of Snooks Arm....

Locally, the argillites are graphitic and fossiliferous (Snelgrove, 1931; see Age and Correlation). Church (1969) reported mafic and felsic schist fragments as well as clasts of granodiorite and of ophiolitic origin within the Snooks Arm graywackes.

Upadhyay (1973) noted pillow lava and diabase sills in both the Bobby Cove and Balsam Bud Cove Formations. The pillow lavas generally vary from a few metres to tens of metres, though at Scrape Point, Upadhyay (1973) observed 300 m of pillow lava. Diabase sills are very common in these units and locally compose up to 50% of the local sequence. The sills generally range from a few metres up to, in rare cases, 200 m.

Jenner (1977) and Jenner and Fryer (1980) noted a major felsic component in the upper portion of the Bobby Cove and the Balsam Bud Cove Formations. They reported lithic fragments of felsic volcanic rocks and subangular grains of quartz and plagioclase as the common constituents of the sedimentary rocks.

Venams Bight and Round Harbour Formations

Upadhyay (1973) described these units as follows:

The Venams Bight Basalt, consisting of about 500 meters of pillow lava, conformably overlies the Bobby Cove Formation. It stretches from Indian Burying Place in the west to Venams Bight in the east, a distance of about 8 kilometers. The Round Harbour Basalt comprises 1000 meters of pillow lava and forms the uppermost part of the Snooks Arm Group. It is separated from the Venams Bight Basalt by the Balsam Bud Cove Formation. It has a much shallower dip (as low as 25°-30°) than the underlying formations of the Snooks Arm Group.

The Venams Bight and Round Harbour Basalts consist primarily of pillow lava and sills with minor pillow breccia; sediments are practically absent. Chert fills the spaces among the pillows which in some cases show the dip of the associated pillow lavas.

Upadhyay (1973) also noted that the Venams Bight lavas are medium gray and slightly darker than the pillow lavas of the Betts Cove Complex; the Round Harbour basalts are dark gray to nearly black.

Upadhyay (1973) also reported that the Round Harbour lavas are less altered in thin section than the Venams Bight pillows; both consist of plagioclase, clinopyroxene, chlorite, epidote, carbonate, and opaques ± amphibole ± quartz. Jenner and Fryer (1980) noted the character of these minerals as follows:

Phenocrysts of plagioclase are commonly altered to sericite-epidote, carbonate-quartz, or quartz-chlorite. Groundmass plagioclase is generally altered to epidote. Probe analyses of fresh portions of phenocrysts give An_{76}-An_{82}. Clinopyroxene is less altered than plagioclase. Its alteration products are actinolite (especially along crystal margins) and chlorite. Prehnite and pumpellyite occur (Upadhyay, 1973; Jenner, 1977).

Upadhyay (1973) noted that the whole of the Snooks Arm Group is locally intruded by late plugs, sills and dikes of silicic composition; he reported the chief constituents of these dikes as quartz, plagioclase, and potash feldspar, with minor chlorite and micas.

Depositional Environment

The Snooks Arm Group directly overlies the Betts Cove Ophiolite Complex. Upadhyay (1973) noted:

The Bobby Cove Formation represents a chert - andesitic tuff - flysch assemblage which is invariably associated with ophiolitic rocks throughout the world. These are characteristic of a marine deepwater environment.

The abundance of pillow lava in this group indicates active volcanism during deposition of the deep-water sediments.

It is uncommon for such a thick sequence of volcanic and volcaniclastic rocks to directly and conformably overlie oceanic crust. Based on this observation and limited geochemistry, Upadhyay (1973) and Neale (1976) suggested that the Snooks Arm Group is of island arc origin. More recent and detailed geochemical studies by Jenner (1977) and Jenner and Fryer (1980) indicate that the Snooks Arm Group is geochemically more like rocks formed at an ocean island, a seamount, or an aseismic ridge.

Age and Correlation

Snelgrove (1931) first reported the following fossils from the group, which were recently confirmed by D. Skevington (reported in Dean, 1978):

> On the southwest shore of Snooks Arm, a third of a mile south of the settlement of Snooks Arm, graptolites occur in the black graphitic slates of the Snooks Arm series. These fossils have been determined by Dr. R. Ruedemann as *Loganograptus logani* and *Didymograptus gracilis.* He pronounces this faunule as of the 2nd and 3rd Deep Kill zone of America, or the Middle Arenig of Great Britain.
>
> [from Snelgrove (1931)]

This Early Ordovician age assignment is significant in that it also brackets the upper age limit of the Betts Cove ophiolite and the lower age limit of the Cape St. John Group and consanguineous Cape Brulé porphyry.

Traditionally, the Snooks Arm Group was correlated with the former Baie Verte Group (present Advocate and Point Roussė Complexes, and Flat Water Pond and Pacquet Harbour Groups). The group shows many similarities to the cover rocks of the Point Rousse Complex, particularly in composition and relationship with the underlying ophiolite, and is comparable lithologically and geochemically with portions of the Pacquet Harbour Group; thus, these correlations are maintained in this report. Correlation with the other units is not supported here.

Church (1977, 1978b) inferred that the graywackes of the Snooks Arm Group may be equivalent to conglomerates at the base of the Flat Water Pond Group. Similar conglomerates are also present in the Advocate Complex. This correlation is not readily apparent lithologically or stratigraphically (Upadhyay, 1979; personal observation). Likewise, correlation of the Snooks Arm Group with the Advocate Complex cover rocks appears at this time to be untenable due to the differences of rock types and stratigraphy between these sequences.

Upadhyay (1973) equated the Snooks Arm Group with the Western Arm Group (Marten, 1971), which occurs immediately east of the Baie Verte Peninsula and conformably overlies ophiolitic lava of the Lushs Bight Group.

FLAT WATER POND GROUP

The Flat Water Pond Group is defined as the thin, homoclinal, eastward younging sequence of dominantly mafic volcaniclastic and volcanic rocks, with subordinate conglomerate and felsic volcaniclastic and volcanic rocks, that outcrops between Pittmans Pond and Baie Verte. The unit reaches an outcrop width of 3 km. It is best exposed at the north end of Flat Water Pond, which is here taken as a representative section through the group. Locally, the section is cut by layer-parallel faults, though these are considered to be of minor significance. Conformable contacts exist, at least locally, between all subdivisions of the group (Kidd, 1974, 1977). The group is distinguished from other mafic cover sequences of the Baie Verte Belt mainly by the occurrence of boulder conglomerates at its base and at higher stratigraphic levels.

The Flat Water Pond Group displays one strong penetrative regional fabric that, in the northwestern portion of the group, is accompanied by a strong lineation fabric. Locally, a later crenulation cleavage is developed. The group has been metamorphosed to the lower greenschist facies.

Rocks of the group have traditionally been considered as part of the Baie Verte Group (Watson, 1947; Baird, 1951; Neale, 1958a, 1959b; Neale and Kennedy, 1967; Church, 1969; Kidd, 1974, 1977). In recent years, the other parts of the Baie Verte Group have been reassigned to the Advocate and Point Rousse Complexes and the Pacquet Harbour Group, leaving only this vestige as representing the group. Kidd (1974, 1977) demonstrated the unique stratigraphy of this sequence compared with other portions formerly included in the Baie Verte Group and, consequently, Williams et al. (1977) informally assigned these rocks to the Flat Water Group. The rationale of Williams et al. (1977) is adopted in this report, but in compliance with the Code of Stratigraphic Nomenclature, the sequence is herein termed the Flat Water Pond Group.

Kidd (1974) elucidated the stratigraphy of the group between Flat Water Pond and Micmac Lake and subdivided the group into seven formations. The geographic place names he assigned to the units have not been formally approved with the Canadian Committee on Geographic Names. Consequently, most of these units are not used herein. For the purposes of this report, four mappable rock associations are recognized, largely on the basis of Kidd's (1974) work, including (i) the Kidney Pond conglomerate and associated rocks, (ii) mainly mafic volcaniclastic rocks, (iii) mixed mafic and felsic volcaniclastic rocks, and (iv) mainly pillow lavas. In addition, mafic dikes and sills occur throughout the unit. Many of the following descriptions rely on the detailed work of Kidd (1974, 1977). I mapped the region north of Flat Water Pond and in the Black Brook area, and locally field checked the area mapped by Kidd.

Kidney Pond Conglomerate and Associated Rocks

The Kidney Pond conglomerate (Kidd, 1974) and associated rocks constitute a distinctive suite of mainly coarse conglomerates, that in most places forms the apparent base of the Flat Water Pond Group. Detailed mapping by Kidd (1974) indicated that the Kidney Pond conglomerate is a thin, extensive unit that locally overlies two separate, less extensive, coarser conglomerate units; Kidd (1974) considered these divisions as separate formations based largely on the differences in clast composition, though it may be more practical to consider all three conglomerates as individual members of the Kidney Pond conglomerate. The Kidney Pond conglomerate of Kidd (1974) is a thin (< 60 m) polymictic conglomerate with a dominantly homogeneous black slaty matrix; locally, the matrix consists of greenish sandstone and siltstone. The matrix-supported clasts range in size from dominantly subrounded pebbles to boulders with exposed dimensions up to 50 by 20 m (Kidd 1974); they have been described by Kidd (1974) as follows:

> The clast assemblage in the conglomerates is dominated by ophiolite complex-derived gabbroic lithologies, with subordinate diabase, mafic volcanic rocks, and argillite. A minor proportion of the clasts are granodiorite and foliated silicic volcanic rock, but these are found throughout the mapped extent of the formation. In an outcrop of green sandy matrix conglomerate opposite central Slink Pond [north of Micmac Lake], proportions of gabbroic to diabase and mafic volcanic clasts were estimated as about 2:1, and these comprise about 95% of the total clasts. The remaining 5% are mostly granodiorite clasts.

East of Trap Pond, in an outcrop along the Baie Verte highway, Williams et al. (1977) noted blocks of virginite (quartz-magnesite-fuchsite schist) within the conglomerate. In addition, in the area immediately west of northern Micmac Lake, Kidd (1974; also personal observation) noted a single block, approximately 1 by 2 m, of dark gray, crenulated, fine grained muscovite-quartz-albite schist. Schistosity in the block is oblique to the internal layering, which is tightly folded. Kidd (1974) recognized this as a predeformed block in the conglomerate and considered its source terrain to be the Fleur de Lys Belt. The top of the Kidney Pond conglomerate is characterized, in many places, by a thin cap of clast-free black slate (Kidd, 1974).

A unique occurrence of the conglomerate forms the northernmost isolated outcrop of this unit and has been described by Kidd (1974) as follows:

> ...7.8 km north of the Burlington Road Junction, outcrops of deformed boulder conglomerate are found on either side of the Baie Verte Road. It consists of pebbles and boulders originally from about 1 to 50 cm across, mostly of green quartzite and quartz-pebble conglomerate, with minor brown siltstone, and rare reddish jaspery chert, ultramafic, and vein quartz clasts. The matrix is dark green chlorite, probably derived from an ultramafic matrix, and it coats all the clasts.

In places, the highly lineated clasts strongly resemble boudinaged bedding. This isolated member is included in the Kidney Pond conglomerate mainly because (i) of its apparent near basal setting against the Baie Verte Road Fault; (ii) black slate, similar to the matrix of the conglomerate elsewhere, is associated with this conglomerate in nearby outcrops; and (iii) clastic quartz is not common in the Flat Water Pond Group, and the Kidney Pond conglomerate is one of the few divisions with significant clastic quartz content. The quartz-pebble conglomerate clasts in this outcrop are very similar to outcrops of quartz-pebble conglomerate of the White Bay Group exposed along the Trans Canada Highway immediately east of Route 420 (to Hampden), though no predepositional fabrics were found in the clasts.

In the area south of Kidney Pond, the Kidney Pond con glomerate (*sensu* Kidd, 1974) conformably overlies a conglomerate composed of enormous gabbroic boulders set in a scant argillaceous matrix (Kidd, 1974). This division has a minimum thickness of 200 m (Kidd, 1974). The gabbro boulders range up to at least 100 m across and their composition has been noted by Kidd (1974) as follows:

> They range from clinopyroxenite to anorthosite and include gas-brecciated gabbro [Plate 5-14], gabbro with high temperature foliation and minor trondhjemite. Most is somewhat leucocratic granular gabbro cut by parallel diabase dikes.

The matrix to this conglomerate has been observed at only two localities; Kidd (1974) described the best locality, in the farmlands north of Micmac Lake, as follows:

> ...[It] shows about 1.3 metres of normally striking, vertically dipping green and minor gray banded argillite with fine-grained sandy mafic volcaniclastic beds up to 10 cm thick, between gabbro which has vertical contacts conformable with the sediments. The top 10 cm of the sediment consists of pebble-size clasts of green argillite with a few pebbles of gabbro and diabase in a green argillite matrix. A crack in the overlying gabbro subparallel with bedding is seen filled with the pebbly argillite but soft sediment deformation of the other sediment beds is not apparent. A few metres to the east in this outcrop some similar sediment is seen, but its relationship to the gabbro is not well-exposed.

Plate 5-14: *Brecciated gabbro that forms part of a boulder in the boulder conglomerate south of Kidney Pond; Flat Water Pond Group.*

Gabbro and diabase at the outlet of Flat Water Pond may represent part of this unit in the north, though intense deformation of this area precludes any definite interpretation.

In the area of Flat Water Pond, the Kidney Pond conglomerate overlies an assortment of rocks that locally bear pebbles and boulders. Kidd (1974) considered these rocks as being either partly or wholly equivalent to the gabbro boulder conglomerate and he briefly described them as follows:

> ...northwest of the Burlington Road Junction, most of the rocks are sandy and silty mafic volcaniclastics, and subordinate green and gray, and minor black and maroon, banded argillite, cherty argillite, chert and slate. Beds of mafic volcaniclastics range up to at least several metres thick, and the banded argillaceous beds form units up to at least 10 metres thick... black, grey and green slate with fairly abundant pebble-size slaty argillite clasts are at least 10 metres thick. A boulder of clinopyroxenite at least 5 metres across occurs in silty volcaniclastic rock...

In addition, Kidd (1974) noted marble blocks and interlayers within this member; a block at least 10 m thick of gray brecciated marble outcrops in sandy volcaniclastic rocks northwest of Flat Water Pond, whereas elsewhere, such as south of the outlet to Flat Water Pond, the marble is a layered (centimetre scale) calcarenite, apparently conformable within black and gray slates.

Mafic Volcaniclastic Rocks

Unseparated mafic volcaniclastic rocks with subordinate conglomerate and minor pillow lava constitute most of the Flat Water Pond Group. Locally, they conformably overlie the basal conglomeratic rocks. These rocks comprise a wide band (up to 2 km wide) north of Flat Water Pond; south of the pond, their outcrop pattern bifurcates. The westerly unit conformably overlies the conglomerates and pinches out at Kidney Pond, whereas the easterly unit, sandwiched between two pillow lava units and locally overlying felsic and mafic volcaniclastic rocks, extends southward almost to Micmac Lake, where it is truncated by faulting (Figure 1-1). Kidd (1974) informally assigned these individual belts to separate formations based mainly on their stratigraphic position with respect to the intervening pillow lava unit. The unseparated rocks also form the entire group southwest of Micmac Lake. The extensive areas of these unseparated rocks at the northern and southern ends of the group probably reflect both the scale of mapping undertaken and the available exposure in these regions. Kidd (1974) mapped the central portion of the group in detail whereas only reconnaissance traverses have been conducted at the extremities. The central portion is better exposed than the extremities, as the northern area is densely forested and the southern area has a thicker covering of drift. Detailed mapping may facilitate subdivision of the rocks in these areas in the future.

The unseparated rocks are mainly pale to dark green and buff, silty and sandy volcaniclastics that commonly contain mafic clasts. Layering within the volcaniclastics ranges from a few centimetres to a metre. Conglomerate layers are common throughout the section and range in thickness from 20 cm to 2 m. Clast assemblages of these conglomerates appear to change through the stratigraphic succession; lower conglomerates are composed mostly of mafic volcanic and volcaniclastic rock whereas ophiolitic diabase and gabbro clasts up to 30 cm in diameter occur near the top of the succession. North of Flat Water Pond, outcrops of gabbro boulder conglomerate (with clasts up to 50 cm diameter) can be continuously traced through the inland area, as far north as the La Scie highway (Plate 5-15). In the same area, low in the section near the Baie Verte highway, blocks of ophiolite-like leucogabbro and slabs, up to 2 m long, of diabase occur in the volcaniclastics. These may represent the basal conglomerates of the group, which have been well documented by detailed mapping (Kidd, 1974) south of Flat Water Pond.

A distinctive clinopyroxene-bearing fragmental rock occurs within the unseparated unit south of Flat Water Pond; Kidd (1974) mapped and described two members of this rock type as follows:

> The rock in these members is gritty to conglomeratic and the clasts range from single crystals of clinopyroxene (?augite), up to 1 cm across, to devitrified glassy mafic lava clasts up to 20 cm across containing abundant identical clinopyroxene phenocrysts. The rock also contains some very flattened recrystallised pumiceous clasts without phenocrysts and very rare clasts of green cherty argillite and fine-grained doleritic rock. Albitised plagioclase crystals occur as clasts and as phenocrysts in the devitrified glassy mafic clasts. These also contain fragments of coarse-grained often cumulate-looking intergrowths of clinopyroxene grains, and of clinopyroxene and plagioclase grains (gabbro). Many of the clinopyroxene phenocrysts are marginally zoned, and some have multiple oscillatory zoned borders....Most of the clinopyroxene crystals are either euhedral or are broken pieces of euhedral crystals.... The rock as a whole is poorly sorted, with all sizes of clast from sand size upward in any rock. In a few places, the pebbles and cobbles are abundant enough to appear to form a self-supporting framework, but most often, the pebbly and larger clasts are matrix supported. Beds range from a few centimetres to about 5 metres thick, and some are very roughly graded.

In addition to the volcaniclastic and conglomeratic rocks, some other distinctive sedimentary rocks occur in this unit, including an outcrop of iron formation, beds of jasper, some

Plate 5-15: *Rounded cobbles of gabbro (outlined in black marking pen on outcrop) in gray-green silty-sandy matrix; from the inland area south of the La Scie highway; Flat Water Pond Group.*

quartz sandstone and graywacke beds, and conglomerate containing jasper and quartz sandstone clasts. The iron formation is poorly exposed in a small quarry on the western side of the Baie Verte highway, southwest of Flat Water Pond; only a small area of rubbly, thinly layered (< 1 cm) reddish brown chert and magmetite is exposed. Considering the common occurrence of boulder beds in the group, this limited rock type may represent a depositional boulder within the unseparated volcaniclastics; alternatively, it may represent a very localized depositional environment at the time of formation of the group. The other distinctive beds occur in the area of Black Brook, and are best exposed in the brook. Here again, the succession is dominated by green volcaniclastic rocks, though the monotony of these rocks is broken by interspersed sections of jasper and quartz sandstone. The purplish to red jasper is thinly layered (up to 2 cm) and forms a section approximately 30 m thick about 1 km east of the Black Brook bridge. Along the brook, thick beds (up to 1 m) of quartz-rich sandstone occur both above and below the jasper section; they also occur inland south of the brook, as far south as Pittmans Pond. These beds are characterized by coarse sand grains of quartz, blue quartz, and pink weathering feldspar in a scanty, but strongly foliated, chloritic matrix. Approximately 0.75 km west of the jasper section, a greenish gray conglomeratic bed contains pebbles of both reddish jasper and quartz sandstone in addition to cream-green tabular cherty pebbles. This member is somewhat similar in aspect and stratigraphic position to the isolated quartzite- and jasper-bearing conglomerate at the base of the group north of Flat Water Pond; the two may be equivalent. Kidd (1974) also reported quartz-bearing sandstone and graywacke from only a few other locales in the unit; thus, the rarity of clastic quartz in the group supports this correlation.

Minor members of pillow lava, rare pillow breccia, and agglomerate (Plate 5-16) are locally interleaved with the unseparated volcaniclastics. Thin pillow lava members occur on the northern shore of Flat Water Pond, on the Baie Verte highway north of the pond, and on the highway to the southwest of Micmac Lake. These may be equivalent to the main pillow lava units in the group.

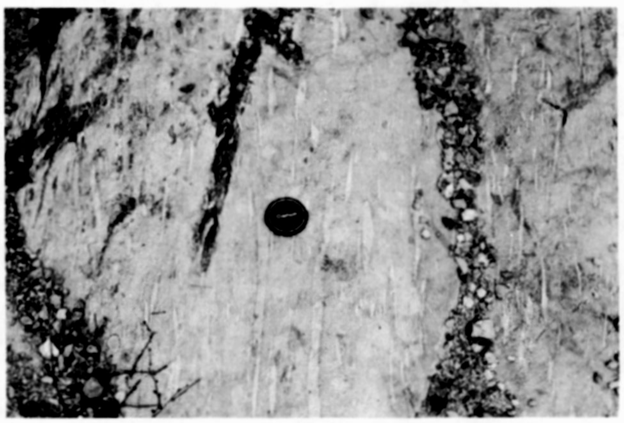


Plate 5-16: *Highly flattened mafic fragmental rock in the area directly south of Black Brook; Flat Water Pond Group (photograph by W. Muggridge).*

Mixed Mafic and Felsic Volcanic and Volcaniclastic Rocks

These interlayered mafic and felsic volcanogenic rocks overlie the unseparated rocks north of Flat Water Pond. This sequence also forms a small tongue between pillow lavas and the easterly volcaniclastic belt immediately southwest of Flat Water Pond (Figure 1-1). The felsic rocks are mainly cream to pink weathering volcaniclastics that are massive to cleaved, and range in thickness from 1 to 80 m; they appear to be dacitic. Minor massive dacite, up to 1 m thick, occurs in the northern portion of the unit. Overall, the felsic rocks constitute approximately 5 to 10% of the section. Based on thin section study and the absence of mafic-felsic mixing, Kidd (1974) suggested that these units were deposited as tuffs.

The mixed mafic-felsic sequence also contains boulder conglomerate and mafic-mineral-bearing volcaniclastics similar to those elsewhere in the group. The conglomerate is composed mainly of mafic clasts of ophiolitic derivation, though one large olistolith of serpentinized ultramafic occurs at the northeastern corner of Flat Water Pond (Kidd, 1974). Gabbroic blocks in the conglomerate range up to 50 m across (Kidd, 1974). Kidd (1974) described the mafic-mineral-bearing volcaniclastics as follows:

> Several beds 5-20 cm thick in an outcrop 0.4 km west of the northeast corner of Flatwater Pond contain abundant mafic crystal clasts, as well as albitised plagioclase and small shreds of recrystallised flattened mafic pumiceous clasts. The mafic mineral, dark green in outcrop, is a bright orange amphibole (?Ti-rich).

Kidd (1974) suggested that these beds may be lateral equivalents of the clinopyroxene-bearing volcaniclastics in the unseparated rocks south of Flat Water Pond.

Pillow Lava

The major pillow lava units in the Flat Water Pond Group form two distinct belts separated by volcaniclastics between Flat Water Pond and Micmac Lake (Figure 1-1). They appear to lens out to the north and south of this area. The thicker western pillow lava belt overlies the western volcaniclastic unit and underlies the tongue of mixed mafic-felsic volcaniclastics, whereas the eastern pillow unit overlies the eastern volcaniclastic belt and locally forms the top of the group (Figure 1-1). The western unit reaches a maximum thickness of 1600 m and was described by Kidd (1974) as follows:

> It is mostly composed of pillow lava, with minor very vesicular massive lava, and rare pillow agglomerate. A minor proportion (approx. 5-10%) consists of sandy and silty mafic volcaniclastic beds with rare beds and units of banded green, and very rare gray-black, cherty argillite, forming sediment units mostly less than 10 metres thick between some flows. Green white-weathering chert is only found as interstitial fillings between pillows but very rare banded maroon chert is found in beds....
>
> Most of the pillow lava in the formation is a very pale green, with pale yellow epidotic rims on the pillows typically about a centimetre thick. Pillow breccia/agglomerate is very uncommon. Pillows are typically 30 cm to 1 metre long, but in some flows are mostly 1-2 metres long... The pale green pillow lava is sometimes mildly amygdaloidal, with calcite and/or clinozoisite fillings. It very often has abundant white-weathering green chert in the interstices between the pillows. Massive lava is relatively rare, and all that was seen is pale green and highly amygdaloidal. It forms thin units 1 to 2 metres thick, that are seen in some cases to be at the base of mostly pillowed flows, but seem to be sharply defined and separate from the pillowed part...
>
> Dark green pillow lava forms a minority of flows.... It has a distinctive bluish tinge on the weathered surface, and usually has orange-colored epidotic rims to the pillows, that contain either or possibly both of altered vermicular chlorite and stilpnomelane. The dark green pillow lava is sometimes amygdaloidal, occasionally with large (approx. 1 cm) vesicules filled with pink clinozoisite. A few flows of dark green pillow lava have a relatively minor amount of small (< 5 mm) lathshaped (altered) plagioclase phenocrysts. Pillow agglomerate (dark green) occurs in places, but is not common. This pillow lava is clearly associated with beds of maroon banded chert up to 20 cm thick, that are not found among the pale green pillow lava. Some maroon chert and, in places, green, white-weathering chert are occasionally interstitial to the dark green pillows.

Microscopic textures in the pale green pillow lavas have been preserved in the southern portion of the unit (Neale, 1962; Kidd, 1974). Kidd (1974) described both pillow types in thin section as follows:

> [The light green pillow texture]...consists of a very fine grained randomly-oriented felted aggregate of albitised plagioclase needles, with very pale green interstitial actinolite at least partly replacing clinopyroxene. Generally, pale green pillow lava only occasionally contains a minor amount of altered plagioclase phenocrysts....
>
> ...In thin section, the dark green lava that is least deformed, east of Slink Pond, also shows a pseudomorphed original microscopic texture.

Like the pale green pillow lava, it also has a randomly oriented aggregate of plagioclase laths, but these are much larger and less elongate than in the pale green lava. The interstitial spaces are filled with a green, as opposed to a very pale green, actinolite accompanied by some chlorite, which is not seen in the pale green lava. Some (partially altered) ilmenite grains also occur, not seen in the pale green lava.

The easterly pillow lava member is poorly exposed. It was described by Kidd (1974) as follows:

> [The member is composed of] mainly medium-green rather cleaved pillow lava with minor interbeds of sandy mafic volcaniclastic rocks. To the north massive lava or fine-grained dolerite occurs... and sandy mafic volcaniclastic beds, minor green banded argillite, and one thin bed of red banded cherty argillite are found. The maximum preserved thickness... is about 430 metres.

In addition to the mappable units, Kidd (1974) described mafic sills from the group as follows:

> Mafic sills are common throughout most of the Baie Verte Group in the map area and most are medium grained dolerite. Sills range from about 10 cm to at least 120 metres in thickness. They are estimated to form 25% of the stratigraphic thickness in the section east from Kidney Pond; elsewhere they form no more than 15% of any section.

Kidd (1974) noted a rough correlation between the abundance of sills and the amount of pillow lava in any section; significantly, there are fewer sills above the main pillow unit than below it. Thus, he suggested that the sills and pillow lavas are comagmatic (Kidd, 1974). In thin section, the sills are coarser, but similar to the pillow lavas.

Contact Relationships

The Flat Water Pond Group is in tectonic contact to the west with the Birchy Complex and the Advocate Complex along the subvertical Baie Verte Road Fault. Based on relationships within the Advocate Complex, the contact between the group and the complex was probably originally an unconformity (see Age and Correlation, below). The contact between the Flat Water Pond Group and the Advocate and Point Rousse Complexes in the Baie Verte area is somewhat ambiguous, though presumably tectonic. Mafic rocks north of the La Scie highway, where these three units converge, are greenschists that lack features diagnostic of any of the units. They are herein included in the Flat Water Pond Group mainly because of their apparent structural continuity with most of the group to the south and also because of the minor occurrence of felsic volcaniclastic rocks, similar to those of the Flat Water Pond Group, in the eastern portion of this area. These rocks are in tectonic contact with Advocate Complex gabbros to the west and the Point Rousse Complex to the northeast, at South Brook. In this same general area, the Flat Water Pond Group is in tectonic contact with the Burlington Granodiorite. Southward, along the same fault, the group tectonically overlaps the Micmac Lake Group. South of Flat Water Pond, these two groups are separated by the Micmac - Flat Water Fault, though locally in this area the units are in stratigraphic contact. Neale and Kennedy (1967) first interpreted this contact as conformable, with the Micmac Lake Group being the older unit; later detailed studies by Kidd (1974, 1977) showed conclusively that the Flat Water Pond Group is overlain by the Micmac Lake Group along an erosional, and slightly angular, unconformity (Figure 1-1).

Depositional Environment

The Flat Water Pond Group represents a submarine sequence, though its substrate is uncertain. It is very likely that the group was deposited upon the ophiolitic components of the Advocate Complex (see next section). The most striking feature of the group is the preponderance of coarse conglomerates, indicating a depositional environment proximal to their source. The very coarse gabbro boulder and block conglomerates at the base of the group indicate probable deposition at the base of a submarine scarp (Kidd, 1974). Kidd (1974) interpreted most of the volcaniclastics in the group to have been deposited by grain flow (i.e. sand avalanching) mechanisms, which further supports the concept of deposition near a steep submarine slope.

The variety of clasts present indicates that a diverse assortment of emergent terrains supplied sedimentary detritus. Kidd (1974) called upon the eastern Baie Verte Belt as the source for most of the nonophiolitic clasts in the Flat Water Pond Group. Indeed, the foliated felsic clasts in the Kidney Pond conglomerate resemble Pacquet Harbour felsic volcanics and the granodiorite clasts appear identical to the nearby Burlington Granodiorite (Church, 1969; Kidd, 1974, 1977). However, it appears that other clasts had a western source. In particular, the blocks of quartz pebble conglomerate in the northern portion of the group and the muscovite schist block near Micmac Lake are very similar to rock types in the Fleur de Lys Supergroup. In addition, the closest source for the ophiolitic detritus is the Advocate Complex, which lies immediately to the west. Based on clast assemblages, it appears that the Flat Water Pond Group was deposited in a submarine basin trapped between emergent lands to both the east and the west. The gargantuan size of the ophiolitic debris in the group suggests a steep fault scarp, at least locally, on the western side of the group. Virginite boulders within the base of the group may reflect scraps of ultramafic rocks deformed and altered along this fault scarp prior to deposition.

In contrast to the dynamic sedimentation indicated by the coarse conglomerates, the jasper, siliceous argillite, and black slate present locally in the group all indicate relatively stable depositional environments. These may represent either time or space gaps between conglomerate deposition and volcanism.

Age and Correlation

Direct evidence for the age of the Flat Water Pond Group is lacking. Samples of black slate and marble from the group have been sampled and checked for acritarchs and conodonts, respectively, without success. However, there is an abundance of indirect evidence that places limitations upon the age of the group.

Traditionally, the Flat Water Pond Group was considered part of the former Baie Verte Group, and thus correlative with the Arenigian Snooks Arm Group. Neale and Kennedy (1967) dispensed with this tradition; they considered it to overlie the Micmac Lake Group and, hence, to be Siluro-Devonian in age. Kidd (1974, 1977) later showed the group to be older than the Micmac Lake Group, and reverted to an Early Ordovician age for it. Based on the clast assemblage in the Flat Water Pond Group and on regional geological relationships, Williams et al. (1977) noted that the group is probably younger than cover sequences of the Point Rousse and Advocate Complexes.

It is now apparent that there are major stratigraphic differences between the Flat Water Pond Group and the Point

Rousse Complex and Snooks Arm Group; in fact, the Flat Water Pond Group conglomerates are unique. Thus, the Flat Water Pond Group is lithically different from the cover sequences of other ophiolites and is not necessarily Arenig in age. On the basis of the clast assemblage in the group, it is younger than the formation of the ophiolite basement, and postdates the formation of virginite in the ultramafic portions of this basement. It appears also to postdate polydeformation in the Fleur de Lys terrane, as the Fleur de Lys-like clasts in the group are predeformed and recrystallized, though to a lesser extent than the Fleur de Lys rocks to the west. Based on regional relationships, this tectonism could be as young as 460 Ma (see Chapter VII). Additional circumstantial evidence for a post-460 Ma age for the group derives from the granodiorite clasts in the conglomerates. If these clasts are pieces of the Burlington Granodiorite, the group must be younger than 460 Ma, the approximate age of the pluton (see Burlington Granodiorite).

Other circumstantial evidence of the age of the group is based on regional correlation. The Flat Water Pond Group stratigraphy is very similar, and most likely equivalent, to that of the Advocate cover sequence (see Advocate Complex). The Advocate cover appears to unconformably overlie the Advocate ophiolite; the regional scale positioning of coarse ophiolite-detritus conglomerates at the base of the Flat Water Pond Group and adjacent to remnants of the Advocate ophiolite indicates that the Flat Water Pond Group, similar to its counterpart in the Advocate Complex, was probably deposited unconformably on top of the ophiolite. This suggests that a sequence conformably overlying the ophiolite, such as the Snooks Arm Group, would have to be removed before deposition of the Flat Water Pond Group. Alternatively, the possibility remains that the Advocate ophiolite was uplifted and stripped away either before or during continuous deposition of the Snooks Arm Group to the east.

Further circumstantial evidence of the age of the Flat Water Pond Group succession comes from correlation with the Llanvirnian Crabb Brook Group in the Bay of Islands area. The significance and the details of this correlation have already been discussed in considering the age of the Advocate cover sequence.

On the basis of all the evidence, the Flat Water Pond Group is probably Llanvirnian or younger in age.

MICMAC LAKE GROUP

The succession of dominantly silicic volcanic rocks that outcrop as a narrow strip in the center of the peninsula, between Birchy Lake and Flat Water Pond (Figure 1-1), was originally termed the Micmac sequence by Neale and Kennedy (1967). Kidd (1974) assigned these rocks to the Micmac Lake Group; this latter name is formally recognized herein.

The Micmac Lake Group is composed of silicic volcanic rocks (dominantly ignimbrites), conglomerate, subordinate sandstone, and mafic volcanic rocks. Most of the sequence lies to the east of the major north-south trending Micmac - Flat Water Fault (Figure 1-1) and forms a moderately westerly dipping, westward younging homocline; dips steepen to the west. On the western side of the fault, minor though significant slivers of the group are eastward younging (Kidd, 1974). The maximum outcrop width of the group is approximately 6 km, in the area south of Black Brook.

Kidd (1974) mapped and described in detail the portion of the group between Flat Water Pond and Micmac Lake. I have only briefly field-checked the portion of the group on the east side of Flat Water Pond and carried out broadly spaced traverses in the poorly exposed terrane south of Micmac Lake. Thus, the following description of the group is taken totally from Kidd's (1974) work and refers to the area north of Micmac Lake. All of the following data are either paraphrased or directly quoted from his work (Kidd, 1974), unless otherwise stated.

In the following general description of the group, Kidd used the Camp 166 road as a geographic reference; it is a gravel road that branches eastward from the Baie Verte highway, just north of Kidney Pond, and extends to Skippens Ridge, the prominent hill approximately 6 km east of Kidney Pond.

> The main homoclinal sequence of the Mic Mac Lake Group is divided into two sequences by a major internal erosion surface... The lower sequence is almost wholly restricted to the area south of the Camp 166 Road, where it overlies...the Burlington Granodiorite. It consists almost wholly of silicic volcanic rocks with minor conglomerate. The major internal erosion surface at the top of this sequence is overlain by the upper sequence, whose lower part consists of mafic lava flows and conglomerate. Further up, interbedded porphyritic silicic volcanic rocks occur in places, and then there is a rapid upward change to wholly porphyritic silicic volcanic rocks. The major erosion surface between the two sequences is not traceable northward of where the mafic lava and conglomerate of the upper sequence disappear 1 km south of the Camp 166 Road.... If it is not everywhere faulted out, the erosion surface must lie within the section consisting wholly of silicic volcanic rocks north of the point where it is lost....
>
> [Kidd, 1974]

The slivers of the Micmac Lake Group that lie to the west of the Micmac - Flat Water Fault (Figure 1-1) all represent the lower sequence. Based on my reconnaissance work, rock types characteristic of both sequences occur in the poorly exposed area south of Micmac Lake. Thus, with detailed work, it might be possible to distinguish the two sequences here.

The Micmac Lake Group displays a single penetrative fabric that is locally affected by a crenulation cleavage. The group is metamorphosed to the lower greenschist facies.

Lower Sequence

This sequence is distinguished from the Upper Sequence on the basis of (i) stratigraphic position relative to the internal unconformity of the group; (ii) the presence of nonporphyritic silicic volcanics; (iii) the relative paucity of prophyritic ignimbrites (though locally significant in this sequence) in contrast to the Upper Sequence; and (iv) the paucity of mafic rocks in the main portion of the sequence.

The lower sequence is distributed in two zones; the main zone is a continuous belt that lies immediately west of the Burlington Granodiorite and the other is a thin discontinuous zone that lies to the west of the Micmac - Flat Water Fault (Figure 1-1). The two areas display somewhat different stratigraphic successions, though correlation between them is possible. The main portion of the sequence reaches a maximum thickness of approximately 750 m whereas the slivers preserved along the Micmac - Flat Water Fault range up to approximately 300 m thick.

The main zone is dominated by a single thick unit of flow banded, massive, nonporphyritic rhyolite; this unit is locally autobrecciated at its base and the unit contains single thin interbeds of conglomerate and eutaxitic ignimbrite. The stratigraphy below this unit is a complex intermingling of conglomerate, nonporphyritic rhyolite, and distinctive pink porphyritic ignimbrites. The main rhyolite mass is overlain by a unit of red quartz-feldspar porphyritic ignimbrite near the Camp 166 road, though to the south this member is missing due to the internal unconformity in the group.

The western zone of the Lower Sequence is characterized by the presence of mafic lavas and distinctive pink porphyritic ignimbrite in most of the outcrop slivers. In addition, this zone contains massive rhyolite, conglomerate, sandstone and siltstone, and minor nonporphyritic ignimbrite.

The correlation of these two belts is based primarily upon (i) the presence of distinctive pink porphyritic ignimbrites in both areas (this rock type is confined to these portions of the group); (ii) the similarity in proportions of clast types in conglomerates of both areas, including mostly granodiorite and nonporphyritic rhyolite clasts, subordinate mafic lava clasts, and minor quartz-feldspar porphyry clasts; and (iii) the presence of nonporphyritic silicic volcanics that do not occur in the Upper Sequence. The rock types in each part of the sequence are essentially the same, and representative examples are described below.

The massive nonporphyritic rhyolite unit in the western portion of the sequence was described by Kidd (1974) as follows:

> ...It is all crimson to maroon in color, and while much is flow banded with a millimetre-scale color banding, some is homogeneous. Flow folding of the flow banding is common, but is irregularly oriented.... The thickness of the whole unit is about 440 metres. At the base, 0.25 km and from 1.1 to 1.8 km south of the Camp 166 road, a rhyolite autobreccia up to 15 metres thick is developed. Flow banded and homogeneous rhyolite clasts from small pebbles to boulders a metre across are contained in a scarlet rhyolite matrix. The clast/matrix ratio is large. The clasts may be angular brittle fragments or partly to wholly plastically deformed. Where most clasts are deformed, the rock strongly resembles the flinty eutaxitic and slightly parataxitic ignimbrites. The autobreccia locally overlies the unconsolidated (now cleaved) top of a porous eutaxitic ignimbrite 0.25 km south of the Camp 166 Road....
>
> [Kidd, 1974]

Locally, the unit displays eutaxitic structure and zones and layers of incipient brecciation; these features indicate that the unit may, in part, be a rheoignimbrite.

A pink to maroon flow that forms the northernmost exposure of the Lower Sequence was presumably fed by two poorly exposed dikes that extend eastward into the Burlington Granodiorite (Figure 1-1).

The nonporphyritic ignimbrites are restricted to the Lower Sequence; most underlie the main rhyolite unit, though one member occurs in the western zone.

> ...They show a thin (approx. 1 m or less) cleaved purple non-welded and non-vapour phase crystallised base and top in a few places. A large proportion of these eutaxites is porous-weathering, compared to the porphyritic ignimbrites. The porous-weathering rocks are pale buff, or bright pink, or pinkish purple, with dark brown or red or black fiamme.... Scarlet flinty weathering parts of these ignimbrites appear to form the central part of a cooling unit 1.1 km south of the Camp 166 Road....
>
> The porous-weathering non-porphyritic eutaxitic ignimbrites show the megascopic features of eutaxitic ignimbrites better than any other ignimbrites in the map area... The fiamme are seen differentially flattened around pebbles of rhyolite or granodiorite that are commonly found within the ignimbrites...
>
> [Kidd, 1974]

In thin section, the nonporphyritic rocks are generally uninformative:

> The rhyolite flows...and the matrix of the ignimbrites consist of an almost irresolvable homogeneous quartzo-feldspathic and hæmatite groundmass. Fiamme are very slightly more coarsely crystalline than the groundmass, but always consist of a mosaic of equant grains. They always contain a larger proportion of hæmatite than the matrix.
>
> [Kidd, 1974]

The pale pink porphyritic dark ignimbrites, which occur in both zones of the sequence, are distinctive and separable from all other porphyritic ignimbrites in the group based on the following criteria:

> (1) The pale pink colour contrasts with the scarlet red to maroon or purple of the main parts of all other porphyritic ignimbrites. (2) They are slightly to moderately porous weathering.... This is only seen elsewhere in part of the non-porphyritic ignimbrites...and very rarely in a thin development of buff ignimbrite near the base of two porphyritic ignimbrites [in the Upper Sequence].... (3) They contain almost no flattened pumice clasts (fiamme), while other porphyritic ignimbrites almost always have an abundance of them. (4) Rare small pebble-size angular inclusions of red chert are found locally in both areas in these rocks, but have not been seen elsewhere. (5) Very rare fuchsite-green (chrome-bearing) highly altered serpentinite clasts are found in places in these rocks in both areas, but have not been found elsewhere. These are mostly between a millimetre and a centimetre across. (6) These are the only ignimbrites in the map area that possess a well-preserved devitrified welded glass shard texture in thin section. In all others examined this texture has been wholly destroyed. They are the only ignimbrites in the map area that locally contain altered plagioclase phenocrysts as well as quartz and K-feldspar phenocrysts.
>
> [Kidd, 1974]

The porphyritic ignimbrite overlying the main rhyolite unit in the eastern portion of the sequence is like ignimbrites of the Upper Sequence and described with them.

In addition to these silicic rocks, a minor occurrence of a quartz-feldspar porphyry sill occurs on the brook feeding Flat Water Pond at its southern end.

The mafic lavas are confined to the western portion of the sequence and are described by Kidd (1974) as follows:

> The mafic lavas...are all basically massive lava flows.... Almost all mafic lava is green, purplish-green or purplish weathering... A few outcrops of strongly cleaved mafic lava are cream weathering... The purplish cast of many flows, especially in their vesicular bases and tops, is due to hæmatite, but red strongly haematised mafic lava is very rare. Most of the mafic lava is not porphyritic, but it is often possible to see the 'groundmass' altered plagioclase laths on the weathered surface of outcrops without using a hand lens. They range up to 1 cm long, but are usually not more than 5 mm long. Rocks in which they are not visible are sometimes found to be wholly recrystallised, although they were probably fine-grained before alteration. In the few mafic lavas sectioned, clinopyroxene (augite?) interstitial to the altered (albitised or saussuritised) plagioclase laths is often partly preserved, even in strongly hæmatised rock. Small magnetite grains are usually abundant. The texture of the lavas is usually a random 'basaltic' aggregate of the plagioclase laths, but occasionally a weakly trachytic preferred orientation is seen. The few examples of porphyritic flows contain very large altered plagioclase phenocrysts up to 3 cm long.... Units of mafic lava flows range up to 180 metres thick and each probably represents one relatively rapid sequence of eruptions. Individual flows are found down to a few metres thick, but a maximum thickness is not well defined.

Neale et al. (1960) and Neale and Nash (1963) suggested that pillow lavas located on islands in Micmac Lake are part of the group; they are not typical of rocks in the Micmac sequence and, thus, may be a portion of the Flat Water Pond Group.

The sedimentary rocks of the Lower Sequence are very similar to those of the Upper Sequence and are described below, with the latter.

Upper Sequence

This sequence forms the remainder of the group between the Micmac - Flat Water Fault and the erosional surface above both the Lower Sequence and the Burlington Granodiorite. The sequence is divided by a fault that splays northward off the Micmac - Flat Water Fault at Park Pond (the small pond on the Micmac - Flat Water Fault and south-southeast of Kidney Pond, Figure 1-1). The lower portion of the sequence is composed mainly of mafic lavas and conglomerates that are overlain and locally interleaved with porphyritic ignimbrites; these lower members grade southward into a thick unit of mainly conglomerate and sandstone. In the area directly south of Flat Water Pond, the lava and conglomerates are overlain by trachyte flows. The thickness of this eastern portion of the Upper Sequence is approximately 1150 m. The western, tectonically bounded portion of this sequence is composed almost entirely of red porphyritic ignimbrite with minor nonporphyritic rhyolite sills. The thickness of this portion is approximately 470 m.

The mafic lavas are identical to those described above with the Lower Sequence; they are commonly associated with conglomerate, sandstone, and siltstone. These clastic sediments were described by Kidd (1974) as follows:

> Most of the sediments in the Mic Mac Lake Group are conglomerates. They range from pebble to boulder conglomerate with clasts usually not more than 1 metre across, but occasional larger clasts up to 8 metres across are found. The clasts can all be matched with local lithologies. That is, the clast assemblages consist of variable proportions of Burlington Granodiorite, pink aplite from the granodiorite, purplish mafic lava, rare epidosite pebbles from the mafic lava, homogeneous and flow-banded nonporphyritic maroon rhyolite, red quartz-feldspar porphyry with or without ignimbrite fiamme, and rare nonporphyritic eutaxitic ignimbrite. One clast of maroon flow-banded rhyolite agglomerate is similar to the autobreccia, but is not precisely matchable within the Micmac Lake Group, although it obviously belongs to the same silicic volcanic suite. Two or three clasts of a purple fine-grained rock with K-feldspar phenocrysts up to 1 cm across were seen. This lithology is apparently common as a marginal facies to the syenite intrusions in the ring complex to the east (E.R.W. Neale, personal communication)....
>
> Clasts always form a self-supporting framework. Bedding is relatively uncommonly seen within the thick mapped units of conglomerate, but interbedded units of coarse sandstone (granule-size clasts) and sandstone 1 to 2 metres thick are seen in places. These show that the boulder conglomerate occurs in individual beds at least 4 metres thick, and thicker beds are probably present.... The clast size and proportions of clast types present are usually fairly constant within most of any one of the mapped conglomerate units. The only consistent exception to this is that there is usually an overall fining upward in the upper 10 to 20 metres of the thicker map units, which consist mainly of pebble conglomerate, coarse sandstone and some sandstone....The matrix of the conglomerates is mostly a granule to coarse sand-size mixture of material from the granodiorite, phenocrysts from the porphyries, and lithic grains of the silicic and mafic volcanics.... The conglomerates are rather poorly sorted....

The conglomerate matrix ranges in color from pinkish red to gray and gray-green. Locally, in the thick conglomerate members, there is a general coarsening upward trend.

Sandstone and siltstone are uncommon in the group; where present, they are generally pinkish red to gray-green and are thickly bedded. Primary structures are not common, but include crossbedding, ripple marks, small channels, and mud-cracks. The composition of these sediments was noted by Kidd (1974) as follows:

> The clastic grains in the siltstone are mainly angular to sub-angular quartz, with minor feldspar and magnetite, and much hæmatite in the shaly parts. Clastic grains in the sandstones consist of angular to sub-angular quartz, with about an equal proportion of K-feldspar, albitised plagioclase, and magnetite. Minor quantities of rhyolite fragments, muscovite, biotite, sphene and epidote occur. Heavy mineral laminae are found occasionally, and in abundance in one outcrop. These contain mainly magnetite and sphene, with lesser epidote, zircon, and apatite, and rare hornblende and rutile. As in the conglomerates, all clastic material is therefore referable to the granodiorite and the volcanic rocks. Sandstone grain size is generally 1.5 mm or less, although beds with granule-size grains (2-4 mm) are common....The matrix of all sandstones examined in thin section is mostly calcite with some sericite.

Trachyte flows occur only in the lower portion of the Upper Sequence, in the area extending approximately 3.5 km south of Flat Water Pond. They are surrounded by strongly deformed rocks and are themselves altered; the fresher flows were described by Kidd (1974) as follows:

> [They] consist of a flinty, dark purplish, slightly mottled rock, discolored to pale buff near thin quartz veins. In thin section, the strongly aligned mat of feldspar laths contains abundant fine-grained magnetite, scattered phenocrysts of K-feldspar, a few larger magnetite grains, a few small apatite grains, and rare phenocrysts of sphene. Areas of granular quartz are common.

The maximum thickness of any one flow is 48 m. The probable magmatic source of these trachytes is represented, now, by quartz syenitic plutons to the east in the Middle Arm Ridge area.

Porphyritic ignimbrites outcrop mainly in the upper portion of the Upper Sequence, though some occur in the lower portion; one ignimbrite from the upper portion of the Lower Sequence is very similar to the Upper Sequence examples. Those at the top of the Upper Sequence appear to form three cooling units (Smith, 1960). The units are difficult to delineate due to poor preservation and poor exposure, but appear to be in the order of 15 to 120 m thick, with individual flows approximately 3 m thick at a minimum. The aspect of these ignimbrites was summarized by Kidd (1974) as follows:

> The bulk of these consist of the flinty-weathering scarlet to crimson central parts of the cooling units. Near the base of two cooling units, a fairly discrete layer of up to 20 metres thick is found, consisting of a non-cleaved porous-weathering pale buff eutaxite with dark brown fiamme, and they are probably flow units. The cleaved purple, originally little consolidated material, forming the tops and bases of the cooling units is far more common. It often contains diffuse-margined off-white patches where the hæmatite has been reduced during deformation. The cleaved marginal facies contains sericite forming the cleavage. Occasional isolated pseudomorphs of glass shards are seen in it, but otherwise shard textures in the porphyritic ignimbrites have been destroyed....
>
> Included pebbles (and occasionally cobbles) of rhyolite and granodiorite are common in the porphyritic ignimbrites, and the former are easily distinguished from the fiamme. In the marginal cleaved facies and the rare buff porous-weathering near-basal facies, the brown fiamme are clearly flattened pumice clasts....
>
> In thin section the porphyritic rocks contain a fairly equal proportion of mostly equant quartz and K-feldspar phenocrysts, of a uniform size (typically 0.5 to 2 mm), and uniform proportion of the rock (20-30%) throughout the whole sequence. Most phenocrysts are broken and/or

resorbed, with K-feldspar more commonly retaining some euhedral faces than quartz. The K-feldspar now appears to be a cryptoperthite or a microperthite, and shows Carlsbad and Baveno twins. The perthite is sometimes a diffuse patch type, but more commonly is a type that mimics chessboard albite. It is possible that some of the feldspar is a chessboard albite, but this has not been positively identified.

Kidd (1974) also reported numerous maroon to dark purplish rhyolite sills associated with the porphyritic ignimbrites; they either are homogeneous or contain attenuated devitrified clasts. The sills also occur locally throughout the sequence; in places, they are spherulitic.

Contact Relationships

The Micmac Lake Group unconformably overlies both the Burlington Granodiorite and the Flat Water Pond Group. Neale and Nash (1963) first recognized the nonconformity on the granodiorite and, locally, Kidd (1974) recognized a regolith developed at this contact. Kidd also indicated that this unconformity is actually composed of two erosional surfaces. His evidence relies mostly upon the stratigraphic position of a mafic lava and conglomerate unit, which he interpreted as part of the Upper Sequence as follows:

> North of a point 1.8 km north of the Camp 166 Road, a sequence consisting almost wholly of mafic lava and conglomerate [Plate 5-17] rests on the unconformity with the Burlington Granodiorite, and appears to underlie the upper silicic volcanics, although a stratigraphic contact between the two is not exposed. It is most likely that this is the same unit as that overlying the internal erosion surface to the south. Therefore, the unconformity on Burlington Granodiorite under this northern mafic lava/conglomerate sequence is the same as the erosion surface within the group to the south. The unconformity on the Burlington Granodiorite southward from the disappearance of this northern mafic lava/conglomerate sequence is therefore a different, slightly older erosion surface.

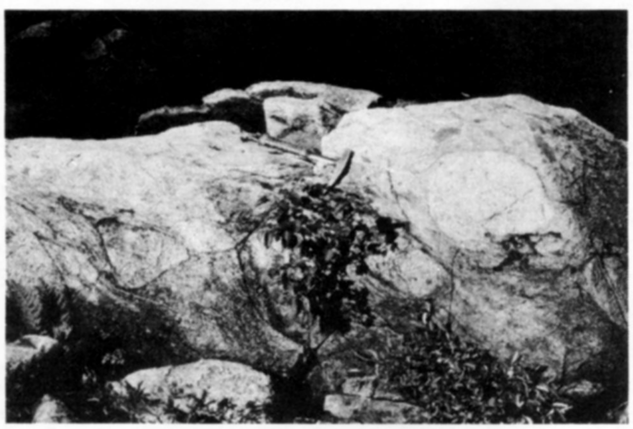


Plate 5-17: *Micmac Lake Group conglomerate containing clasts of Burlington Granodiorite nonconformably overlying the pluton on the east side of Flat Water Pond.*

Neale and Kennedy (1967) interpreted the Micmac Lake Group rocks to be conformably overlain by rocks here considered to be in the Flat Water Pond Group; Kidd (1974) has since shown that the Micmac Lake Group unconformably overlies the Flat Water Pond Group along an erosional and slightly angular discordant surface. This unconformity is preserved only in local areas west of the Micmac - Flat Water Fault (Figure 1-1). Kidd (1974) described only three places where this contact is exposed and at only two of these locales can an unconformable relationship be demonstrated. However, Kidd's (1974) work shows that, on a larger scale, this contact is a very slightly angular unconformity.

The contact between the Micmac Lake Group and the Advocate Complex ultramafic bodies south of Pittmans Pond is unexposed but is presumed to be a fault due to the highly disturbed appearance of the rocks near the contact (personal observation). Carboniferous sediments west of Indian Pond presumably unconformably overlie the group since the sediments are relatively flat lying in comparison to the moderately dipping Micmac Lake Group strata nearby.

Depositional Environment

The abundance of ignimbrites in the Micmac Lake Group strongly indicates that it was deposited subaerially. The apparently oxidized mafic flows, the primary structures preserved by the sediments, and the abundance of obvious erosional contacts support this interpretation. The immature nature of all of the sediments indicates that they were probably deposited near their source. Their characteristics most resemble alluvial fan deposits; local upward coarsening may be related to uplift of source lands and progradation whereas the normal fining upward sequences may reflect erosional exhaustion of the source. The close association of mafic lava and the conglomerates may indicate a close relationship between uplift and the formation of alluvial fans and mafic volcanism.

Age and Correlation

Two Rb/Sr whole rock isochrons on silicic rocks of the Micmac Lake Group have yielded ages of 404 $\pm$ 24 Ma (Wanless, *in* Neale and Kennedy, 1967) and 386 $\pm$ 15 Ma (Pringle, 1978). These ages are compatible with stratigraphic relationships; the group is thus most likely Late Silurian to Early Devonian in age.

Neale (1958a) originally mapped the northern portion of the Micmac Lake Group as part of the Cape St. John Group; later workers all agreed upon the resemblance and many correlated the two. The only obstacle to unanimous acceptance of this correlation was the concept of an eastern Fleur de Lys terrane (Church, 1969) in which the rocks here called the Cape St. John Group were considered as part of the Fleur de Lys Supergroup because of their deformational history and, thus, were considered older than Early Ordovician. The abandonment of this concept (DeGrace et al., 1976; this report, see Chapters VII and IX) removes any problems in correlation. Therefore, the Micmac Lake and Cape St. John Groups are herein considered broadly equivalent.

In addition, Neale and Kennedy (1967) noted the similarity of the Micmac Lake Group rocks to those of the Springdale Group and rocks of western White Bay, the Sops Arm Group (Lock, 1969a,b). The latter have yielded poorly preserved early Middle Silurian brachiopods (Neale and Nash, 1963; Lock 1969a,b) and conodonts (S. Stouge, personal communication, 1981). These units are all considered to be broadly correlative.

CAPE ST. JOHN GROUP

The Cape St. John Group is herein defined as the sequence of dominantly subaerial volcanics, the main outcrop area of which occupies the Cape St. John Peninsula north of Tilt Cove and Red Cliff Pond and east of Gooseberry Cove Head; smaller outliers of the group occur immediately north of Rogues Harbour and west of North Yak Lake, previously assigned to the informal Rogues Harbour Group (Schroeter, 1971) and the Pacquet Harbour Group (DeGrace et al., 1976), respectively. Extrusive felsic rocks and sediments mapped by Neale et al. (1960) in the Middle Arm (Green Bay) area are reportedly similar to the Cape St. John rocks and are tentatively considered as part of the group. More detailed work is needed to determine the stratigraphic status of these geographically isolated rocks. The main outcrop area of the group is disposed in a major, east trending, upright syncline. The sequence was estimated to be approximately 3500 m thick (DeGrace et al., 1976). Neither a type section nor a reference section for the group has been erected; the choice of such a section is deferred pending future detailed work.

Rocks of the Cape St. John Group were first termed the Goss Pond and Red Cliff volcanics by Snelgrove (1931). Baird (1951) later mapped the full extent of these rocks from the area west of Grand Cove to Cape St. John and assigned them to the Cape St. John Group. Neale (1957, 1958b) subsequently included mafic rocks along the northern coast of the Cape St. John Peninsula, previously considered part of the Baie Verte Group, in the Cape St. John sequence. Neale (1958b) included small slivers of felsic volcanics and conglomerate on the east side of Flat Water Pond in the group; these rocks are herein included in the Micmac Lake Group. A major diversion in stratigraphic concepts of the peninsula was introduced by Church (1969) when he assigned the northern, more deformed rocks of the Cape St. John Peninsula to the Grand Cove Group, a division of his Fleur de Lys Supergroup (Church, 1969); within this framework, the Cape St. John nomenclature was retained for rocks to the south. However, this stratigraphic nomenclature has since been abandoned by many workers (Neale et al., 1975; DeGrace et al., 1976; Williams et al., 1977; Hibbard, 1982) as well as in this report.

The main outcrop belt of the group was mapped by DeGrace et al. (1975, 1976); all of the following descriptions are taken from their work. I mapped part of the outlier immediately to the west of the Cape Brulé porphyry, as did DeGrace et al. (1976); descriptions of these rocks are from both sources. I rely entirely upon the descriptions of Maclean (1947) and Neale (1962) for the outliers of extrusive felsic rock in the Middle Arm area.

The Cape St. John Group is polydeformed and polymetamorphosed in the area north of the La Scie highway (Figure 1-1), where it attains lowest amphibolite facies metamorphism. South of the highway, tectonic effects gradually diminish such that the southern portions of the group exhibit a single regional penetrative fabric and greenschist facies metamorphism.

The best available descriptions of these rocks and their depositional environment in the main outcrop area come from DeGrace et al. (1976) and are reproduced below; the basal unconformity that they mention is described later under Contact Relationships.

Coarse calcareous sandstone units [DSjs] and minor conglomeratic beds comprise the basal part of the Cape St. John Group in the Beaver Cove area. The sandstones are light to purple-gray in color, are well-sorted and show large scale trough and wedge-shaped cross-bedding with set thicknesses ranging from 1 metre to over 2 metres. Paleocurrent directions are generally from north to south. Mudcracks were noted in some fine-grained interbeds. The sandstones consist predominantly of clasts of quartz, feldspar, red and green ultramafic rocks, rhyolite, andesite (rare) and chromite; the matrix consists of quartz and rare feldspar fragments and is calcareous, sericitic and, in places, chloritic. The conglomerate beds occur sporadically throughout the unit and contain clasts of the same lithologies (1-5 cm in size). Conglomerate overlying the basal unconformity contains clasts of volcanic rocks resembling the Snooks Arm Group (andesite, argillite and tuff), red ultramafic fragments, rhyolite, chert and minor quartz, in an andesitic to arkosic sedimentary matrix presumed to be derived from the Snooks Arm and oldest Cape St. John Groups. Generally speaking, these sandstones and conglomerates were deposited under fluviatile conditions.

Sandstone, conglomerate and breccia [also] occur sporadically throughout the Cape St. John Group. The conglomerates and breccias contain boulders up to a metre in maximum dimension which vary from subangular to rounded in shape (mostly the latter). These boulders consist of maroon, red, black and gray, porphyritic and nonporphyritic rhyolite, massive and scoriaceous andesite, quartz-K-feldspar porphyry and tuff, and minor red cherty (ultramafic?) fragments. White, well-rounded quartz cobbles and pebbles are common in some of these units. Sandstones associated with these conglomerates and breccias may be calcareous, andesitic, tuffaceous, and feldspar and quartz-rich, and commonly contain many of the fragments present in the conglomerates.

Basic volcanic flows [DSjm] occur throughout the Cape St. John Group as flows and sills generally varying in thickness from 1 to 50 m, but being much thicker in places, especially in the north part of the map-area. They are generally dark-green, fine-grained and equigranular but are commonly oxidized to a maroon-purple colour. They are commonly highly vesicular and amygdaloidal and lack any pillow structures, and are thus more massive and vesicular than volcanic rocks of the Snooks Arm Group. Associated with the flows in places are extremely fine-grained, black to gray-green andesitic tuffs. Scoria breccias, consisting of highly vesicular basic and andesitic lava fragments in a fine to medium-grained matrix of basic andesitic and dacitic tuff, occur above some of the basic and andesitic flows. Near the base of the Cape St. John Group in the Snooks Arm area, basic flows were extruded into mud, resulting in poorly-developed pillow-like structures, and breccias. North of the La Scie Highway, basic volcanic rocks of unit 15 are commonly metamorphosed to amphibolite grade, but deformed amygdules were noted in places, indicating the presence of flows in the section, and the common distribution of metamorphic hornblende crystals in well-defined layers is presumed to reflect original tuffaceous layering.

Microscopically, [these] volcanic rocks....consist of phenocrysts and microphenocrysts of plagioclase that have undergone varying degrees of alteration to saussurite, calcite and sericite; and mafic minerals which are indicated by clots of epidote. These are set in a matrix which consists of a fine-grained intergrowth of feldspar (microlitic in places), opaque minerals and alteration products (chlorite, epidote and calcite), with the texture varying from intergranular-intersertal to pilotaxitic. Secondary minerals consist of chlorite, calcite and quartz in amygdules, fractures and the groundmass.

Quartz-feldspar tuff (pseudo-porphyry - Neale, 1958[b]) [DSjt] consists of attrition-rounded, broken and bipyramidal, partially resorbed quartz crystals, euhedral to broken K-feldspar crystals, and rock fragments in a pale gray-white to orange sericitic quartz-feldspar matrix. Spherulites were noted in a few places. Green and red subangular to rounded ultramafic clasts and rhyolite fragments, ranging in size from 1 mm to 30 cm, are common, and rarer clasts of quartz-feldspar tuff are also present. Clasts of mafic volcanic rock, sediments and chert are less common. In places, [the] tuffs are overlain by thin, pink rhyolite flows that may be genetically related.... [The] tuffs appear to be intrusive and crosscutting in places. In the Long Pond area, however, the unit is reworked, and contains graded beds of arkose.... [The tuffs] can be traced into a near-vent explosive facies...in the Red Cliff Pond area. There, the fragments vary in shape from subrounded to cuspate and lunate, and are embedded in a fine-grained, carbonate-rich matrix of comminuted rock fragments and minor quartz and feldspar.

Intrusive breccia [DSjt] ...was pointed out to us in the field by George Cockburn (1974, pers. comm.). The breccias occur as dikes and irregular masses in the Nippers Harbour and the Greenwood Pond - Kitty Pond areas. They contain subangular to rounded fragments of ophiolitic rocks and granodiorite in addition to acid volcanic rocks presumed to belong to the Cape St. John Group. The fragments reach a maximum dimension of several metres, and are set in a fine-grained matrix of comminuted rock fragments, and crystals of quartz and feldspar.

The breccias are massive except for local crude stratification and graded bedding west of Kitty Pond. They clearly intrude [the Betts Cove] ophiolite. Both sharp and gradational contacts are present, however, between the breccia and the Cape Brulé Porphyry, and it seems that the breccia is at least in part older than the porphyry.

The breccia is intimately associated with acid flows and pyroclastic rocks, and in some places, intrusive breccia can be traced into pyroclastics. This suggests that the intrusive breccia provided some detritus for the tuffs [DSjt] to the east. In general, the intrusive breccia closely resembles, in both field aspect and composition, the conglomerates above the nearby basal unconformity of the Cape St. John Group. It is therefore suggested that the conglomerates are the reworked surface equivalent of the intrusive breccia.

Possibly related to these breccias are very small, numerous diatremes or intrusive breccias which cut peridotites and gabbros...in the vicinity of North and South Yak Lakes. The breccias contain subangular to well-rounded fragments of quartz-feldspar porphyry (similar to the Cape Brulé) in a mafic groundmass containing quartz, K-feldspar and plagioclase crystals. These breccias are generally massive, but the fragments are in places elongated parallel to the contact with country rocks, indicating flowage during emplacement.

Andesitic and silicic pyroclastic rocks and flows [DSja] south of the La Scie Highway are probably facies equivalent of basic volcanic rocks... exposed to the north between Cape Cagnet and La Scie. The andesites are generally dark to light-green, massive, vesicular and, in places, porphyritic. They are readily distinguished in the field from the basalts...being fine-grained, generally lighter colored, and of higher silica content. Calcite and chlorite amygdules are common. The andesites are intimately associated with dark-colored, fine-grained to aphanitic rocks of dacitic-rhyolitic composition, which are commonly flow-banded. Coarse, flow-aligned volcanic breccia is common...and, less frequently, fine grained, green to black andesitic tuff. Microscopically, the acid rocks...were noted to contain scattered moderately sericitized plagioclase microphenocrysts and K-feldspar microphenocrysts in a felsophyric matrix of alkali feldspar, opaque minerals and quartz, with minor calcite, sericite and epidote. The [andesitic] rocks grade imperceptibly into [felsic] volcanics [DSjp]. Similar gradations occur down to a flow-banding scale in [the andesite] unit... itself.

Acid volcanic rocks [DSjp] exhibit a complex volcanic stratigraphy and consist of pink to peach-colored subaerial lapilli tuff, ash-flow tuff, welded crystal (quartz and feldspar) tuff, and minor pisolitic tuff and agglomerate. Mafic rock fragments up to 1 m in maximum dimension were noted in a few places, and, less commonly, small ultramafic and acid fragments. Flow-banded spherulitic rhyolite and trachyte flows are common and are pink to maroon in color, aphanitic, and in many places contain feldspar phenocrysts. Microscopically, the flows and ignimbrites are very similar. Phenocrysts and microphenocrysts of K-feldspar, with perthitic textures in places, moderately to extensively sericitized plagioclase (rare checkerboard albite) and quartz (commonly corroded) occur separately or together in a fine grained to aphanitic matrix of moderately to extensively sericitized alkali feldspar and quartz, devitrified glass and minor pyrite, and, in addition, devitrified shards in the ignimbrites. Secondary carbonate and quartz are common in the matrix and also as fracture fillings and amygdules.

Similar rocks were described by Schroeter (1971) and DeGrace et al. (1976) in an outlier in Rogues Harbour (too small for resolution in Figure 1-1) and in the area north of Northwest Arm.

The dominantly felsic rocks immediately west of North Yak Lake (Figure 1-1) were previously considered as part of the Pacquet Harbour Group (Neale, 1958b; DeGrace et al., 1976). DeGrace et al. (1976) described these rocks as follows:

...a thick sequence (about 1800 m) of quartz-feldspar crystal tuffs predominating over gray felsic tuffs. The crystal tuffs, which are closely similar to tuffs east of Tilt Pond, contain gray-green anhedral to euhedral and broken feldspar crystals up to 7 mm in maximum dimension, and lesser amounts of anhedral quartz crystals up to 2 mm in maximum dimension. The tuffs are well-bedded, and suggestions of cross-bedding in places indicate some sedimentary reworking.

These rocks most closely resemble the Cape St. John lithic types; they are intruded by the Cape Brulé porphyry, yet, significantly, contain rounded clasts of quartz-feldspar porphyry similar to the Cape Brulé. The aspect of these rocks and their relationships, similar to those of the Cape Brulé porphyry and the Cape St. John Group, leave little doubt that these felsic rocks are equivalent to the Cape St. John Group. Where gray-green fine grained varieties of these felsic rocks abut the Pacquet Harbour Group, it is difficult to distinguish them from dacitic flows of the latter. In the field, the Cape St. John Group rocks commonly contain clasts of quartz-feldspar porphyry, though these are not ubiquitous; these are absent from the Pacquet Harbour rocks. In thin section, the Cape St. John rocks contain nearly completely sericitized feldspars and abundant sphene whereas the Pacquet Harbour Group contains dominantly clear albite and no sphene.

Cape St. John-like strata in the area of Middle Arm were described by Maclean (1947) and Neale (1962). The outlier of extrusive felsics and associated rocks north of the community of Middle Arm as well as at Middle Arm Ridge were first described by Maclean (1947) as follows:

Atop Middle Arm Ridge and near the White Hills are several patches of flat-lying or gently folded volcanic and sedimentary rocks. The main rock type is flow rhyolite. Breccia layers are always associated with the rhyolite. In places a few thin layers of bedded grit, sandstone, and conglomerate are interbedded with the flows.

These patches of rhyolite and other rocks appear to rest on the granodiorite and porphyry intrusives that form the plateau in that area. No contacts between the volcanic-sedimentary section and the intrusives were seen. The rhyolite is not known to be cut by dikes, but very thin sheets of rhyolite porphyry seem to have been injected along bedding planes in the sandstone and breccia. The rhyolite porphyry intrusives around these patches of flow rhyolite are full of blocks of similar flow rock.

Neale (1962) described the remaining outliers of dominantly felsic rocks in the Middle Arm area as follows:

Most common are flow-layered felsites which include very light gray to grayish black and moderate reddish orange to very dusky red varieties. Individual flow layers are commonly 1 to 2 mm thick and are due to lighter, quartz-rich layers alternating with feldspathic layers. In many places the flow layers are folded in a swirly, disharmonic pattern that may have resulted from deformation or flowage in a semi-consolidated state. In other places the attitude of the layering, save for local small scale flexures, is fairly constant over areas of a square mile or more. Massive, silicic aphanophyres which are intercalated with the flow-layered rocks generally contain less than 10 per cent bipyramidal quartz and microperthite phenocrysts, 1.0 mm in average maximum dimension. Spherulitic structures are common in these rocks; individual spherulites range up to 1.5 cm in diameter. The cryptocrystalline quartzo-feldspathic-devitrified groundmass of both flow-layered and massive rocks commonly contains small amounts of biotite and/or sodic amphibole and magnetite. On the basis of mineralogy these rocks are classified as rhyolites, soda rhyolites and quartz trachytes.

The pyroclastic rocks include agglomerates with felsic fragments up to 20 cm in greatest dimension, lapilli tuffs, and tuffs. Colors vary but red hues are much more common than gray. A few specimens collected as flow-layered lavas in the field were recognized as crystal tuffs in which thin, cryptocrystalline, recrystallized ash beds are disrupted by and gently arched over quartz or feldspar clasts. Also, as noted above, several specimens collected as porphyry were recognized as tuffs by microscopic examination.

Neale also noted that all of these rocks in the Middle Arm area are most likely related to the Cape St. John Group.

Contact Relationships

The Cape St. John Group unconformably overlies the Snooks Arm Group with an angular discordance (Plate 5-18) (Neale, 1957) and nonconformably overlies the Betts Cove Complex. The stratigraphic sections preserved at these unconformities were described by DeGrace et al. (1976) and are summarized in Figure 5-3. The contact between the Cape St. John Group and the Pacquet Harbour Group appears to be unexposed. Neale and Kennedy (1967) believed that the contact was either a fault or an unconformity, with the Cape St. John rocks being younger; other workers (Church, 1969; DeGrace et al., 1976) implied that both groups form a conformable sequence. I feel that this contact, though unexposed, was originally a significant unconformity based upon (i) major lithic contrasts in the groups (the Pacquet Harbour Group is dominantly mafic with subordinate dacitic rocks, whereas the Cape St. John Group is largely rhyolitic with subordinate mafics); (ii) sharp contrasts in depositional environment (the submarine Pacquet Harbour Group in contrast to the subaerial Cape St. John Group); (iii) major differences in geochemistry (see Chapter VI), which imply a significant change in magma sources for the groups; and (iv) the inferred age of the Pacquet Harbour Group being similar to that of the Betts Cove Complex and Snooks Arm Group, thus suggesting that the relationship between the Cape St. John and Pacquet Harbour Groups is likely to be an unconformity like that between the Cape St. John and Snooks Arm Groups. Along the northern coastal section, this proposed unconformity was probably obliterated by both the intrusion of the Cape Brulé porphyry between the two groups and Acadian polytectonism in this area. It is more likely that this contact may be preserved in the forested area west of North Yak Lake.

The Cape St. John Group is intruded by the Cape Brulé porphyry and the plutonic rocks near Cape St. John, informally referred to here as the La Scie igneous suite. In addition to the intrusive relationship, some workers have locally noted gradational contacts between the porphyry and the Cape St. John Group (Neale, 1957; DeGrace et al., 1976).

Age and Correlation

Five radiometric age dates that span a wide time range have been reported for the main outcrop belt of the Cape St. John Group. Rb/Sr whole rock studies have produced results of 353 $\pm$ 15 Ma and 441 $\pm$ 50 Ma (Pringle, 1978) and 385 $\pm$ 15 Ma and 520 $\pm$ 40 Ma (Bell and Blenkinsop, 1978a), whereas the single U/Pb date on a zircon is 475 $\pm$ 10 Ma (Mattinson, 1977). The location given by Mattinson for the zircon sample is in an area mapped by DeGrace et al. (1976) as Cape Brulé porphyry (Figure 1-1). It is probably in a minor zone of extrusive rock within the porphyry (see Cape Brulé porphyry). A preliminary Rb/Sr whole rock isochron of 480 $\pm$ 60 Ma by Bell and Blenkinsop and reported by DeGrace et al. (1976) is disregarded here because the researchers reported that the Cape St. John samples... "do not define a unique isochron" (Bell and Blenkinsop, 1975).

The youngest dates (353 $\pm$ 15 and 386 $\pm$ 15 Ma) are similar to numerous $^{40}Ar/^{39}Ar$ cooling ages on the eastern part of the peninsula that reflect Acadian tectonism in this area (see Chapter VII). Pringle's (1978) date of 441 $\pm$ 50 Ma appears to fit best with the age of formation of the group as inferred from circumstantial evidence. Indirect evidence strongly suggests that the Cape St. John Group is Silurian to Devonian in age, mainly because (i) it unconformably overlies Lower Ordovician rocks (Snooks Arm Group) and (ii) it resembles the Siluro-Devonian Micmac Lake Group and fossiliferous Silurian Sops Arm Group (Neale, 1957, 1958b; Neale and Kennedy, 1967; Neale et al., 1975; DeGrace et al., 1976; Williams et al., 1977). The oldest dates (475 $\pm$ 10 and 520 $\pm$ 40 Ma) are of questionable geological significance. These have been interpreted as reflecting the crystallization age of volcanics in the group (Mattinson, 1975; Bell and Blenkinsop, 1978a). I feel that these ages reflect other phenomena. The dates are comparable with the probable age of the Cape St. John unconformity on the Snooks Arm Group and, in both cases, the very youngest limits of the dates must be considered merely to keep in accord with this stratigraphic evidence. Acceptance of these dates as the age of the group would signify (i) a very rapid change in depositional environment from the submarine of the Snooks Arm Group to the subaerial of the Cape St. John Group and (ii) an even more remarkably rapid change in magmatic regimes (see Chapter VI). Another pertinent consideration is that the fossils in the Snooks Arm Group occur in its lower half; thus, the upper part of the group may be significantly younger. These rapid changes are not impossible, but I feel that they are highly unlikely in light of the circumstantial stratigraphic evidence presented above. The 475 $\pm$ 10 Ma date may reflect the age of an inherited zircon, as zircon morphology analyses were not conducted on this sample (J. Mattinson, personal communication, 1981). The oldest date (520 $\pm$ 40 Ma) may reflect Rb/Sr systematics during the petrogenesis of the distinctly bimodal Cape St. John volcanics and may indicate separate sources for the mafic and felsic members of the group (B. Fryer, personal communication, 1981). This is supported by Sr initial ratios of about 0.7030 for the 520 Ma date, whereas the Sr initial ratios on the felsic rocks range from approximately 0.7050 to 0.7100.

In summary, the Cape St. John Group is considered to be Siluro-Devonian in age based on a Rb/Sr whole rock isochron and indirect stratigraphic evidence. Outliers of felsic rock to the west and southwest are identical to the group and show similar relationships to surrounding rocks as do the Cape St. John Group rocks; this indicates that they all probably formed within the same tectonic regime and are probably similar in age.

CARBONIFEROUS(?) SEDIMENTARY ROCKS

A small patch of very poorly exposed, flat lying to gently dipping sedimentary rocks occurs at the southern edge of the peninsula, at Indian Pond; these rocks were first mapped by Dean and Strong (1975a). Their extent, as shown in Figure 1-1, is interpreted largely from topography and the very few bedrock exposures in this area. The rocks are predominantly red conglomerate and sandstone with minor siltstone and

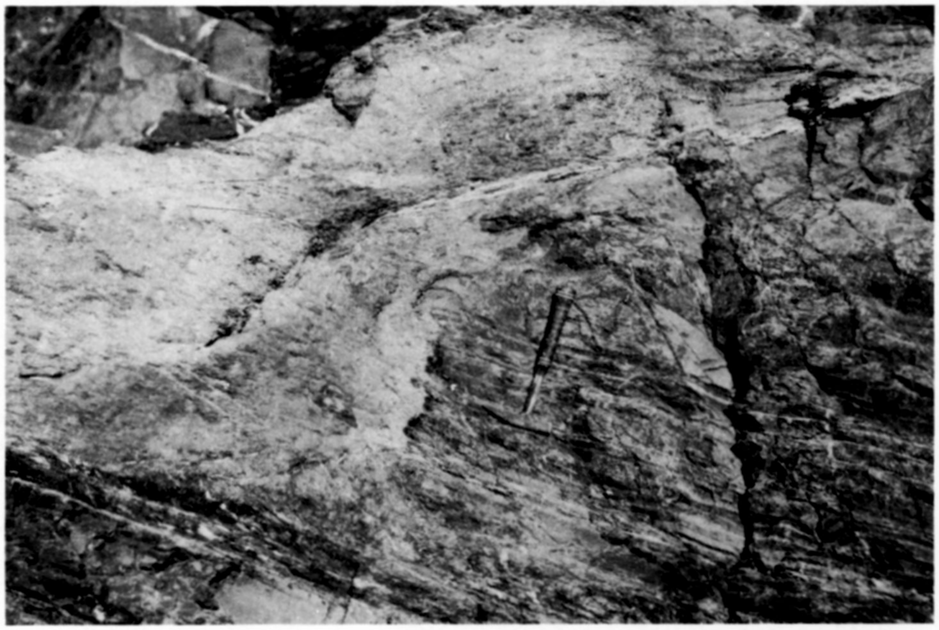


Plate 5-18: *Erosional unconformity between Cape St. John sandstone and Snooks Arm Group mafic volcaniclastics at Beaver Cove; note the truncation of the Snooks Arm beds by the overlying sandstone.*

mudstone (Dean, 1978). The conglomerate is composed mainly of angular to rounded clasts, up to cobble size, of felsic volcanic rocks and pinkish granite, though clasts of metamorphic tectonites have also been found (P. Dean, personal communication, 1981); matrix to the conglomerate is a friable red sandstone (Plate 5-19).

These sediments are probably unconformable on surrounding rocks, based on their structural attitude, but contacts are unexposed. They have been interpreted as Carboniferous in age based on their lithic similarity to Carboniferous rocks of the Deer Lake Basin, to the west, and because similar strata unconformably overlie Siluro-Devonian granitoids immediately south of the peninsula (Dean and Strong, 1975b).

INTRUSIVE ROCKS

Several major intrusions crosscut the rocks of the Baie Verte Belt, and one of these spans the boundary between the Baie Verte and Fleur de Lys Belts. These plutons are mainly granitoids, though locally they grade into diorite and gabbro; one major body of gabbro outcrops at Cape St. John.

Based on age, the Baie Verte Belt intrusions are readily separable into two suites. One suite appears to be Early Ordovician in age and is composed of the Burlington Granodiorite and the Dunamagon Granite. The latter pluton also intrudes the Fleur de Lys Belt in the Pacquet Harbour area. The other suite is Siluro-Devonian in age and comprises the Cape Brulé porphyry and an identical porphyry in the Middle Arm Ridge area, granitic to syenitic rocks near Middle Arm Ridge, a monzonitic to leucogabbroic intrusion southwest of Gull Pond, and three intrusions at Cape St. John ranging from granite to gabbro that are collectively termed the La Scie intrusive suite. The Siluro-Devonian plutons can be subdivided, based on geography, into northeasterly and southwesterly terranes, connected by a semicontinuous partial ring dike of granitic porphyry (Figure 1-1).

The northeasterly terrane comprises the Cape Brulé porphyry and the La Scie intrusive suite; these do not appear to be in mutual contact. To the southwest, in the Middle Arm Ridge area, there is porphyry identical to the Cape Brulé that locally grades into the granitic and syenitic rocks (Neale et al., 1960; Neale and Nash, 1963; Neale, 1962). The southwestern porphyry is herein described with the Cape Brulé porphyry because of their lithic similarity and the present lack of data to divide these rocks in the area of the partial ring dike. Future studies may in some way distinguish between these porphyries, so I suggest that the southerly porphyry and associated granitoids be collectively termed the Middle Arm Ridge igneous suite. The monzonitic to leucogabbroic intrusion to the southwest, which is informally termed here the Gull Pond Ridge pluton, may be a part of the suite. The Siluro-Devonian suite appears to be related to a cauldron subsidence event (Neale, 1962); this hypothesis is pursued in the summary at the end of this section. Numerous mafic and felsic dikes that intrude the suite and nearby rocks and are most likely related to this suite are described following the major intrusive rocks.

I have mapped only portions of the Burlington Granodiorite and very minor portions of the Dunamagon and Cape Brulé bodies; in addition, I have seen exposures along logging roads of a small portion of the Middle Arm Ridge rocks and the monzonite-leucogabbro near Gull Pond. Therefore, most of the following descriptions of these intrusions, particularly the thin section descriptions, are taken from earlier workers.

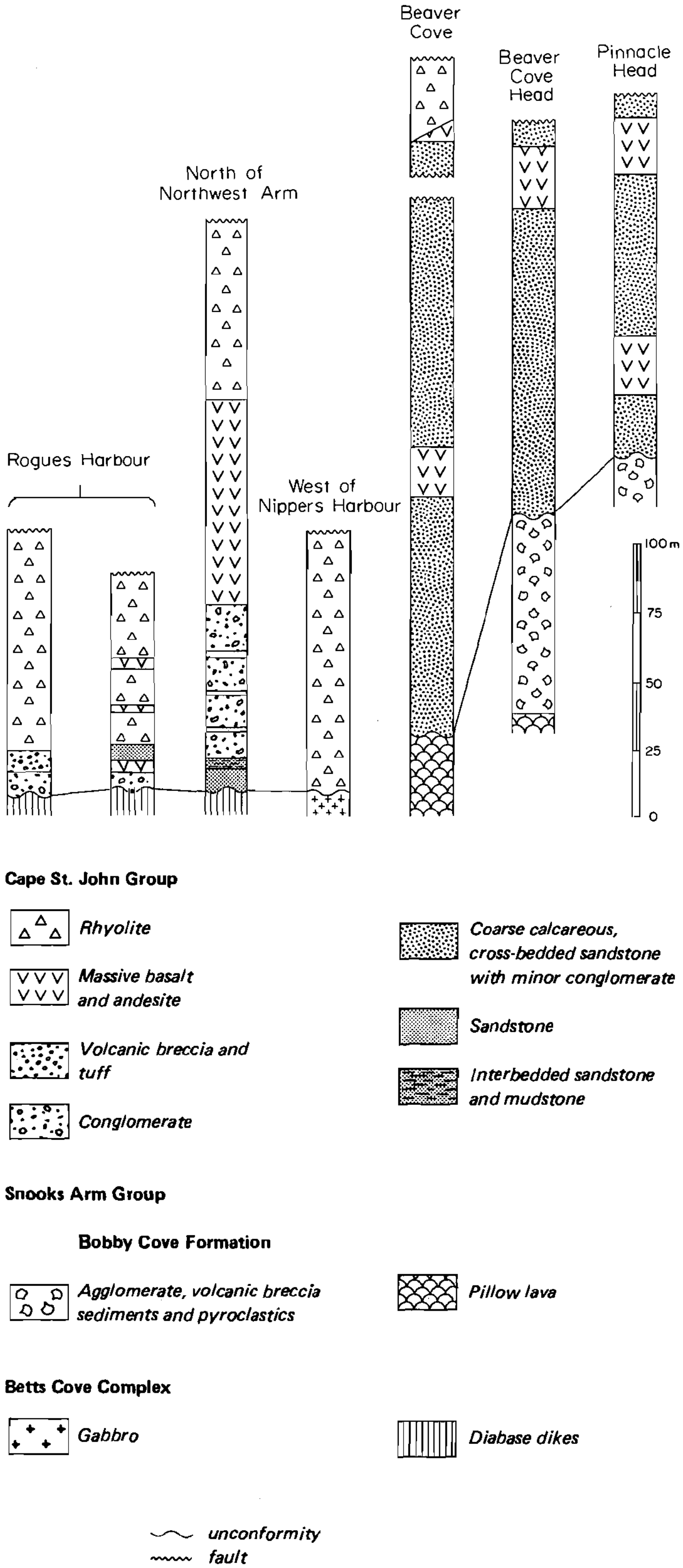


Figure 5-3: *Stratigraphic sections through the Cape St. John - Snooks Arm Group unconformity (from DeGrace et al., 1976).*

Plate 5-19: *Angular to subrounded clasts dominantly of felsic volcanics and granitoids in friable red Carboniferous sandstone, near the Baie Verte highway - Trans Canada Highway junction.*

BURLINGTON GRANODIORITE

Baird (1951) first termed the granitoid rocks in the north-central portion of the peninsula the Burlington granitic rocks. Watson (1947) had previously mapped this body as quartz diorite. Neale (1958a,b), Neale et al. (1960) and Neale and Nash (1963) outlined the full extent of this body and demonstrated that granodiorite and quartz diorite are the main phases, though granite and quartz monzonite occur locally; thus, the body has been commonly called the Burlington Granodiorite by subsequent workers.

Rocks here included in the Burlington Granodiorite extend for almost the whole length of the peninsula, from Indian Pond in the south to the hills bordering Baie Verte in the north; the maximum width of the batholith is approximately 25 km in the area of the Burlington road, but it narrows locally to less than 1 km wide, as in the area of Fox Pond. The batholith is fairly homogeneous and is characterized by a greenish gray weathering surface, though subtle differences are noticed over broad areas; for example, the granodiorite near the southernmost Gull Pond appears to be coarser grained and more quartz-rich than the granodiorite near the two Gull Ponds further to the north. I have observed three distinct phases of the batholith in the Burlington area, and detailed mapping by Epstein (1983) indicated a rather complex intrusive history for this portion of the batholith. Thus, the batholith may be a composite pluton; its overall homogeneity may reflect very similar lithic phases. If this is so, it will be difficult to document inland because of poor exposure.

The Burlington Granodiorite intrudes both the Pacquet Harbour Group (Plate 5-20) and the Betts Cove Complex; locally, along these contacts, narrow zones of agmatite have formed. On its west side, the batholith is overthrust by the Flat Water Pond Group in the north and is nonconformably overlain by the Micmac Lake Group further south. Locally, along this unconformity, east of Kidney Pond, cleaved and highly altered rock, interpreted as paleoregolith on the granodiorite, is preserved (Kidd, 1974).

Plate 5-20: *An apophysis of the Burlington Granodiorite intruding amphibolite of the Pacquet Harbour Group, on the woods road south of Consolidated Rambler Mines.*

The granodiorite appears to be foliated and more intensely altered in the area of the La Scie highway and northward. Watson (1947) noted a strong, moderately northeasterly plunging lineation in the pluton in the area of South Brook (Baie Verte). $^{40}Ar/^{39}Ar$ cooling ages indicate that the northerly portion of the batholith was subjected to Acadian resetting (see Age and Correlation).

Neale's (1962) description of this suite in the King's Point map area is representative of most of the batholith and is reproduced here as follows:

The rocks of this plutonic suite are commonly medium-grained, with the largest crystals up to 1 cm in greatest dimension. Moss-stripped surfaces are chiefly very light gray with, in some specimens, a bluish cast imparted by light bluish gray quartz. Fresh surfaces are speckled and/or mottled by dark mafic minerals, light bluish gray quartz, and moderate orange pink potash feldspar which stand out against the greenish gray... and, rarely, light gray background of altered plagioclase.

The rocks are commonly massive with a hypidiomorphic granular texture. Anhedral quartz and potash feldspar are molded around subhedra and euhedra of plagioclase. Less commonly gneissic structure has been imparted by rough alignment of mafic minerals and elongated, recrystallized quartz pods.

Thin section examination shows that the plagioclase in these rocks is very turbid due to extensive alteration of finely divided white mica and epidote. Albite twinning is not completely masked, in most specimens, by the alteration and a compositional range from An_5 to An_{16} was established. Presumably, the original plagioclase was considerably more calcic. In a few specimens clots of chlorite-epidote-carbonate-quartz have formed at the expense of plagioclase. In many specimens the plagioclase has thin, relatively unaltered albitic rims interposed between it and adjacent anhedra of potash feldspar and quartz. The potash feldspar, which is either interstitial to or mantles the plagioclase, is chiefly orthoclase microperthite. Microcline microperthite was noted in only one of 15 specimens studied. Exsolved blebs and patches within the microperthite show the mica-epidote alteration typical of the plagioclase in these rocks in contrast to the unaltered orthoclase host. Quartz anhedra show strain shadows and annealed, "sutured" boundaries in most specimens. Hornblende, pleochroic from pale brown to green, is the most abundant mafic mineral. Small amounts of brown or green biotite is commonly associated with the hornblende. Both hornblende and biotite are partly altered to chlorite in most and wholly altered in a few specimens. Accessory minerals include epidote, magnetite, sphene and, rarely, muscovite and zircon.

Modal analyses of seven typical specimens from this map-area are presented below [Table 5-1]. On the basis of the ratio of potash feldspar to total feldspar (Williams et al., 1954), they range from quartz monzonite through granodiorite to quartz diorite. Field estimates of mineral abundance suggest that granodiorite and quartz diorite are the dominant types and quartz monzonite is relatively rare....

Small xenoliths are common in the granodiorite and associated granitic rocks. Many of these are dark, greenish gray and consist chiefly of actinolite and chlorite with relatively minor amounts of albite and carbonate. Mafic mineral content is generally higher than normal in the granitic rocks which enclose them. Others are rounded, faded, "ghost-like" xenoliths that are distinguished from the host rock only by their 10 to 20 per cent higher content of mafic minerals.

In addition to the phases described above by Neale, the batholith locally includes pinkish aplite and various granitoid phases. The aplites are most common at the margins of the granodiorite in the area of Gull Pond, situated south of Rambler Mines, and along the La Scie highway. The pluton appears to be most diverse in the Burlington area; there, along

Table 5-1: *Modal analyses of granodiorite and associated intrusive rocks from King's Point and Baie Verte map areas (from Neale, 1962).*

	NA2552	BV.173	NA2125	NA2606	NA3076	BV.99	NA2535	NA2609	W.301.9**
quartz	25.8	15.13	40.5	39.4	18.1	19.6	16.9	28.0	28.8
orthoclase	25.0	9.65	16.5	16.2	14.1	9.2	3.9	2.6	3.1
plagioclase*	40.0	58.46	38.2	36.8	51.7	54.5	63.8	66.3	58.1
hornblende	3.3	6.63	--	--	8.0	5.2	3.4	tr	--
biotite	0.3	7.46	--	--	2.3	4.4	1.7	tr	11.0
chlorite	5.3	0.88	3.8	6.4	4.0	3.9	5.7	2.2	--
magnetite	0.1	0.77	0.2	0.3	1.0	--	tr	0.1	--
epidote	0.2	--	0.8	0.9	--	1.0	2.6	0.7	3.7
sphalerite	--	0.64	--	--	0.8	0.9	2.0	--	0.2
muscovite	--	--	--	--	--	0.9	--	0.1	1.0
zircon	tr	--	--	--	tr	--	--	--	tr
apatite	tr	0.38	tr	tr	tr	--	--	--	0.1
calcite	--	--	--	--	--	0.4	--	--	--
TOTAL	100.0	100.00	100.0	100.0	100.0	100.0	100.0	100.0	106.0

* Alteration products of plagioclase (i.e. epidote, muscovite, calcite, etc.) included in plagioclase count.
** Taken from Watson (1947, page 16).

a logging road north of Rix Cove, I noted a small body of muscovite leucogranite that is locally garnetiferous and intrudes typical granodiorite. Detailed mapping in the Burlington area (Epstein, 1983) indicates that there are many granitoid phases here, that range from quartz diorite to muscovite-garnet leucogranite, and which display complex intrusive relationships.

Age and Correlation

Numerous radiometric isotope studies, employing a variety of techniques on a variety of systems, have yielded a wide range of dates from the Burlington Granodiorite. However, the results of the various methods generally cluster around the following three dates: 460 Ma, 410 Ma and 345 Ma (Figure 1-1; Table 5-2). The oldest dates are interpreted as the emplacement age of the pluton, the intermediate dates may reflect slow magmatic cooling of the pluton, and the youngest dates almost certainly reflect an Acadian thermal event. A 380 Ma date from the Middle Arm area appears to be anomalous, but may fit the history of the pluton as outlined below.

The oldest dates indicate an Early Ordovician time of intrusion, particularly the U/Pb zircon date of 461 ± 15 Ma and the $^{40}Ar/^{39}Ar$ hornblende cooling date of 464 ± 5 Ma (Appendix IV), from an aplite at the margin of the pluton. These are supported by a $^{207}Pb/^{206}Pb$ zircon date of 451 ± 5 Ma (Mattinson, 1977) and fall within the uncertainty range of a Rb/Sr whole rock isochron of 494 ± 34 Ma. The approximate magmatic age of 460 Ma for the granodiorite is consistent with stratigraphic relationships. The batholith intrudes the Lower Ordovician Pacquet Harbour Group. The syngenetic massive sulfides at the Ming and Rambler Mines have yielded a Pb/Pb isotope date of 460 Ma; this date may reflect a disturbance of isotopic ratios by the intrusion of the Burlington Granodiorite.

Table 5-2: *Burlington Granodiorite age dates.*

Method		Age (Ma)	REFERENCE (if other than this report)
Rb/Sr	whole rock	494 ± 34	
$^{40}Ar/^{39}Ar$	hornblende	463 ± 5	
U/Pb	zircon	461 ± 15	
Pb/Pb	zircon	451 ± 5	Mattinson (1977)
U/Pb	sphene	434 ± 9	Mattinson (1977)
$^{40}Ar/^{39}Ar$	hornblende	418 ± 5	
$^{40}Ar/^{39}Ar$	hornblende	417 ± 8	
$^{40}Ar/^{39}Ar$	hornblende	414 ± 5	
$^{40}Ar/^{39}Ar$	biotite	414 ± 10	
$^{40}Ar/^{39}Ar$	hornblende	413 ± 5	
$^{40}Ar/^{39}Ar$	hornblende	412 ± 5	
$^{40}Ar/^{39}Ar$	biotite	412 ± 10	
$^{40}Ar/^{39}Ar$	biotite	409 ± 10	
$^{40}Ar/^{39}Ar$	hornblende	406 ± 5	
K/Ar		380	Lowdon et al. (1963)
U/Pb	apatite	345 ± 30	Mattinson (1977)
$^{40}Ar/^{39}Ar$	biotite	345 ± 5	
$^{40}Ar/^{39}Ar$	biotite	343 ± 5	

A single U/Pb sphene date of 434 ± 9 Ma has been attributed to the low blocking temperature of sphene during cooling of the pluton (Mattinson, 1977).

Nine $^{40}Ar/^{39}Ar$ dates from the central portion of the pluton, near Burlington and Flat Water Pond, all cluster around 410 Ma. The overlap of hornblende and biotite dates indicates that the cooling was rapid. The significance of these dates is somewhat ambiguous; they may reflect either final magmatic cooling of this portion of the pluton or the age of this portion of the body, thus suggesting a composite character for the whole pluton. The latter suggestion appears to be the less likely since, although minor phases of the granodiorite are recognized in the Burlington area, the batholith otherwise appears homogeneous from Burlington to Flat Water Pond.

A single K/Ar date of 380 Ma (Lowdon et al., 1963) probably reflects the thermal disturbance of the granodiorite by nearby, younger granitoids (Neale and Nash, 1963).

The youngest group of dates averages 345 Ma and consists of two $^{40}Ar/^{39}Ar$ biotite dates and a U/Pb apatite date (Mattinson, 1977) from the northern part of the pluton. These dates are identical to cooling dates from metamorphic rocks on the northern portion of the Cape St. John Peninsula and can be directly related to an Acadian thermal event (Hibbard, 1982; see Chapter VII).

In summary, the history of the pluton includes (i) intrusion at about 460 Ma as indicated by a U/Pb zircon date and an $^{40}Ar/^{39}Ar$ hornblende date from an aplite at the chilled margin of the pluton, (ii) extended cooling of the intrusion until about 410 Ma, and (iii) Acadian thermal disturbances at about 380 Ma in the Middle Arm area and 345 Ma in the northern portion of the batholith.

Neale (1962) noted that the Burlington Granodiorite is similar to plagioclase-rich granitic rocks that are associated with the early tectonic stages of "geosynclinal" regions. Dean (1978) briefly described all of the Early Ordovician plagiogranites in the northern portion of the Dunnage Zone and, of these, it appears that the Burlington is most like the Colchester, Wellman's Cove, and Dollard plutons of western Notre Dame Bay, and possibly the granodioritic phase of the South Lake Igneous Complex in north-central Newfoundland. Like the Burlington, all of these plutons appear to intrude Cambro-Ordovician ophiolitic rocks. In addition, the Burlington Granodiorite may be equivalent to lithically similar portions of the Hungry Mountain Complex (Kean, 1979), just north of Red Indian Lake, in central Newfoundland. Small granodioritic bodies within the Fleur de Lys Supergroup have also been correlated with the granodiorite (Kidd, 1974) (see Fleur de Lys Belt).

DUNAMAGON GRANITE

Baird (1951) first gave the name Dunamagon Granite to the pink biotite granite that outcrops in the highlands west of Pacquet Harbour. The main body of this pluton is approximately 10 km long and a maximum of 4 km wide; a small satellite, less than 1 km^2, outcrops along the eastern side of Ming's Bight. In addition, numerous dikes and pegmatites between Pacquet Harbour and Cape Hat are probably related to this granite (Baird, 1951).

The granite spans the boundary between the Fleur de Lys and Baie Verte Belts and intrudes both the Pacquet Harbour and Ming's Bight Groups. DeGrace et al. (1976) indicated that intrusive contacts are exposed on the southern side of Pacquet Harbour (with the Ming's Bight Group) and in a stream at the northeastern end of Belly Pond. They also indicated that xenoliths resembling rock types in each of the groups are sparsely distributed in the pluton.

The granite has been deformed and displays at least two sets of inhomogeneously developed structures (see Chapter VII).

Limited exposures of the granite were observed in the present study, so the following descriptions are taken from Baird (1951).

> The granite is not uniform in texture or composition. Feldspars may vary from 5 mm in length to as much as 15 mm; the groundmass is composed of fine-grained quartz and feldspar, with no mafic minerals, whereas in others biotite may constitute as much as 20 percent of the rock; and foliation may be marked, or almost lacking.
>
> The following essential minerals are seen to occur in thin sections: microcline and orthoclase, 40 to 60 percent; quartz, 25 to 40 percent; biotite, 2 to 20 percent; and albite, 3 to 10 percent.
>
> Quartz occurs in aggregates of irregularly interlocking grains, which show conspicuous Boehm lamellae and undulatory extinction. Trains of inclusions, some of them liquid and some of an unidentified dark mineral in dust-like particles, occur commonly in some specimens. The quartz is interstitial to the large feldspar grains.
>
> Microcline crystals as much as 3/4 inch in length and of irregular outline commonly show characteristic grid twinning on a fine scale. Albite (An_{7-20}) occurs as minor, small crystals. Alteration of the feldspars has produced a scaly aggregate in which sericite and epidote are recognizable. Perthitic intergrowths are common, and tear-drop intergrowths of myrmekite were observed in some thin sections. Fissures in the large feldspar crystals are commonly filled with secondary albite.
>
> Biotite occurs in clumps and patches, which are attenuated parallel with the foliation, and characteristically shows many pleochroic haloes. Apatite and titanite are common accessory minerals, and are almost always associated with biotite. Much of the biotite is partly scattered through the rock as irregular wisps and shreds.
>
> Occasional patches of a brownish, gel-like material were observed in thin sections, and probably represent an iron oxide. Magnetite occurs as well-crystallized octahedra, and as abundant irregular masses apparently developed during the alteration of biotite to chlorite.

In addition, Baird (1951) described the dikes associated with the granite along the coast northwest of Pacquet Harbour as follows:

> ... The dykes are as much as 20 feet wide and 3,000 feet long, but generally are much smaller. They are largely concordant with the formations they intrude, but crosscutting dykes are not rare.
>
> These pegmatites are composed largely of feldspar and quartz, with minor muscovite and biotite, and accessory garnet, smoky quartz, and tourmaline. The composition of the dykes varies considerably, and gradations exist from those of nearly pure feldspar to those that could be termed quartz veins. The feldspar is orthoclase, microcline, albite, and perthite, and graphic intergrowths of quartz and orthoclase were observed in some of the smallest dykes of this region. Although commonly distributed uniformly, biotite has been noted in crystals up to 6 inches across and may form as much as 30 percent of the rock. Greenish muscovite occurs in some dykes as crystals from 1 inch to 3 inches across. Most of the quartz is of the milky variety, but glassy and smoky varieties also occur.

Age and Correlation

Eight radiometric dates have been obtained by a variety of methods and techniques from the Dunamagon Granite. Two U/Pb zircon dates of 460 ± 12 Ma (see Appendix IV) and 435 ± 15 Ma (Mattinson, 1977) appear to indicate intrusion of the granite at about 450 Ma. This date is consistent with the intrusive relationship between the granite and the Early Ordovician, or older, Pacquet Harbour Group. A single Rb/Sr whole rock isochron of 425 ± 10 Ma has been interpreted as the original age of the pluton (Pringle, 1978), but in light of the U/Pb dates, the significance of this age is questionable. The remaining five dates include Rb/Sr biotite dates, a $^{40}Ar/^{39}Ar$ biotite date, and a K/Ar biotite date and range from 343 to 368 Ma; these reflect the affects of an Acadian thermal disturbance in this area (see Chapter VII).

Correlation of the Dunamagon Granite with other intrusive rocks on the Baie Verte Peninsula is uncertain at this time; the pluton appears to be the same age as the Burlington Granodiorite, but it is lithically more like the younger La Scie Granite further to the east (see La Scie igneous suite).

Plate 5-21: *Layering within the Cape Brulé porphyry, along the La Scie highway.*

Plate 5-22: *Close-up view of layering shown in Plate 5-21; note gradational nature of layering boundaries and mafic fragment in upper right corner.*

CAPE BRULE PORPHYRY AND SOUTHERLY EQUIVALENTS

Snelgrove (1931) first gave the name Burtons Pond porphyry to the granitic porphyry that outcrops in the inland area north of Nippers Harbour; subsequently, Baird (1951) mapped the full extent of this porphyry (Figure 1-1) and proposed the name Cape Brulé porphyry for the entire mass.

The main outcrop area of the Cape Brulé porphyry is roughly diamond-shaped and extends for approximately 25 km from Cape Brulé southward almost to Stocking Harbour and approximately 20 km eastward from Pacquet Brook to Tilt Pond. Numerous minor stocks and dikes accompany this main body and occur east of Rambler Pond, in the Nippers Harbour area, at Tilt Cove, in Brent's Cove, and immediately south of Seal Island Bight. In addition, Neale (1958a) and Neale et al. (1960) mapped an extensive tract of identical granite porphyry in the Middle Arm Ridge area; these rocks appear to be discontinuously linked to the main outcropping of the porphyry by a partial ring dike immediately north of the Burlington road. These rocks are herein designated as "southerly equivalents" to the Cape Brulé porphyry and, because they are identical to the main body (Neale, 1962), they are described below with the Cape Brulé. Future detailed studies of both areas may warrant their separation.

The porphyry has been polydeformed and polymetamorphosed in the area between Cape Brulé and the La Scie highway; southward, this tectonism gradationally diminishes, and in the area of Middle Arm Ridge the porphyry is relatively fresh compared to that of the main mass.

I have mapped only a small portion of the westerly main mass of the porphyry and the poorly exposed partial ring dike so the following descriptions are largely either paraphrased or directly taken from DeGrace et al. (1976) for the Cape Brulé body and Neale (1962) for the Middle Arm Ridge area. Based on these descriptions, as well as from my own reconnaissance field work, I have a notion that portions of these porphyries are extrusive rocks, possibly massive ash-flow tuff. In particular, the locally gradational contacts with surrounding rocks, the presence of local layering (Plates 5-21, 5-22) and hexagonal jointing, and the fragmental microscopic textures of many specimens all suggest an extrusive origin. Thus, the porphyries may comprise composite bodies with both intrusive and extrusive portions.

Cape Brulé Lithic Types

DeGrace et al. (1976) most recently described the Cape Brulé porphyry as follows:

> Mapping in the course of this project has shown that the porphyry can be resolved in the field into two units - a finer-grained porphyry with a felsic matrix [DScj] and a coarse-grained porphyry [DScm] with a matrix generally rich in mafic minerals. The coarse-grained porphyry in places intrudes and is in places in gradational contact with the fine-grained porphyry.

> The fine-grained porphyry contains feldspar phenocrysts ranging in size from 2 to 3 mm in maximum dimension and quartz phenocrysts from 2.5 to 3.5 mm. Phenocryst content in the rock is variable. Feldspars comprise from 10 to 35 percent of the rock volume (generally about 20 percent) and quartz from 10 to 25 percent (generally about 15 percent). Colour variations in the feldspars are extreme and include white, gray, green, orange, brown-orange, purple and pink. In the generally aphanitic matrix, colours are similarly variable. The feldspar and quartz phenocrysts are generally anhedral to subhedral in outline.

> In the coarse porphyry, feldspar phenocrysts range from 5 mm to 2.5 cm in maximum dimension, and quartz phenocrysts from 4 to 7 mm in maximum dimension. Feldspars may comprise locally from 15 to 55 percent of the rock volume, but generally are in the range of 30 to 35 percent. Quartz phenocrysts vary from 15 to 30 percent of rock volume,

but generally are about 20 percent. Feldspar types are more readily identified than in the fine porphyry and both orthoclase and plagioclase are widespread, occurring in shades of pink-orange and gray-green. The feldspars vary from subhedral to euhedral in outline, and quartz from anhedral to subhedral.

In thin section, orthoclase was noted to generally predominate over plagioclase. Plagioclase composition was generally about An_{15} but rarely ranged up to about An_{60}. Quartz and feldspar phenocrysts are commonly embayed, corroded and broken. In the fine-grained porphyry broken to comminuted feldspars are more abundant than in the coarse-grained porphyry, in which embayed and corroded phenocrysts predominate. The groundmass is generally very fine-grained, but in the coarse-grained porphyry a granophyric texture was noted in places between the coarser feldspar phenocrysts. In the fine-grained porphyry comminuted quartz and feldspar phenocrysts are common. Mafic constituents in the groundmass consist of biotite altered to chlorite, with minor epidote, and sphene and subhedral zircon.

Mafic mineral abundance in the groundmass of the porphyry is extremely variable. In most of the area underlain by the coarse-grained porphyry, the percentage of mafic minerals ranges up to 20 percent of the rock volume and gives the rock a very dark colour. In the fine-grained porphyry, and extending up to about 500 m inside the boundary of the coarse-grained porphyry, mafic minerals in the groundmass amount to a maximum of only about 7 percent.

The contact relations of the two porphyry types are not clear. In most places, the contact appears to be gradational over 2 or 3 metres. In a few places the gradation takes place over about 50 metres and this is perhaps a reflection of a complexly oriented contact. In places, veins of coarse-grained porphyry occur in the fine-grained porphyry; and although no chilled margins were noted, this texture, taken along with the spatial distribution of the rock types, is taken as indicating that the coarse-grained porphyry is the younger unit.

The Cape Brulé Porphyry contains numerous inclusions of fine-grained mafic, gabbroic and ultramafic rocks [resembling those of the Betts Cove Complex] [Plate 5-23], siliceous to intermediate volcanic rocks resembling those of the Cape St. John and Pacquet Harbour Groups, and, more rarely, porphyry fragments. The inclusions range in size from less than a centimetre up to hundreds of metres in maximum dimension, and the largest of these may in fact be roof pendants. In the coarse-grained porphyry, fragments of possible "Cape St. John" origin are rare, and mafic inclusions seem to be concentrated near the margins of the unit. It could be either that the presence of more abundant mafic inclusions in the coarse-grained porphyry is obscured by the dark-colored matrix, or that the mafic mineral-rich matrix of that porphyry is a product of resorption of former mafic inclusions. The latter explanation is considered to be the more probable.

Bichan (1958) suggested that the porphyry formed either a sill or a sheetlike body and, in contrast to DeGrace et al. (1976), stated that the large gabbroic-ultramafic rafts in the area of North Yak Lake represented windows through this body.

DeGrace et al. (1976) noted the widespread occurrence of well developed hexagonal jointing, particularly in the finer grained felsic facies around South Yak Lake; they noted that this was consistent with a tabular form for the intrusive. Alternatively, as noted above, the jointing, as well as characteristics described above, may indicate that the finer grained felsic porphyry is, in part, extrusive.

Lithic Types of the Southerly Equivalents

Neale (1962) described these rocks from the Middle Arm Ridge area as follows:

Most of the rocks classified as porphyry are very similar in appearance although wide variations in all aspects are known locally. Typically they are light brown rocks that consist of about 50 percent phenocrysts set in aphanitic groundmass. The phenocrysts average 2 to 3 mm long and include feldspar slightly to greatly in excess of quartz, and small amounts of mafic mineral as cumulophyric aggregates up to 1.5 mm in greatest dimension. Microscopic study shows that the feldspar is almost invariably orthoclase microperthite with up to 40 percent exsolved albite. The feldspar phenocrysts are anhedral to subhedral, the quartz phenocrysts are slightly rounded and embayed, and both are partially resorbed along their contacts with the groundmass. The mafic minerals include traces of pale green pyroxene, and "moss-like" and radiating clumps of amphibole with very strong absorption, and pleochroism in blue green. The amphibole is tentatively identified as riebeckite or arfvedsonite. The groundmass consists of anhedra of quartz, plagioclase and orthoclase that average between 0.04 mm and 0.2 mm. Small grains of magnetite and clumps of amphibole are sparsely disseminated throughout the groundmass.

Less common varieties of porphyry include pale yellowish brown, brownish gray and grayish red rocks. Phenocrysts range in abundance from 35 to 70 percent and in average size from 1 mm to 1.5 cm. In some greyish red dyke rocks that represent off-shoots from bodies of typical porphyry, quartz and feldspar phenocrysts are not developed, and small (1.5 mm) aggregates of soda amphibole form the sole phenocrysts in a cryptocrystalline, felsic groundmass.

Contact Relationships

The Cape Brulé porphyry intrudes the Betts Cove Complex, the Pacquet Harbour Group, the Burlington Granodiorite, and the Cape St. John Group. The contact of the main porphyry mass and the Pacquet Harbour Group is either unexposed or faulted but, locally, quartz-feldspar porphyry dikes identical to the main mass intrude the group east of Gull Pond and at Rambler Pond. The contact with the Cape St. John Group is locally intrusive, though Neale (1957) and DeGrace et al. (1976) reported gradational contacts between these units (see also Cape St. John breccias, described above).

Neale (1962) reported similar relationships between the southerly porphyry and surrounding units as follows:

Dykes of quartz-feldspar porphyry are known to cut the Ordovician [Burlington] granodiorite and associated rocks at several localities between Gull Pond and Rabbit Pond [Bartletts Pond] along the western margin of the main outcrop area. Numerous porphyry dykes cut these old plutonic rocks along the shores of Middle Arm and a large, arcuate-ring dyke of porphyry intrudes them immediately north of the map-area (Neale, 1958[a]).

Quartz-feldspar porphyry, containing characteristic sodic amphibole, occurs along the shores of Moose Antler Pond [immediately south of Wolverine Pond] within the [Advocate Complex] ultrabasic belt, apparently as an intrusion into the ultrabasic rocks....

The relationship of the quartz-feldspar porphyry to the syenitic and granitic plutons that disrupt its continuity around Middle Arm and Middle Arm Brook is generally gradational although obscure in places due to lack of outcrop. Thin granite sills on some of the tributaries of Rattling Brook about 3 miles west of the village also appear to be gradational into surrounding porphyry.

The relationship of the porphyry to the associated silicic volcanic rocks [see Cape St. John Group] has already been described. In many places they are conformably intermingled and, in some places, small-scale cutting relationships suggest that porphyry forms sills within some varieties of the volcanic rocks... However, in several localities near the outcrop margins of the volcanic rocks, this writer had the impression they represented large, rather gently dipping blocks 'floating' in porphyry. MacLean (1947, p. 7) apparently had a similar impression of their relationships on Middle Arm Ridge and White Hills.

Age and Correlation

Three isotopic dates, one by U/Pb methods, and the others by Rb/Sr whole rock studies, have been obtained from the Cape Brulé porphyry. The oldest, a U/Pb zircon date of 475 ± 10 Ma, has been interpreted as the original age of the pluton (Mattinson, 1977), whereas the two Rb/Sr whole rock ages

Plate 5-23: *Large xenolith of altered ultramafic rock in the Cape Brulé porphyry, along the Nippers Harbour road. R. Talkington holding hammer for scale.*

of 404 ± 25 Ma and 334 ± 14 Ma have been interpreted as disturbed ages (Bell and Blenkinsop, 1977; Pringle, 1978). Considering the relationships outlined above for the porphyries, it is apparent that they are younger than the Ordovician Burlington Granodiorite (circa 460 Ma); hence, a 475 ± 10 Ma age is dubious. This date comes from samples retrieved very close to those Cape St. John specimens that also yielded a 475 ± 10 Ma date and, likewise, may represent an inherited age (see Cape St. John Group). The porphyry appears to be a lithic and geochemical (see Chapter VI) correlative of the Cape St. John Group and the two share a gradational contact. In view of this correlation and the age constraints on the Cape St. John Group, the Rb/Sr whole rock date of 404 ± 25 Ma appears to be most consistent with stratigraphic evidence and probably reflects the age of the intrusion. The 334 ± 14 Ma date is similar to the many dates along the northern Cape St. John Peninsula that have been related to an Acadian disturbance (see Chapter VII). Hence, the porphyries appear to be Late Silurian to Early Devonian in age.

LA SCIE INTRUSIVE SUITE

The term La Scie intrusive suite is informally applied here to the three intrusions, including the Reddits Cove Gabbro (DSr), Seal Island Bight Syenite (DSs), and the La Scie Granite (DSl), that underlie an area approximately 2 by 8 km at Cape St. John. DeGrace et al. (1976) suggested that these rocks are genetically inter-related. All of them intrude the Cape St. John Group. Cockburn (1971) determined the internal intrusive sequence as follows:

> The earliest major intrusive is the Rettis [Reddits] Cove layered metagabbro complex (exposed as a coastal section almost the full length of the belt). It is cut by early diabase, and by granitic dykes which formed locally abundant hybrid rocks and agmatites. The granitic dykes are related to the two-mica (lepidomelane, muscovite) La Scie granite which also intrudes the peralkaline... granite. Quartz-albitite dykes cutting the Cape St. John metasedimentary rocks and Rettis Cove metagabbro complex are tentatively placed next in the intrusion sequence. A second generation of diabase dykes intrudes the entire sequence.

Only limited studies have been carried out on these rocks; the best available descriptions are brief and reproduced here from DeGrace et al. (1976):

> The Reddits Cove Gabbro [DSr] is a fine- to medium-grained equigranular intrusion in which igneous layering is present in places as 1-5 cm alternate pyroxene and feldspar-rich bands [Plate 5-24]. The gabbro contains up to six percent TiO_2 in thickly-disseminated ilmenite. Microscopically, the gabbro consists of plagioclase (An_{60-70}) and pyroxene altered to chlorite, the texture being subophitic to intersertal. Accessory minerals are ilmenite and apatite, and rare patches of possible relict olivine are present.

Plate 5-24: *Layering within the Reddits Cove Gabbro near Reddits Cove (photograph by J.R. DeGrace).*

The Seal Island Bight Syenite [DSs] is fine- to medium-grained and homogeneous, but contaminated with inclusions of mafic rocks in places. It consists primarily of orthoclase and microcline (commonly perthitic) and rare plagioclase, with riebeckite and opaque minerals as the mafic constituents. Quartz comprises less than 10 percent of the rock and is commonly graphically intergrown with orthoclase.

The La Scie Granite [DSl] is a heterogeneous fine- to medium-grained pink biotite granite which is K-feldspar porphyritic in places....

Cockburn (1971) also identified muscovite in the granite and aegirine in the Seal Island Bight Syenite.

Age and Correlation

Three $^{207}Pb/^{206}Pb$ zircon dates on the La Scie Granite and a U/Pb zircon date and a Rb/Sr whole rock isochron for the Seal Island Bight Syenite are available (Figure 1-1). The La Scie Granite dates of 488 $\pm$ 35 Ma and 486 $\pm$ 40 Ma are most likely the result of inherited zircons, as noted by Mattinson (1977). The 462 $\pm$ 40 Ma Pb/Pb date on the granite is of uncertain significance. The U/Pb date of 435 $\pm$ 15 Ma on the syenite is an inferred age (Mattinson, 1977) as the data lie slightly above the concordia; this date indicates a Silurian age for the pluton, which is consistent with available stratigraphic evidence. The youngest date of the syenite of 324 $\pm$ 25 Ma (Bell and Blenkinsop, 1977) falls within the range of dates on the Cape St. John Peninsula that record an Acadian disturbance.

The suite shows alkaline to peralkaline petrochemical affinities; the petrography of some similar granitoids in the Middle Arm Ridge area indicates that these rocks may be geochemically similar; tentatively, they are all considered correlative. These rocks may all be related to the Siluro-Devonian peralkaline Topsails Igneous Complex (Taylor et al., 1980) that occurs immediately to the south of the Baie Verte Peninsula.

GRANITE AND SYENITE IN THE MIDDLE ARM RIDGE AREA

Neale et al. (1960), Neale and Nash (1963), and Neale (1962) are the only workers to have reported on these rocks in the Middle Arm Ridge area. I have only briefly observed these rocks along a logging road between Middle Arm and Kidney Pond. The following descriptions of the granite and syenite in this area are reproduced from Neale (1962):

Distribution: These rocks outcrop briefly around and west of Middle Arm, Green Bay, as either several plutons or as a single large pluton which contains partial screens and inclusions of porphyry and silicic volcanic rocks. Exposures are relatively scarce and traverses were widely spaced in this region so that the outline of the pluton (or plutons) shown on the map may be grossly in error. In part, e.g. immediately south of Middle Arm, contacts were drawn on the basis of vegetation distribution. This part of the area was burned over about 40 years ago and regrowth has generally been least on rocks of [granitic porphyry], more on the granitic and syenitic rocks being discussed, and greatest on the basic volanic rocks [of the area].

Another, separate pluton is located along and west of Southwest Arm, between Rattling Brook and Southwest Head. Small offshoots of this body, not mappable at this scale, occur within older granodioritic rocks on the east shore of the Arm. A small pluton outcrops along Patty Brook; at least two occur in granodioritic rocks north of Middle Arm; and small amounts of red granitic and syenitic rocks are associated with granodiorite 'screens' in porphyry east of Gull Pond.

... These granitic and syenitic rocks cut the older granodiorite and associated rocks - intrusive contacts are best displayed along and near the north shore of Middle Arm.... The granitic and syenitic rocks are, in many places, apparently gradational into the quartz-feldspar porphyry (Cape Brulé equivalents) but are known to cut and contain inclusions of the silicic volcanic rocks (related to this unit)....

Lithology: The granitic and syenitic rocks of the smaller plutons, e.g. that along and west of Southwest Arm, are generally brighter coloured than those of the large pluton(s). Chiefly, they are pale reddish brown to moderate reddish orange whereas those in the large pluton(s) west of Middle Arm are pale brown and light brownish gray. The brighter coloured varieties commonly contain abundant megascopically visible quartz and were classified as granites whereas the darker varieties lacked easily discernible quartz and were classified as syenites in the field. Microscopic study shows that many of the darker varieties contain micrographic intergrowths of quartz and feldspar and that quartz is commonly present in amounts sufficient to warrant classification as quartz syenite or even granite. Mineralogically, they generally consist almost wholly of orthoclase or orthoclase microperthite, quartz and plagioclase - in that order of abundance. The plagioclase is common albite; in a few subporphyritic varieties it is zoned albite-oligoclase-andesine. Mafic minerals are scarce and generally total between 2 and 5 percent. They include very dark brown biotite, amphibole - common hornblende in some specimens and a deeply pleochroic sodic variety in others, chlorite, epidote and magnetite. Traces of sphene, zircon and apatite occur in most specimens. Sodic pyroxene occurs with sodic amphibole in one specimen examined. Stilpnomelane occurs with riebeckite in subporphyritic granite near the mouth of Rattling Brook. In two other specimens studied, stilpnomelane and magnetite are the only mafic minerals present.

Texturally the brighter coloured rocks of the small plutons are chiefly equigranular with an average grain size of 2 to 3 mm. However, subporphyritic, medium-grained varieties are known. Megascopically the darker varieties convey the appearance of medium-grained equigranular rocks but microscopically most of them prove to be subporphyritic or serial porphyritic with orthoclase and zoned plagioclase phenocrysts up to 1 cm and average groundmass minerals about 1.5 mm.

In the area of Middle Arm and Middle Arm Brook, I have noted that the granitoid rocks commonly contain numerous rounded "globules," up to 3 cm in diameter, of mafic material. In the area of Middle Arm Ridge, the granitic rocks locally grade into massive porphyritic patches in which amphibole(?) up to 2 mm across and green altered feldspar form phenocrysts in an aphanitic, brownish, brick red matrix.

Age and Correlation

Direct evidence for the age of these rocks is lacking. Neale (1962) tentatively interpreted them to be genetically related to the surrounding porphyry (see Cape Brulé porphyry) and felsic rocks (see Cape St. John Group) in the Middle Arm Ridge area. He based this correlation upon the gradational relationships between the granitoids and the porphyry and the sodic affinities of both rock types, as reflected in their constituent amphiboles. Stratigraphic evidence and isotopic dates indicate that the porphyry is either Late Silurian or Early Devonian in age; this age is therefore tentatively assigned to the rocks of the Middle Arm Ridge area. Isotopic data indirectly support this age assignment; the granitic and syenitic rocks are sodic in character and possibly correlative with the Seal Island Bight syenite, which has yielded a Rb/Sr age of 435 $\pm$ 15 Ma. In addition, a K/Ar date of 380 Ma from older granodioritic rocks in Middle Arm area has been interpreted as a disturbed age, marking the thermal effects of the granitic and syenitic rocks on the older pluton (Neale and Nash, 1963; Neale, 1962).

The sodic nature of these plutons suggests that they are related to the Siluro-Devonian peralkaline Topsails Igneous Complex (Taylor et al., 1980) that occurs south of the peninsula.

MONZONITE AND RELATED ROCKS AT GULL POND RIDGE

Monzonite and related rocks in the south-central portion of the peninsula were first mapped and described by Neale et al. (1960) and Neale (1962). I have seen only limited exposures of these rocks along the woods road that leads into Gull Pond; thus the geology of this pluton has been compiled (Figure 1-1) from Neale et al. (1960). Neale (1962) noted:

> Exposures [of the pluton] are relatively sparse but the general outcrop area is characterized by higher aeromagnetic intensity than that of the surrounding greyish green granodiorite... The boundaries of the pluton are drawn from this aeromagnetic data.

The contact zone between this body and the surrounding granodiorite is poorly exposed along the Gull Pond road and is cryptic. The zone is characterized by a complex aggregation of fine grained granitic rocks and local intrusive breccias containing fine grained mafic fragments; this collection of rocks grades in both directions into rocks typical of the respective pluton on either side. Neale (1962) noted red pegmatitic material cutting the nearby granodiorite, indicating that the monzonite body may intrude the granodiorite.

Neale (1962) described the rocks of this pluton as follows:

> Most of these rocks are light brown, pale brown and pale reddish brown mottled with 10 to 20 percent greenish black hornblende crystals. These rocks were classified as medium-grained syenites in the field but microscopic work shows that some of them contain up to 18 percent quartz and all of them contain plagioclase equal to or in excess of orthoclase. Hence they are now classified as quartz monzonites and monzonites. In several places these rocks grade into medium dark grey, medium-grained pyroxene hornblende diorite. In two locations, the southwestern and the south central parts of the pluton, they apparently grade through diorite into coarse-grained leucogabbro that consists of 15 to 20 percent pyroxene, hornblende and ore minerals and the remainder plagioclase.

Along the Gull Pond road, I have noted grayish pink and pink felsic dikes, locally feldspar porphyritic, crosscutting gabbroic members of the suite as well as local stocks of medium grained monzonite bearing blobs of mafic material. Locally, along the road, the gabbro contains up to 2% biotite.

Age and Correlation

The age of this body is unknown. The monzonitic portions somewhat resemble the granitoids of the Middle Arm Ridge area, and the local inclusion of mafic material is characteristic of the latter bodies. Tentatively, the pluton is considered to be related to the Middle Arm Ridge suite; it may be a link related to both the suite and the mafic dikes described next.

MAFIC DIKES

Mafic dikes are widespread throughout the Baie Verte Belt, though they compose less than 1% of the belt. Generally, all of these dikes are locally sparsely feldspar porphyritic and trend east to northeast, indicating that they may all be related. They occur in three settings, namely (i) as intrusions into the Burlington Granodiorite, (ii) as dikes in plutonic rocks of the Middle Arm Ridge area, and (iii) as tectonized dikes within all units along the northern coast of the peninsula.

Mafic dikes intrude the Burlington Granodiorite at three locales, including the Burlington area, the area of Flat Water Pond and immediately to the south, and the area of Gull Pond and Gull Pond Ridge, at the southern end of the batholith. These dikes are generally fine grained and are locally feldspar porphyritic. Bimodal dikes, noted at two localities, comprise mafic dikes that were subsequently intruded along their central axes by whitish pink felsic dikes (Plate 5-25). Two such dikes were noted on the Smith's Harbour logging road and one was identified near Gull Pond. Kidd (1974) suggested that these composite dikes may be related to Micmac Lake Group mafic volcanism.

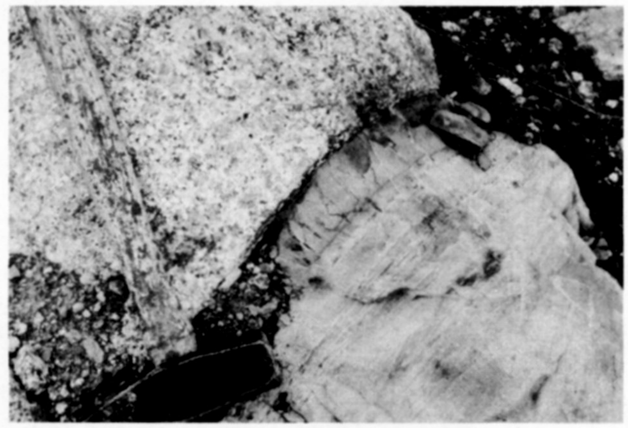


Plate 5-25: *Bimodal dike exhibiting a thin mafic margin and a thicker, flow banded felsic core; intruding the Burlington Granodiorite in the area southwest of Gull Pond.*

Numerous diabasic to gabbroic dikes crosscut the granitic rocks of the Middle Arm Ridge suite. These dikes generally trend in a 070° direction and are locally up to 3 m wide. They range from dark grayish-greenish black diabase and feldspar porphyritic diabase to medium and coarse gabbro with intersertal to subophitic textures; locally, the latter contain large feldspar phenocrysts, up to 3 cm long (Plate 5-26). In places, these dikes reveal multiple intrusive features, in which finer grained gray diabase forms the matrix to brecciated medium grained diabase to gabbro. North of Middle Arm Brook (Green Bay), these mafic dikes truncate fine grained felsic aphanites that crosscut the granitic and syenitic rocks (DSg) of the area.

DeGrace et al. (1976) briefly described a similar set of dikes along the northern coast of the Cape St. John Peninsula that were deformed and metamorphosed to amphibole-biotite schists. These dikes are generally less than 3 m wide and occur in all of the units along the coast between Cape St. John and Ming's Bight, including the Dunamagon Granite and the Ming's Bight Group. Church (personal communication, 1980) noted that the petrochemistry of one of these dikes, in the Ming's Bight Group, is characterized by high concentrations of TiO_2 and P_2O_5. This is similar to the geochemistry of mafic rocks in the Cape St. John Group at Mother Burke Rock; hence, Church (personal communication, 1980) informally termed the dikes the Mother Burke dikes. At this time, it is uncertain if all of these dikes are petrochemically similar, but they have similar field aspects.

Since all of these dikes are lithically similar and most postdate the younger volcanic and intrusive rocks of the Baie

Plate 5-26: *Plagioclase porphyritic gabbro dike in the area directly west of the Middle Arm townsite; shade variation of outcrop surface due to inhomogeneous weathering.*

Verte Belt, they are probably Siluro-Devonian or younger in age. The deformed and metamorphosed mafic dikes must be pre-Acadian (see Chapter VII). Considering these age constraints and the collective similarity of these rocks, all of the dikes are probably related to a late pulse of the same Silurian - Early Devonian magmatic activity that formed many of the plutons of the Baie Verte Belt. Individual groups of dikes, though, may not necessarily be directly related. The monzonitic to leucogabbroic body near Gull Pond is, in part, lithically similar to the coarser mafic dikes and may be directly related to them. Lithically similar mafic dikes with trends similar to those described above also occur on the southeastern corner of Eastern Island of the Horse Islands. The dikes there are undeformed and have been only partly recrystallized. If these are correlative with the dikes described above, then the tectonization affecting dikes on the peninsula may not have affected the Horse Islands rocks.

SILURO-DEVONIAN MAGMATIC ACTIVITY

The Baie Verte Belt was a locus of vigorous magmatic activity during the Siluro-Devonian periods. Neale (1962) was the first to note that there are two major Siluro-Devonian magmatic centers on the Baie Verte Peninsula, one on the Cape St. John Peninsula and the other in the Middle Arm Ridge area, apparently connected by a partial ring dike. In each area, he noted a similar sequence of granitoids, porphyry, and extrusive rocks and sediments that grossly exhibited the same relationships. On the Cape St. John Peninsula, the Cape Brulé porphyry both intrudes and is gradational with the extrusive Cape St. John Group; the contact between the porphyry and the intrusive rocks of the La Scie intrusive suite is not exposed, but the latter rocks intrude the Cape St. John Group. Similar relationships are found in the Middle Arm Ridge area, where granitic porphyry is both intrusive into and transitional with extrusive rocks in the area; it is also flanked immediately to the west by the extrusive Micmac Lake Group and southward, off the peninsula, by the lithically identical Springdale Group. Both the porphyry and the extrusive rocks in the Middle Arm Ridge area are intruded by granitic to syenitic plutons.

The similarity of the individual components of each sequence is striking. The La Scie intrusive suite and Middle Arm Ridge granitoids are similar, and both have sodic mafic minerals. The granitic porphyries in each area are virtually identical, and the extrusive rocks flanking both areas (Cape St. John and Micmac Lake Groups) are similar and both display a bimodal character. Furthermore, late mafic dikes intrude the sequences in both areas.

These data prompted Neale (1962) to write the following:

> ... the logical conclusion to be drawn from these relationships is that the porphyry plutons and arcuate ring-dike represent a center of post-Ordovician volcanism and the site of major surface cauldron subsidence.... The large blocks of lava and pyroclastic rocks preserved within the porphyry probably represent some of the earliest products of volcanism. They subsided, probably along ring fractures, to achieve their present structural level within porphyry and granite which were probably emplaced along ring fractures and cone sheets. According to this view, the older granodiorite, ultrabasic and volcanic 'inliers' might actually be screens of country rock separating the ring-intrusions that form these (postulated) composite plutons. Subsidence and intrusion were probably accompanied by widespread fracturing and accompanying volcanism and the flanking Cape St. John (which at the time of writing included Micmac Lake Group rocks) and Springdale rocks were probably formed at this time.

Neale obtained further evidence for this hypothesis from the sedimentary rocks of the Cape St. John and Springdale Groups; all contain fragments of quartz-feldspar porphyry, and the Springdale Group rocks also contain granitic fragments similar to the granites of the Middle Arm Ridge area. Thus, Neale noted:

> This suggests that intrusion, uplift and erosion of some of the porphyry, its enclosed volcanic rocks, and possibly even the associated granite and syenite took place prior to cessation of deposition of the flanking Cape St. John and Springdale Groups.

Kidd (1974) also noted the close association of tectonic activity and magmatism in the Micmac Lake Group. Here, the mafic flows are closely associated with conglomerates, indirectly suggesting a link between uplift and mafic magmatism. Furthermore, mafic dikes in the nearby Burlington Granodiorite trend in the same direction as faults that appear to have controlled the topography of erosional surfaces in the group (Kidd, 1974).

The detailed age relationships between rocks in each center are somewhat equivocal. They may all be of the same age or they may represent diachronous magmatic events wherein the two centers may only partly overlap in age. Based on stratigraphic relationships, both centers must be younger than the Burlington Granodiorite (circa 460 Ma). Isotopic dates interpreted as the age of rocks from both centers range from 441 ± 50 Ma (Cape St. John Group) to 386 ± 15 Ma (Micmac Lake Group). The evidence thus indicates that the age of the ancient caldera(s) can be imprecisely constrained to Silurian to Early Devonian times, though a Late Ordovician initiation of magmatic events is possible.

CHAPTER VI

GEOCHEMISTRY

INTRODUCTION

The main purpose of the present nonsystematic geochemical study is two-fold: to help distinguish the various basic meta-igneous rocks of the peninsula and to determine the magmatic nature of the Wild Cove Pond Igneous Suite. Characterization of the tectonic setting of the metabasic rocks by geochemical methods was a secondary consideration since the rocks are almost totally recrystallized. However, geochemical comparison of these metabasic rocks with more pristine units in the region indicates that alteration is insignificant; the comparison has yielded significant information concerning the tectonic environment of all of these rocks. There are insufficient geochemical data to attempt a detailed documentation of the petrogenesis of the rocks involved.

This study supplements many previous geochemical studies on the peninsula. This chapter is divided into the following three parts: the first section is a compendium of the results of previous geochemical studies, the second section presents geochemical data obtained during this study, and the final section integrates and correlates all of the geochemical studies and briefly discusses their significance in light of the regional geological setting.

PREVIOUS WORK

More than 800 geochemical analyses have been reported for rocks of the Baie Verte Peninsula over the past forty years. Most of the geochemical investigations form parts of local, detailed studies; hence, the distribution of analyzed samples is patchy (Figure 6-1). Almost all of the analyses are of either igneous or meta-igneous rocks, though a few are of sedimentary rocks (see Jenner, 1977; Jenner and Fryer, 1980). Results of systematic studies are outlined in Table 6-1 and units sampled by nonsystematic investigations are summarized in Table 6-2. All of the previous analyses are compiled with the present results in a computer file at the Department of Mines and Energy.

From the studies outlined in Table 6-1, it is apparent that the Fleur de Lys Belt rocks are geochemically distinct from those of the Baie Verte Belt. The Fleur de Lys rocks are most likely related to a rift environment (de Wit and Strong, 1975). In contrast, the Baie Verte Belt rocks appear to record the geochemical evolution of an ophiolite sheet overlain by distinct volcanic sequences. The ophiolites collectively show an east to west geochemical gradient from high MgO and low TiO_2 to normal oceanic basalts (Church, 1977). Based on geochemical data, the Pacquet Harbour Group appears to be of ophiolitic affinity and fits into the observed geochemical spectrum of ophiolites on the peninsula (Gale, 1971, 1973). Significantly, many workers have noted the high magnesian character of the ophiolitic rocks and have suggested that they originated at a fast-spreading ridge (Gale, 1973; Norman and Strong, 1975; Coish, 1977a); thus, perhaps the spectrum reflects accelerated action at an oceanic ridge. Basalts of the Snooks Arm Group, overlying the Betts Cove ophiolite, are tholeiitic, but the interpretation of the tectonic environment in which they formed is controversial; some workers consider these rocks to be of island arc affinity (Upadhyay, 1973; DeGrace et al., 1976; Upadhyay and Neale, 1979) whereas more recent work has indicated that they may be of oceanic island origin (Jenner, 1977; Jenner and Fryer, 1980). The younger Cape St. John Group appears to be related to late stage development of an island arc (DeGrace et al., 1976). The intrusive rocks of the Baie Verte Belt that have been analyzed by DeGrace et al. (1976) all appear to be geochemically related to an alkaline parent magma, with the exception of the Burlington Granodiorite. Their limited analyses on the granodiorite indicate that it has a more calc-alkaline affinity than the other intrusions.

Where nonsystematic study areas (Table 6-2) overlap in distribution with the systematic study areas, the results appear to support the findings of the more detailed investigations. The most significant result of the nonsystematic studies comes from analyses of the Middle Arm Ridge granite in the Rattling Brook, Southwest Arm area (Church, personal communication, 1981); the average of these five samples has normative acmite and, based on their Al_2O_3/FeO(total) ratio (MacDonald, 1975), these rocks are comenditic. The peralkaline character of the granite strongly suggests that the related porphyries (Cape Brulé and southerly equivalents), which locally contain sodic amphibole, have alkaline affinities; however, comprehensive geochemical data is lacking for these units.

RESULTS OF PRESENT STUDY

Eighty-three samples were analyzed for major elements during the study and almost all of these were also analyzed for trace elements. The sample distribution is as follows:

FLEUR DE LYS BELT

- 9 - *White Bay Group amphibolite*
- 4 - *Rattling Brook Group amphibolite*
- 7 - *amphibolite and eclogite from the Old House Cove Group and East Pond Metamorphic Suite*
- 7 - *Birchy Complex: 5 of metagabbro and 2 of mafic schist*
- 25 - *Wild Cove Pond Igneous Suite*
- 1 - *Partridge Point Granite*
- 1 - *mafic dike from the Horse Islands*
- 1 - *ultramafic in Rattling Brook Group*
- 1 - *metagabbro pod in Ming's Bight Group*

BAIE VERTE BELT

- 15 - *Pacquet Harbour Group: 10 of pillow lava and massive lava, 2 of mafic dikes, 1 of mafic crystal tuff, and 2 of felsic extrusives*
- 7 - *Flat Water Pond Group: 6 of felsic extrusives and 1 mafic dike*

Table 6-1: *Summary of previous systematic geochemical studies on the Baie Verte Peninsula.*

REFERENCE	STRATIGRAPHIC UNIT	ROCK TYPES	GEOCHEMICAL CHARACTER	INTERPRETED TECTONIC ENVIRONMENT	GEOCHEMICAL CORRELATION
Gale (1971, 1973)	Pacquet Harbour Group	Mafic volcanics	Basaltic komatiite	Ophiolite; oceanic ridge during initial rifting or rapid spreading	Similar to Barberton, South Africa and Cape Vogel, Papua
			Tholeiitic basalt	Uncertain, probably transitional oceanic ridge - island arc	
		Mafic intrusions	Tholeiite		
		Felsic volcanics	Low K_2O with uncertain relationship to mafic rocks		
Upadhyay (1973, 1978a, 1982); Upadhyay and Neale (1976)	Betts Cove Complex	Ultramafics, gabbro, diabase dikes and pillow lava	Derived from a single parental magma; Na_2O enrichment in upper portion of complex yields calc-alkaline trend on AFM diagram	Ophiolite, mid-ocean ridge	
		Selected pillow lavas and dikes	Peridotitic and pyroxenitic komatiites and minor boninites	Close associations with island arc; probably a marginal basin	Similar to rocks at Destor, Quebec and Orthis, Greece
	Snooks Arm Group	Basalt	Tholeiites with slight alkalic affinity	Island arc volcanics	
Norman (1973); Norman and Strong (1975)	Point Rousse Complex	Ultramafics, gabbro, diabase dikes, pillow lava	Derived from a parental magma of low Ti and low K abyssal tholeiite type; locally, pillow lavas comparable to basaltic komatiites	Ophiolite; mid-ocean ridge; possibly fast spreading	Papuan ophiolite, Oman ophiolite
de Wit and Strong (1975)	Mafic dikes of Old House Cove Group and East Pond Metamorphic Suite	Amphibolite and eclogitic amphibolite	Transitional tholeiitic to alkalic basalts	Continental rifting inferred from correlation	Long Range Mountain diabase dikes and mafic flows
DeGrace et al. (1976)	Cape St. John Group	Felsic and mafic volcanics	Bimodal distribution with respect to SiO_2; calc-alkaline affinities. Bimodality suggests felsics and mafics not part of a continuous differentiation series. CaO and SiO_2 appear to have been mobile in mafics; alkalis mobile in felsics	Late stage island arc	Similar to their few "Pacquet Harbour Group" samples
	Snooks Arm Group		Same as Upadhyay above	Island arc volcanics	
	Burlington Granodiorite, Cape Brulé porphyry, La Scie Granite, Seal Island Bight Syenite, Reddits Cove Gabbro		Felsic intrusions *except* Burlington Granodiorite are geochemically similar and may be related to alkaline Reddits Cove magma		
Church and Coish (1976); Church (1977)	Betts Cove Complex	Pillow lava	Low TiO_2, high MgO basalts	Unspecified oceanic ridge	End member of a geochemical series of Newfoundland ophiolites; similar to Thetford Quebec ophiolitic lavas
Coish (1977a); Coish and Church (1979)	Betts Cove Complex	Pillow lava, lower and intermediate levels	Depleted in Ti, Zr, Y, P and REE; also have high Al_2O_3/TiO_2 ratios. Derived from melting of severely depleted lherzolite	Oceanic ridge	Analogous to lower lavas in Arakapas Fault Zone, Cyprus; Khan-Taishir ophiolite, W. Mongolia; Marianas trench lavas
		Upper level	Typical mid-ocean ridge basalt	Off axis oceanic ridge	
		Dikes	Similar to lower lavas		
Coish (1977a,b)	Betts Cove Complex	Entire ophiolitic sequence	Documented geochemical changes during metamorphism. Noted increases in Fe_2O_3, MgO, Na_2O and Al_2O_3; decreases in CaO and Ca; variable changes in SiO_2, total Fe, K_2O, Ba and Rb.	Metamorphism occurred at a mid-ocean ridge during hydrothermal circulation of heated seawater; the intensity of alteration may represent a faster spreading ridge than normal	
Jenner (1977); Jenner and Fryer (1980)	Snooks Arm Group	Pillow lava	Tholeiitic basalt enriched in large ion lithophile elements	Closest to ocean island volcanics	Similar to Icelandic basalts and portions of the Galapagos volcanics
		Sedimentary rocks		Origin and evolution open to interpretation	

Table 6-2: *Summary of previous nonsystematic geochemical studies on the Baie Verte Peninsula.*

REFERENCE	STRATIGRAPHIC UNIT	NUMBER OF ANALYSES AND ROCK TYPE
Watson (1947)	Advocate Complex	3 zoisite-prehnite schist and metagabbro
	Point Rousse Complex	2 ultramafic and mafic plutonic rocks
	Burlington Granodiorite	1 quartz diorite
Neale (1962)	Advocate Complex	1 virginite
	Burlington Granodiorite	2 granodiorite 1 quartz diorite
	Cape Brule porphyry	3 granitic porphyry
Craig (1967)	Betts Cove Complex	3 pillow lavas 1 diabase dike
Bursnall (1975)	Advocate Complex	2 pillow lavas 1 massive lava 1 mafic dike
W.R. Church (unpublished data)	East Pond Metamorphic Suite	3 eclogitic amphibolites
	Birchy Complex	1 metagabbro (South Cove schist)
	Ming's Bight Group	1 metagabbro 1 metasediment
	Betts Cove Complex	4 pillow lavas 2 trondhjemite
	Snooks Arm Group	1 pillow lava 1 felsic flow
	Advocate Complex	2 pillow lavas
	Pacquet Harbour Group	2 pillow lavas 1 mafic flow 1 mafic dike
	Burlington Granodiorite	8 samples
	Flat Water Pond Group	1 pillow lava 2 granodiorite clasts 2 granitic clasts
	Cape St. John Group	5 mafic rocks 1 intermediate volcanic 2 felsic volcanics
	Micmac Lake Group	2 mafic flows
	Middle Arm Ridge Suite	5 granite
	Mafic dikes	1 intruding Ming's Bight Group 1 intruding Betts Cove Complex volcanics
Epstein (1983)	Burlington Granodiorite	28 samples
	Cape Brule porphyry	3 samples
	Dikes	5 intermediate to mafic in Burlington area

1 - Advocate Complex pillow lava
3 - Cape Brulé porphyry
1 - mafic dike, Point Rousse Complex

The methods for sample preparation and analysis are given in Appendix III, along with the final analyses; sample locations are shown on Figure 6-1. Analyses used for determining norms and in formulating geochemical plots in this chapter have been adjusted for oxidation and hydration and recalculated to 100% according to the method of Irvine and Baragar (1971).

The following discussion focuses mainly on the geochemistry of the three major divisions of basic meta-igneous rocks on the peninsula, i.e. (i) pods and layers of amphibolites in the White Bay, Rattling Brook, and Old House Cove Groups and East Pond Metamorphic Suite, collectively termed here the Westerly amphibolites, (ii) mafic rocks of the Birchy Complex, and (iii) mafic rocks of the Pacquet Harbour Group. As well, the geochemical characteristics of the Wild Cove Pond Igneous Suite and Partridge Point Granite are outlined. The remaining "spot" analyses have been undertaken for miscellaneous reasons and are presented only in Appendix III of this report.

ALTERATION

The mafic rocks and felsic volcanics that have been analyzed are almost totally recrystallized, to at least greenschist grade, and rarely exhibit primary textures. Such metamorphism probably produced some chemical alteration of the rocks. Field evidence is abundant for chemical migration, such as calcite, quartz, epidote and chlorite in cavities, veins and vesicles within the units sampled. Petrographic evidence further supports the field evidence for chemical migration: in most of the mafic rocks, plagioclase, represented by albite, is altered to sericite and epidote group minerals. Also, the plagioclase is locally chloritized. The extent of chemical migration is uncertain; the Wild Cove Pond Igneous Suite is relatively fresh compared with other rocks analyzed.

Figure 6-2 gives an indication of the extent of alkali mobility within the mafic rocks analyzed. According to Hughes (1973), the area outlined within the plot statistically represents the range of igneous compositions; thus, analyses that plot outside this envelope, or "igneous spectrum," are considered to be altered with respect to alkalies. In this plot, the Westerly amphibolites display a strong shift out of the envelope toward the soda-rich portion of the diagram, typical of spilitized lavas. The trend of the Birchy Complex samples is markedly different from that of the Westerly amphibolites and appears to indicate that these rocks have undergone alkali exchange involving dominantly potash depletion and soda addition. The Pacquet Harbour Group samples apparently lack any clear trend, though they do range outside of the "igneous spectrum;" this suggests that any alkali alteration in the group was inhomogeneous.

Previous studies by de Wit and Strong (1975) and Coish (1977a,b) addressed the nature of other alteration in mafic rocks on the peninsula. De Wit and Strong (1975) geochemically correlated the Westerly amphibolites with relatively fresh Eocambrian-Cambrian(?) basalts of the Long Range Mountains. They noted that metamorphism of the Fleur de Lys rocks involved a closed system with essentially no bulk alteration of these rocks. It has been shown above that soda enrichment is evident in the Fleur de Lys rocks; considering de Wit and Strong's (1975) findings, it is implied that the alteration predated regional metamorphism.

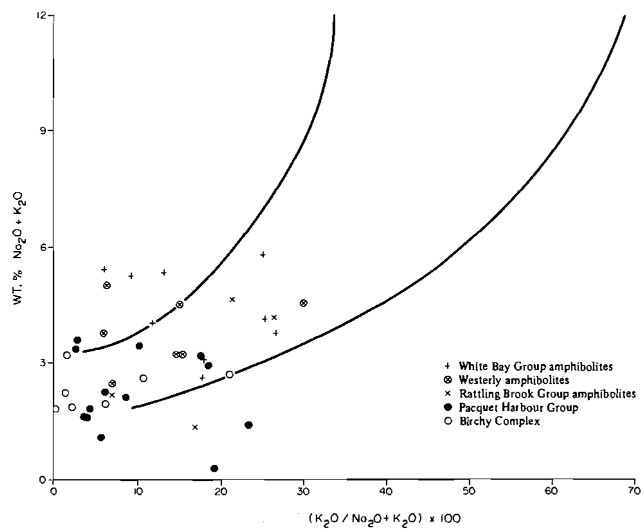


Figure 6-2: *Igneous spectrum plot (from Hughes, 1973).*

Coish (1977a,b) demonstrated that the major alteration of the Betts Cove Complex can be attributed to ocean floor metamorphism; he considered Fe_2O_3, MgO, Na_2O, H_2O, CaO, Cu, SiO_2, FeO(total), K_2O, Ba and Rb as mobile components and Ti, P, Ni, Cr, Zr, Y and FeO(total)/MgO as stable components. The two ophiolitic units considered in this geochemical investigation may be expected to display geochemical alteration patterns similar to those of the Betts Cove Complex. Contrary to Coish's (1977b) findings for Betts Cove, MgO appears to have been a stable component within the Birchy Complex and Pacquet Harbour Group. In these rocks, MgO displays a systematic variation with some of the elements considered to be immobile both by Coish (1977a,b) for the Betts Cove Complex and by Floyd and Winchester (1975) for altered basalts (e.g. Ti, P, Zr, Y and Cr; Figure 6-3). If MgO is considered as an immobile component and used as a differentiation index (Figure 6-4), then CaO values, in addition to the alkalies, appear to be consistent with data from the Betts Cove Complex (Coish, 1977a,b). There is insufficient data to substantiate other common alteration patterns within the ophiolitic units of the peninsula. The apparent lack of P_2O_5 in the Birchy Complex rocks and of Nb in most of the Pacquet Harbour mafics may reflect either alteration or very low original abundances. In addition, individual samples display local alterations that are discussed below where relevant.

On the basis of the foregoing discussion, I feel that the original geochemical traits of rocks in the area are grossly preserved, in some cases so well retained that the data can be a significant aid to stratigraphic correlation and can be used, to some extent, in the tectonic characterization of the units. Also, early alteration appears to have had more influence on element distribution in the Westerly amphibolites than have ensuing tectonism and regional metamorphism.

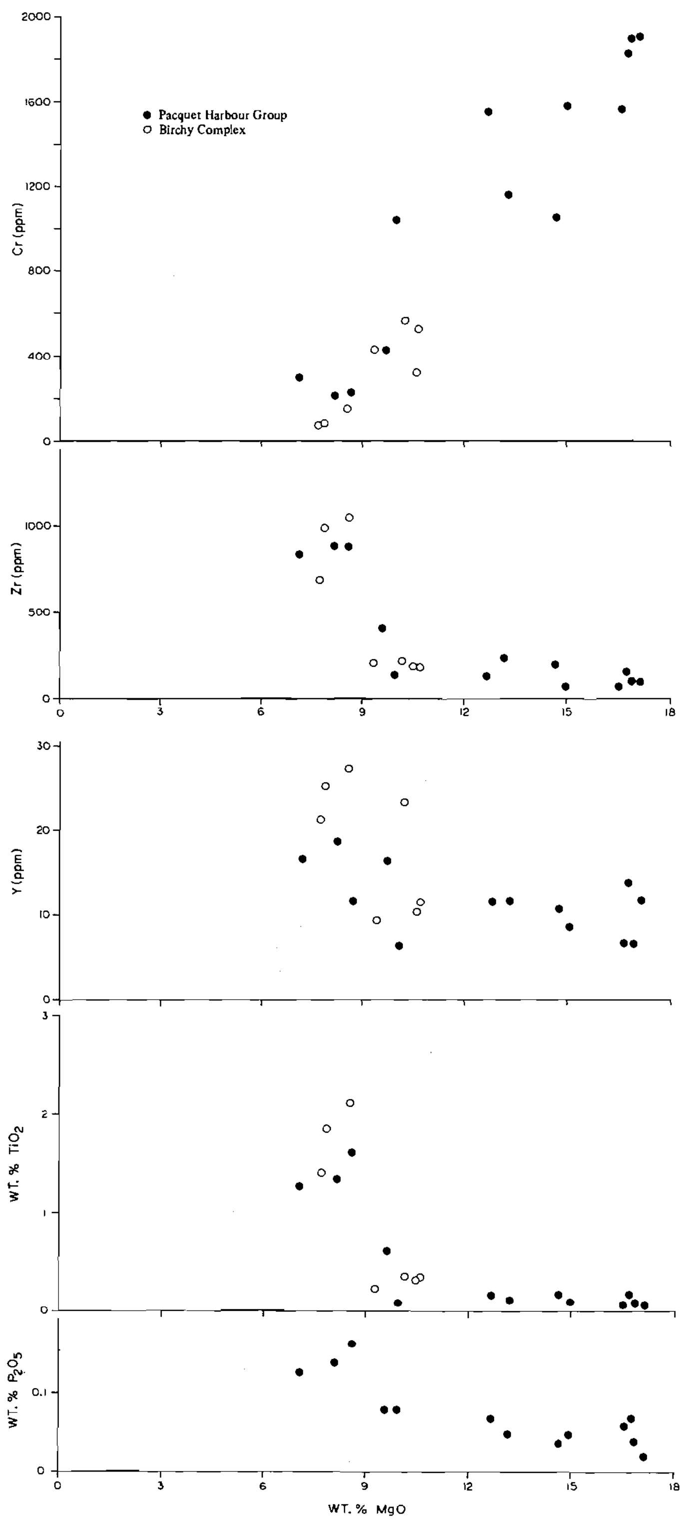


Figure 6-3: *Selected variation diagrams for the ophiolitic units.*

DATA AND RESULTS

The following presentation concentrates upon four distinct geochemical units, the Westerly amphibolites, the Birchy Complex, the Pacquet Harbour Group, and the Wild Cove Pond Igneous Suite. On an AFM diagram (Figure 6-5), it is apparent that the three mafic units show tholeiitic differentiation trends that are distinct from the calc-alkaline trend of the intrusive suite; also, though data are scarce, it appears that the Westerly amphibolites are separable from the ophiolitic rocks. This separation is best displayed in variation diagrams discussed below with each unit.

Westerly Amphibolites

During the present study, it was found that Westerly amphibolites, i.e. those from the White Bay Group, Rattling Brook Group and amphibolite pods from the Old House Cove Group and East Pond Metamorphic Suite, all share similar geochemical traits; hence, they are discussed together here. The rocks are massive, generally featureless, and probably represent either flows or dikes, though a volcaniclastic origin cannot be discounted (see Chapter IV).

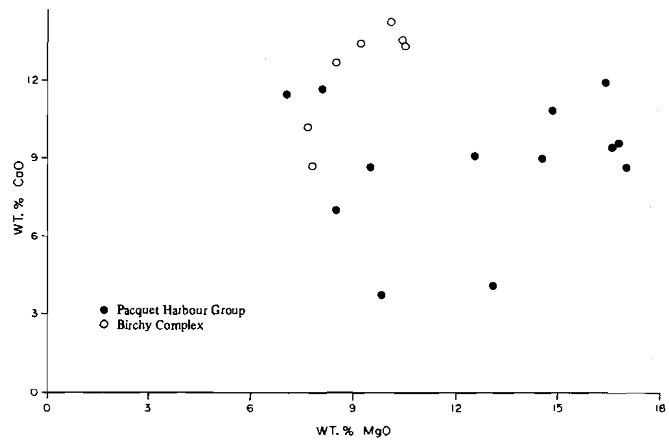


Figure 6-4: *CaO vs. MgO for the ophiolitic units.*

The average major element geochemical features of each unit within the Westerly amphibolites as well as an overall average are given in Table 6-3. Samples 1340021 and 1340023 from the White Bay Group and sample 1340030 from the Rattling Brook Group have been excluded from these averages due to either an inordinately high LOI or a low total for the analysis (Appendix III). However, they have been plotted in the accompanying variation diagrams. Sample 1340023 appears to be depleted in total iron and silica and sample 1340030 is noticeably lacking silica; sample 1340021 has no apparent depletions, though Al_2O_3 is low and may account for the low total for this analysis.

The amphibolites cluster between 45 and 52% silica and are enriched in iron oxides. On the AFM diagram (Figure 6-5) and alkalies vs. silica plot (Figure 6-6), the amphibolites display subalkaline to slightly alkaline trends. The alkaline affinity of some of these rocks is most likely due to alteration involving soda enrichment, as shown in the "igneous spectrum" diagram (Figure 6-2). The subalkaline character of these rocks is borne out by their trends on plots devised by Floyd and Winchester (1975); the plots discriminate alkaline from subalkaline rocks by using immobile elements such as Ti, P, Zr, Y and Nb (Figure 6-7). In theory, these elements are invulnerable to alteration and, thus, should reflect original

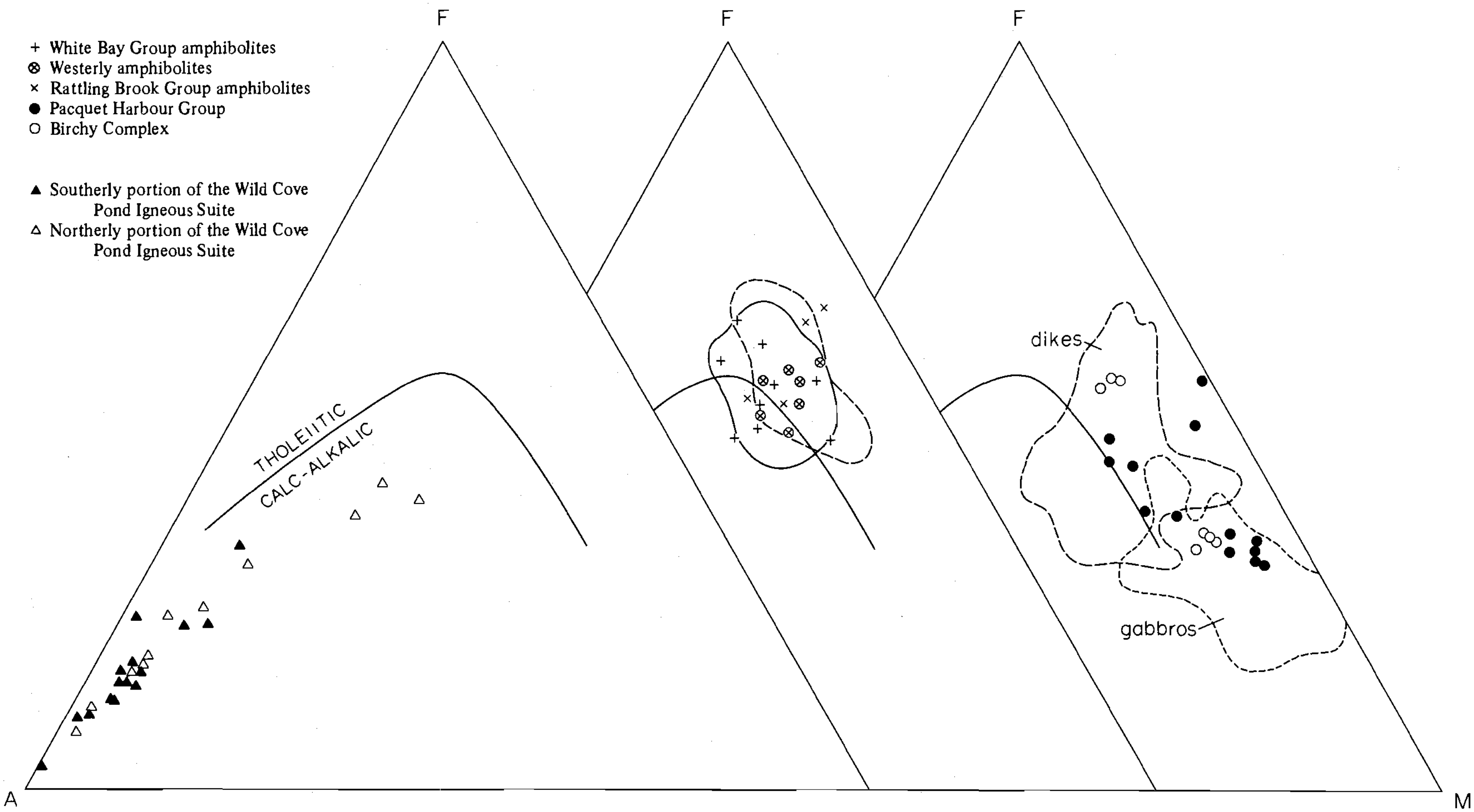


Figure 6-5: *AFM ($Na_2O + K_2O$, FeO^t, MgO) diagrams for all units sampled. Outlined fields are as follows for middle diagram: dashed line is field of Baie Verte Peninsula eclogitic amphibolites (de Wit and Strong, 1975); solid line is field of Long Range dikes and flows (Strong, 1974). For diagram on right, fields are for Betts Cove ophiolitic rocks (Upadhyay, 1974).*

Table 6-3. *Average major element analyses of the westerly amphibolites and comparison with correlative rocks.*

	White Bay Group (7 samples)	Rattling Brook Group (3 samples)	Fleur de Lys amphibolites (7 samples)	Average westerly amphibolites (17 samples)	Fleur de Lys amphibolites (de Wit and Strong, 1975) (10 samples)	Long Range dikes (de Wit and Strong, 1975) (14 samples)	Cloud Mountain flows (de Wit and Strong, 1975) (11 samples)
SiO_2	49.6	48.3	50.4	49.4	47.21	47.09	47.88
Al_2O_3	13.51	14.58	13.79	13.96	13.93	13.99	14.25
TiO_2	2.04	2.01	1.78	1.94	1.64	2.57	2.05
Fe_2O_3	8.65	5.92	4.07	6.21	2.55	2.89	5.56
FeO	4.99	6.86	8.24	6.70	9.48	11.05	7.83
MgO	6.26	6.06	6.74	6.35	6.32	6.10	6.11
CaO	7.65	8.56	12.11	9.44	9.96	10.85	7.97
Na_2O	3.62	2.87	2.16	2.88	2.42	2.29	3.08
K_2O	0.57	0.73	0.53	0.61	0.63	0.53	1.29
MnO	0.22	0.23	0.20	0.22	0.23	0.24	0.22
P_2O_5	0.23	0.21	0.09	0.18	0.26	0.28	0.21
LOI	1.35	2.27	1.01	1.54	2.73	1.48	3.03
AVERAGE TOTAL	98.64	98.60	99.31	98.85	98.36	99.36	99.48

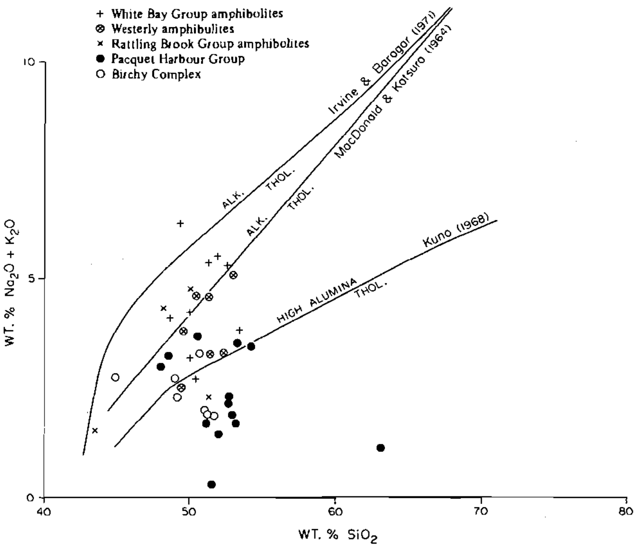


Figure 6-6: *Alkalies vs. silica plot.*

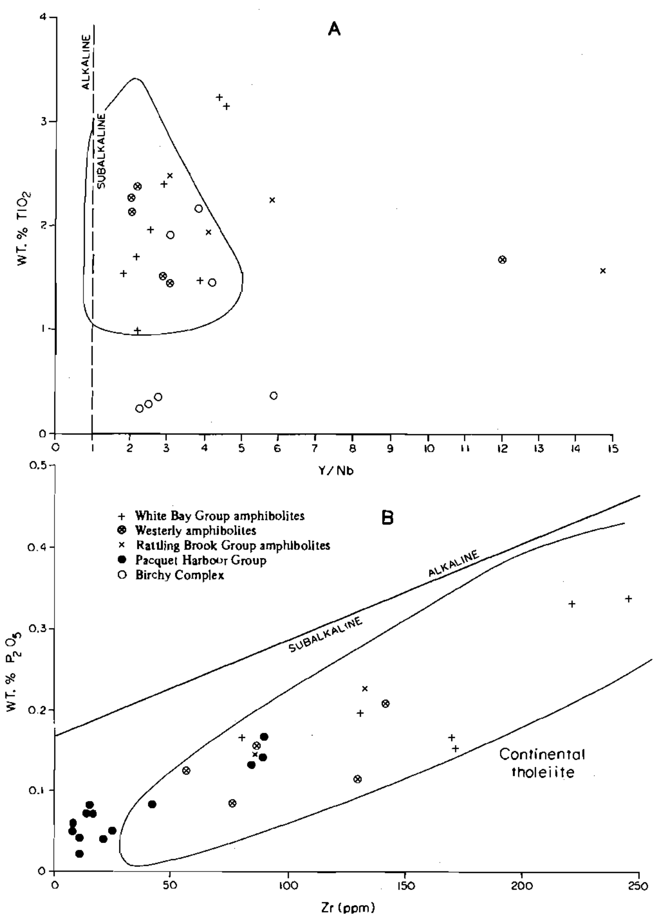


Figure 6-7: *Floyd and Winchester (1975) diagrams for metamorphosed mafic rocks of the Baie Verte Peninsula: continental tholeiite fields outlined on both plots as defined by Floyd and Winchester (1975).* **6-7A:** *TiO_2 vs. Y/Nb diagram.* **6-7B:** *P_2O_5 vs. Zr diagram.*

abundances in spilitized and metamorphosed rocks. On both the TiO_2 vs. Y/Nb and P_2O_5 vs. Zr plots, the Westerly amphibolites show subalkaline affinities; on the former plot, samples 1340047 and 1340057 are omitted due to either extremely low or nondetectable concentrations of Nb, whereas on the latter plot, six samples lack P_2O_5 and have been omitted. On both diagrams, the Westerly amphibolites cluster within the range of continental tholeiites as delineated by Floyd and Winchester (1975). One White Bay Group amphibolite (1340046) plots in the alkaline field on the P_2O_55 vs. Zr diagram; this sample, from the Garden Cove Formation, was collected in an area north and directly along strike from fragmental metavolcanics in the unit and, thus, may represent a volcaniclastic rock. Major element and oxide plots (Jensen cation plot, Figure 6-8; FeO(total)/MgO plot, Figure 6-9) also indicate a strongly tholeiitic character for these rocks.

The geochemical characteristics of the Westerly amphibolites are nearly identical to those of amphibolites and eclogites from the Fleur de Lys Belt sampled by de Wit and Strong (1975) (Table 6-3). De Wit and Strong (1975) demonstrated the geochemical correlation of their rocks with Eocambrian-Cambrian(?) lavas of western Newfoundland (Strong, 1974); a similar correlation is evident for the Westerly amphibolites (Table 6-3). In particular, the Westerly amphibolites overlap the fields of the plateau lavas in the AFM diagram (Figure 6-5), the Zr vs. Ti diagram (Figure 6-10) and the Ni vs. Cr plots (Figure 6-11). In Figure 6-11, some of the Westerly amphibolites show a slight Cr depletion compared with the western Newfoundland lavas; this may reflect either the mobility of Cr during alteration or original abundances in the rocks. In the Zr vs. Ti diagram (Figure 6-10) four of the White Bay Group samples (1340019, 1340021, 1340050 and 1340053) are significantly enriched in Zr with respect to the trends defined by the remaining fourteen Westerly amphibolites and the western Newfoundland plateau lavas. The reason for this enrichment is uncertain, though it may reflect either (i) alteration due to deformation, as three samples are from the area proximal to the Carroll Hills Slide Zone and one comes from the area near the Cabot Fault, or (ii) primary abundances in amphibolites that were originally volcaniclastic rocks.

This geochemical correlation supports the stratigraphic correlation of the western Newfoundland basalts with those of the White Bay Group (see Chapter IV). The western Newfoundland basalts have been related to continental rifting (Bird and Dewey, 1970; Strong and Williams, 1972); trace element data for the Fleur de Lys amphibolites, given above, are compatible with this interpretation.

Birchy Complex

A small number of samples were analyzed from the Birchy Complex, including five metagabbro samples from the South Cove schist and two greenschist samples (Figure 6-1); thus, only tentative conclusions can be made concerning the geochemistry of the complex. These samples plot in the subalkaline field of the alkalies vs. silica diagram (Figure 6-6) and appear to define a tholeiitic trend on both the AFM diagram (Figure 6-5) and the FeO(total) vs. FeO(total)/MgO plot (Figure 6-9). Because these rocks are metamorphosed, the samples have been plotted on a Floyd and Winchester (1975) TiO_2 vs. Y/Nb diagram (Figure 6-7), where they also display a subalkaline affinity.

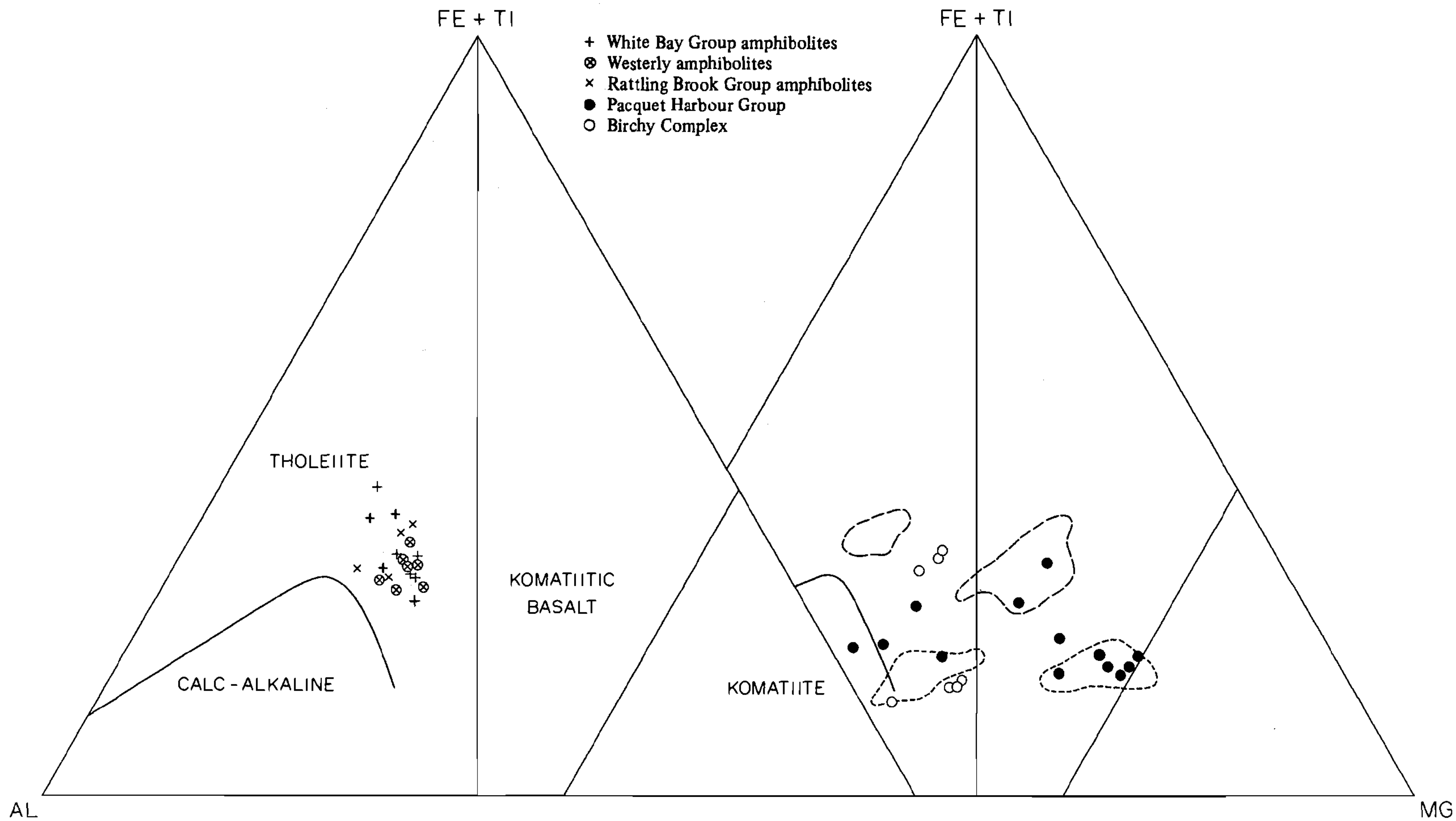


Figure 6-8: *Jensen cation plots for mafic rocks analyzed. Fields outlined for Gale's (1971) Pacquet Harbour Group analyses (long dash) and Upadhyay's (1973) Betts Cove Complex analyses (short dash).*

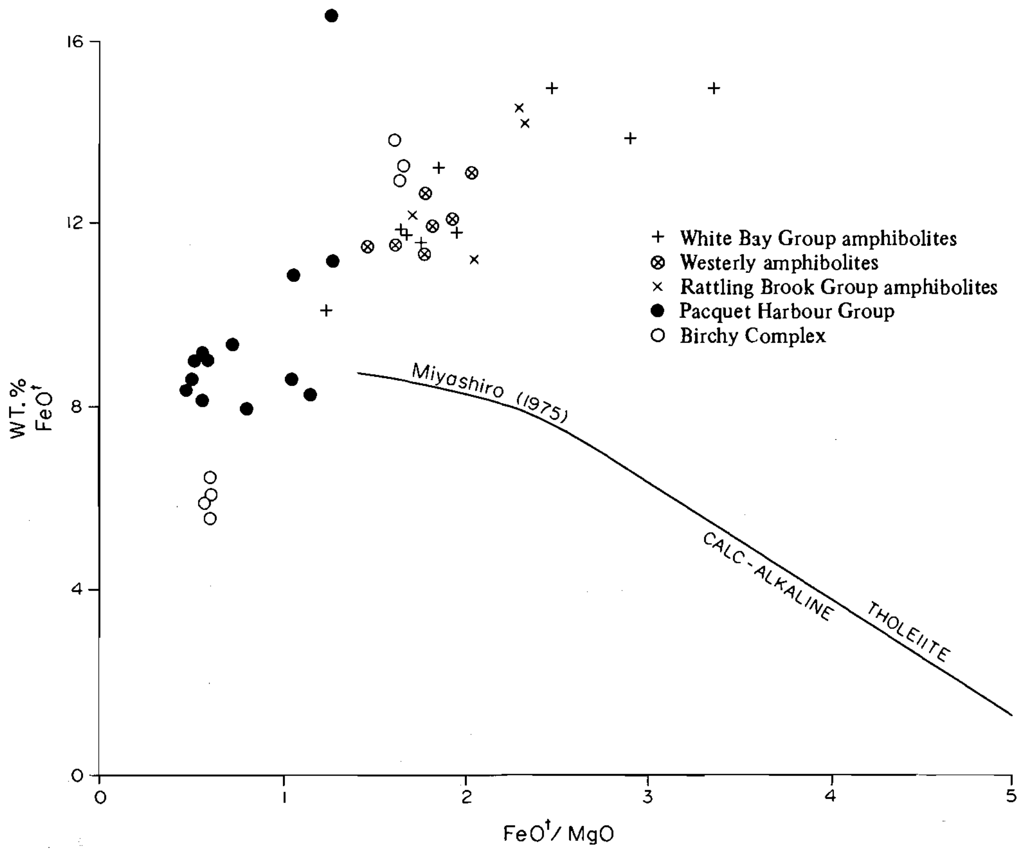


Figure 6-9: *FeO(total) vs. FeO(total)/MgO for mafic rocks analyzed.*

The Birchy Complex metagabbros and greenschists are readily distinguishable from the Westerly amphibolites by the following criteria:

(i) strong bimodalism of the Birchy rocks with respect to TiO_2 content and FeO(total) content;

(ii) higher MgO and CaO content of the Birchy samples;

(iii) lower concentrations of alkalies, P_2O_5, Zr and Y in the Birchy rocks.

Perhaps the most significant finding is that the geochemistry of the complex is similar to that of other ophiolites on the peninsula. The Birchy rocks overlap the Point Rousse ophiolite gabbro and sheeted dike fields (Norman and Strong, 1975) on the AFM diagram (Figure 6-5) and all but one of the metagabbros fall within the gabbro field of the ophiolite (Norman and Strong, 1975) on the Zr vs. Ti plot (Figure 6-10). In addition, the metagabbros of the Birchy Complex are similar to those of the Betts Cove Complex (Table 6-4).

This sparse data thus supports the view, based on stratigraphy, that the Birchy Complex is ophiolitic (see Chapter IV, Birchy Complex).

Pacquet Harbour Group

The Pacquet Harbour mafic rocks are mostly from the area south of Gale's (1971, 1973) investigation (Figure 6-1); the felsic rocks have not been plotted because one is severely altered (1340045) and the other fits into the geochemical fields outlined by Gale (1971) for the Pacquet Harbour felsics.

Thirteen mafic samples collected from the Pacquet Harbour Group include 10 lavas, 2 dikes and 1 tuff. All of the rocks sampled have been totally recrystallized in the middle to upper greenschist facies. The geochemical character of these rocks is summarized in Table 6-5, where the average composition is also presented. Most notably, the Pacquet Harbour rocks display a strong bimodalism with respect to TiO_2 and MgO and, to a lesser extent, Al_2O_3.

Two samples (1340010, 1340018) appear to be altered with respect to the others; they exhibit a severe depletion of alkalies on both the AFM diagram (Figure 6-5) and alkalies vs. silica plot (Figure 6-6), and depletion of CaO on the CaO-MgO-Al_2O_3 (CMA) diagram (Figure 6-12). In addition, sample 1340010 displays an extreme enrichment in iron (Figure 6-9); sample 1340018 is enriched in silica (Figure 6-6) and shows slightly higher concentrations of iron than most other samples (Figure 6-9). Some of the iron enrichment may be attributable to a thick oxidized weathering rind on the sample, possibly incorporated into the rock powder of the specimen. Sample 1340018 is extensively veined with quartz and contains numerous quartz-filled vesicles. Though the mobility of some of the elements of these two samples is extreme, the analyses still reflect some of the geochemical attributes of the group and are, thus, used in the plots discussed below. However, they have not been used in calculating the average analyses presented in tables that follow.

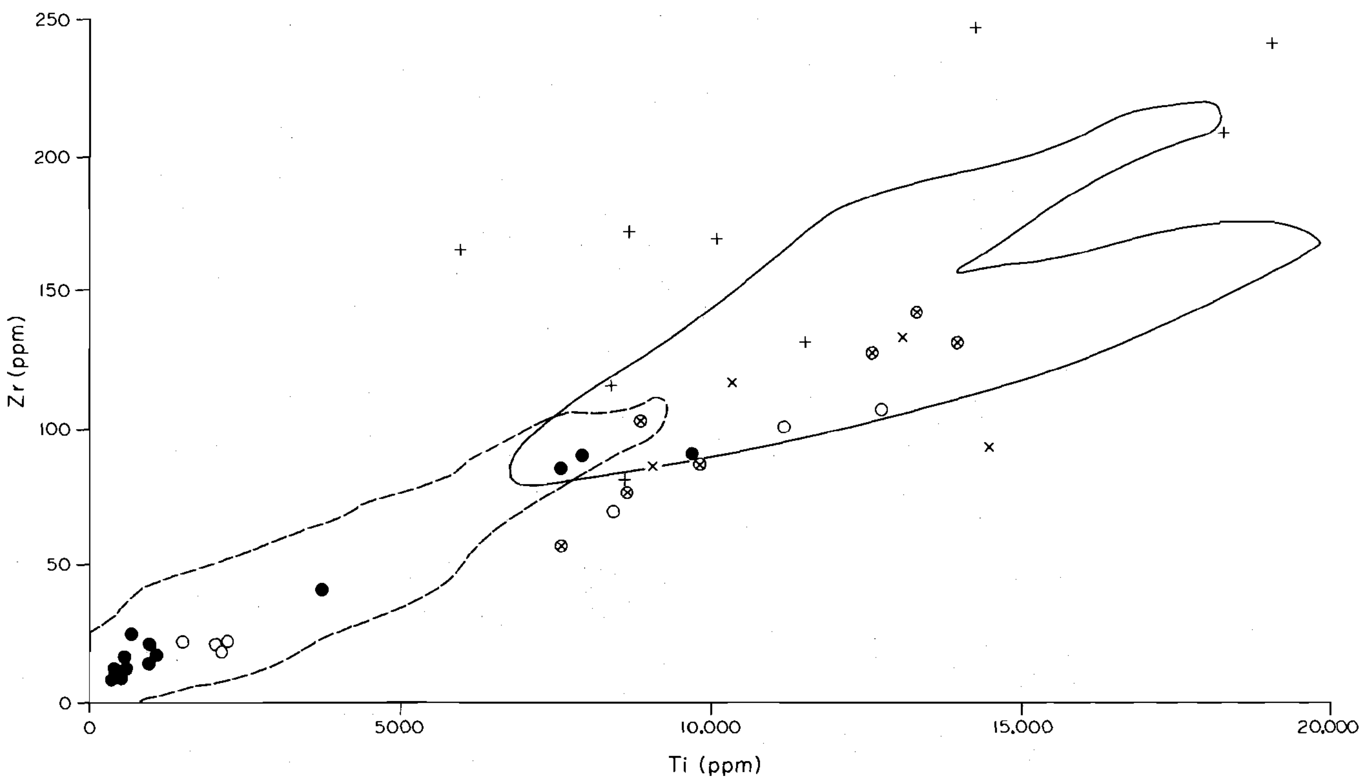


Figure 6-10: *Zr vs. Ti diagram with fields outlined for plateau lavas of western Newfoundland (de Wit and Strong, 1975; solid line) and for Point Rousse Complex gabbros (Norman and Strong, 1975; dashed line).*

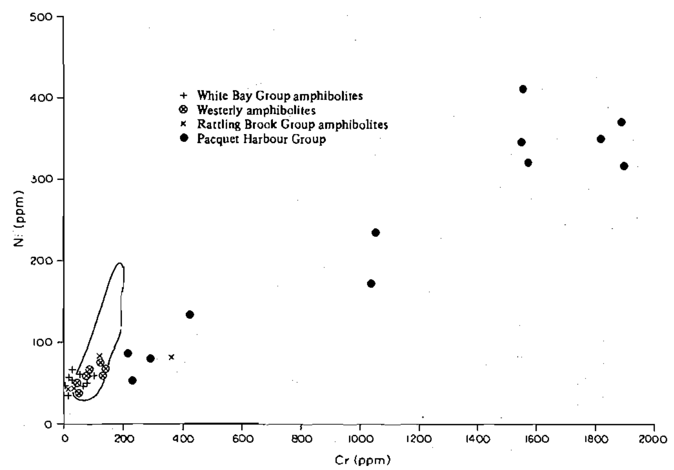


Figure 6-11: *Ni vs. Cr diagram for mafic rocks analyzed; field outlined for western Newfoundland plateau lavas (de Wit and Strong, 1975).*

The samples all display a tholeiitic character on the AFM diagram (Figure 6-5), the alkali vs. silica plot (Figure 6-6), the FeO(total) vs. FeO(total)/MgO diagram (Figure 6-9) and the P_2O_5 vs. Zr diagram (Figure 6-7). However, the Pacquet Harbour Group samples display a distinct bimodalism with respect to both MgO, as best shown on the Jensen cation plot (Figure 6-8), and TiO_2 as displayed on the TiO_2 vs. MgO variation diagram (Figure 6-3). On the Jensen plot, the dikes and one lava appear to be tholeiitic to calc-alkaline, whereas nine lavas plot in the "basaltic komatiite" field; the tuff plots in the tholeiite field but displays characteristics more like the "basaltic komatiite" field with respect to MgO content (see sample 1340007, Appendix III). On the TiO_2 vs. MgO variation diagram (Figure 6-3), the "basaltic komatiites" contain substantially more MgO and less TiO_2 than the remaining samples.

The calc-alkaline trend (Figure 6-8) in the low MgO Pacquet Harbour Group rocks is defined solely by the two dike samples, which display high Al_2O_3 concentrations. The single tholeiitic lava is geochemically very similar to the tholeiites that Gale (1973) reported from the Pacquet Harbour Group (Table 6-5, columns 1 and 2).

The lavas that plot in the basaltic komatiite field of the Jensen diagram are distinguished from the tholeiitic Pacquet Harbour mafic rocks mainly by MgO and TiO_2 contents, yet they also contain more CaO (Figure 6-12), Ni and Cr (Figure 6-11) and less aluminum (Figure 6-12), Zr (Figure 6-10), and P_2O_5 (Figure 6-7) than the tholeiites. These "magnesian" lavas are almost all from the southerly portion of the Pacquet Harbour Group, with the exception of one sample (1340005) from the Rambler Mines area. They are geochemically very similar to the lavas previously termed basaltic komatiites from the Rambler area (Gale, 1971, 1973), though the southerly samples display slightly higher concentrations of MgO and lower iron contents than those reported by Gale (1973) (Figures 6-8, 6-12; Table 6-5).

Table 6-4: *Comparison of Birchy Complex metagabbro compositions with those of the Betts Cove Complex.*

	1	2	3
SiO_2	48.1	50.45	53.6
TiO_2	0.68	0.08	0.10
Al_2O_3	15.02	17.43	15.6
FeO	5.62	4.14	4.77
Fe_2O_3	2.10	1.67	1.32
MnO	0.14	0.11	0.13
MgO	9.66	9.52	9.28
CaO	13.2	8.48	7.12
Na_2O	1.94	2.76	1.97
K_2O	0.15	1.32	2.49
P_2O_5	-	0.01	-
LOI	1.97	3.64	3.40
AVERAGE TOTAL	98.56	99.60	99.63
Zr (ppm)	38	3	ND
Y	16	6	ND
Ni	114	138	ND
Cr	401	243	ND

1. average of 5 samples, South Cove schist, Birchy Complex: this study
2. average of 5 gabbros, Betts Cove Complex: Coish (1977a)
3. average of 2 gabbros, Betts Cove Complex: Upadhyay (1973)

ND - not determined

All of the magnesian Pacquet Harbour lavas are geochemically strikingly similar to lavas from the western Pacific Ocean termed boninites (Cameron et al., 1979) (Table 6-5)[1]. The range of boninite compositions, both petrographically and geochemically, has been shown to overlap that of basaltic komatiites (Cameron et al., 1979); however, boninites, in most cases, have TiO_2 contents less than 0.40% and appear to be associated with ophiolites, whereas basaltic komatiites have slightly higher TiO_2 concentrations and have been considered as characteristic of Archean terranes (Cameron et al., 1979). I prefer the term boninite, rather than basaltic komatiite, for the Pacquet Harbour lavas in order to emphasize their low TiO_2 contents.

The application of the term "boninite" to "magnesian" lavas of the Pacquet Harbour Group is further warranted by their geochemical correlation with other distinctive "magnesian" ophiolitic lavas of the Baie Verte Belt. Pillow lavas geochemically equivalent to the boninites have been reported from the ophiolitic portion of the Point Rousse Complex (Norman and Strong, 1975) and from the Betts Cove Complex (Upadhyay, 1973, 1978a, 1982; Coish and Church, 1979); in addition, komatiitic lavas occur in the Betts Cove Complex (Upadhyay, 1973, 1978a, 1982). All of these "magnesian series" rocks of the Baie Verte Belt are compared in Table 6-6, and the Pacquet Harbour and Betts Cove rocks are plotted in the Jensen and CMA diagrams (Figures 6-8, 6-12) for comparison. It is obvious from these data that the Pacquet Harbour boninites are geochemically correlative with the magnesian lavas of the Point Rousse and Betts Cove Complexes. Boninites of the latter complexes form the lower portions of the pillow lava member in each ophiolite; because of their distinctive geochemical character, identical to the Pacquet Harbour rocks, and their geographic proximity to these rocks, I interpret the Pacquet Harbour lavas as ophiolitic (see also Gale, 1971, 1973), though lower ophiolite members are either unexposed or unrecognized in the group.

Wild Cove Pond Igneous Suite

Twenty-five representative samples of the Wild Cove Pond Igneous Suite were collected and analyzed. As well, one

[1] The Pacquet Harbour lavas are different from Pacific boninites in that the Pacquet Harbour rocks are olivine-normative, whereas most Pacific boninites are quartz-normative. This difference may be due to alteration or SiO_2 depletion in the Pacquet Harbour lavas; however, further study is necessary to document such an alteration.

Table 6-5: *Composition of Pacquet Harbour mafic lavas and comparison with geochemically similar lavas.*

	THOLEIITIC ROCKS		BONINITIC ROCKS			
	This study (1 sample)	Gale (1973) (avg. 3 samples)	This study (avg. 7 samples)	Gale (1973) (avg. 9 samples)	Type locality boninite (reported in Dietrich et al., 1978)	Marianas trench from Dietrich et al. (1978) (avg. 6 samples)
SiO_2	48.3	49.07	51.8	52.83	54.44	55.73
Al_2O_3	15.35	15.39	9.60	9.49	12.90	11.04
Fe_2O_3	2.95	12.18	1.30	10.71	7.08	3.44
FeO	8.06	nr	7.51	nr	nr	5.15
MgO	8.39	7.84	15.70	14.21	12.75	13.29
CaO	6.89	9.70	9.78	10.22	5.12	5.97
Na_2O	3.39	3.61	1.89	2.00	2.06	1.92
K_2O	0.10	0.09	0.13	0.12	0.35	0.84
TiO_2	1.59	1.68	0.12	0.16	ND	0.21
MnO	0.15	0.11	0.14	0.17	ND	0.16
P_2O_5	0.16	0.18	0.05	0.07	ND	0.04
LOI	2.54	*	2.55	*	3.54	3.66
Zr (ppm)	90	149	13	14	ND	33
Y	35	29	5	3	ND	19[1]
Ni	52	43	332	330	ND	280[2]
Cr	227	190	1615	1300	ND	718

nr - not reported
ND - not determined
* - analyses recalculated on a volatile-free basis
[1] - average of only two of six samples
[2] - average of only five of six samples

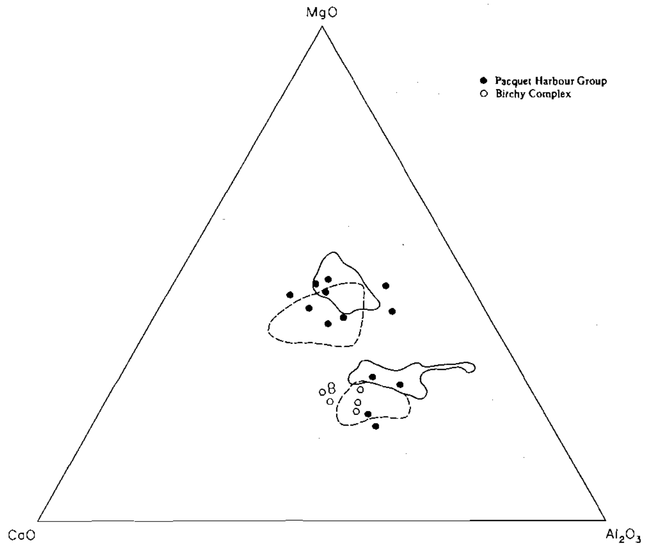


Figure 6-12: *CMA plot for the ophiolitic units with fields for Gale's (1971) Pacquet Harbour Group analyses (dashed lines) and Upadhyay's (1973) Betts Cove Complex analyses (solid lines).*

sample from the Partridge Point Granite was analyzed; because it is similar to those of the Wild Cove Pond Suite, it is included herein with samples from northerly parts of the suite. Most of these samples are corundum-normative granodiorite and granite with only a few more mafic rocks, mainly from the northern part of the pluton (Appendix III).

Figure 6-13 shows that most of the major element oxides display systematic variation trends with respect to SiO_2. Samples from the northern and southern parts of the pluton appear to be geochemically inseparable, but close inspection of variation diagrams reveals that more mafic samples from the northerly part of the body appear to trend toward higher FeO and MnO and lower Al_2O_3 and Na_2O concentrations than the southerly samples. Trace element data support this division. In the northerly part of the pluton, rocks of a given SiO_2 content appear to have lower Sr and Ba concentrations and higher Rb, Y and Nb values than the southerly samples (Figures 6-14, 6-16). In addition, samples from each part of the pluton have coherent and separate Sr/Ca trends (Figure 6-15) that are inexplicable by derivation of one portion of the suite from the other. The Nb/Zr diagram (Figure 6-15) shows that these groups have the same trends, but that the northerly samples are systematically enriched in both Zr and Nb.

In the Nb/SiO_2 diagram (Figure 6-16), the southerly samples are comparable to the field of "crustal melts" whereas the northerly samples, enriched in Nb, overlap on the diagram with "within plate" tin-bearing granites of Bolivia and Nigeria (Pearce and Gale, 1978). The northerly part of the Wild Cove Pond pluton is most similar to the Bolivian granites as they have similar concentrations of Zr and Sn and contain comparable rock types; thus, they may have similar origins.

The geochemical features described above for each part of the suite may be explained by crystal fractionation (cf. Cawthorn et al., 1976), magma mixing (Eichelberger, 1975), or mixing of granitic melt and less silicic restite material (White and Chappell, 1977). The available data are inadequate in quantity and type to evaluate conclusively the role these processes had in the generation of the Wild Cove Pond Igneous Suite, but it is clear (e.g. from Sr/Ca trends) that processes were independent in the northern and southern parts of the suite. In addition, there are some significant differences in trace element concentrations, such as Nb and Zr, which suggest that the parts of the suite have different origins. The most reasonable explanation for all of these features is that the northern and southern parts of the Wild Cove Pond Igneous Suite represent separate magma batches formed by partial melting of different source rocks; variation trends from each part of the suite resulted from differing degrees of contamination with the respective restite material from these sources.

The higher concentrations of Nb, Zr, Rb and K_2O in the northern part, compared to higher Sr and Na_2O in the southern part, may reflect more "sedimentary" versus more "igneous" sources, respectively. However, a more "sedimentary" source would be expected to result in higher concentrations of Al_2O_3 in the northern samples, unlike the observed concentrations (Figure 6-13). Clearly, more geochemical and isotopic data are required before a convincing petrogenetic interpretation of the suite is possible.

DISCUSSION AND SUMMARY

It is apparent from the foregoing data that the units considered in this geochemical study form distinct geochemical groupings which unquestionably aid stratigraphic interpretation in the area. Additionally, the tectonized rocks have grossly retained their primary chemical traits as demonstrated by their coherent groupings and their geochemical similarity to less deformed and metamorphosed units; thus, their geochemistry can aid in deducing the environments in which they formed. Considered in conjunction with other geochemical studies of the peninsula, these analyses can be used to delineate a broad magmatic-tectonic history for the Baie Verte Peninsula.

The earliest magmas on the peninsula appear to be represented by the Westerly amphibolites; they form a coherent geochemical group that is tholeiitic in character. Their composition is nearly identical to that of Eocambrian-Cambrian(?) western Newfoundland plateau basalts, which have been related to the initial rifting of ancient North America immediately prior to the formation of the Iapetus Ocean (Strong, 1974). Based on this correlation a similar tectonic environment is envisaged for the Westerly amphibolites.

Mafic rocks of the Birchy Complex are apparently distinct from the other Fleur de Lys terrane mafic rocks, particularly with respect to TiO_2, FeO(total), MgO, P_2O_5, Zr and Y. The limited number of Birchy Complex samples indicate a closer relationship to the ophiolitic rocks of the Baie Verte Belt; this, in conjunction with stratigraphic relationships in the complex, indicates an ophiolitic origin for the unit.

The Pacquet Harbour Group, though mainly extrusive in character, displays a strong geochemical affinity to ophiolitic rocks of the Baie Verte Belt, in particular the lower ophiolitic pillow lavas at Betts Cove and some of the ophiolitic lavas of the Point Rousse Complex. These lava members all form

Table 6-6: *Comparison of boninitic and komatiitic lavas of the Baie Verte Belt.*

	POINT ROUSSE COMPLEX	PACQUET HARBOUR GROUP		BETTS COVE COMPLEX		
	1	2	3	4	5	6
SiO_2	51.20	52.83	51.8	48.81	52.35	50.68
Al_2O_3	15.92	9.49	9.60	11.06	12.39	9.73
Fe_2O_3	8.98	10.71	1.30	2.22	1.78	1.28
FeO	nr	nr	7.51	6.63	7.00	8.84
MgO	10.02	14.21	15.70	14.57	14.11	19.74
CaO	10.61	10.22	9.78	10.56	9.11	8.54
Na_2O	2.08	2.00	1.89	0.80	2.33	0.33
K_2O	0.63	0.12	0.13	0.24	0.86	0.42
TiO_2	0.40	0.16	0.12	0.15	0.17	0.09
MnO	0.16	0.17	0.14	ND	0.18	0.18
P_2O_5	0.004	0.07	0.05	0.02	0.05	ND
LOI	nr	*	2.55	4.31	3.61	5.28
Zr (ppm)	18	14	13	14[a]	nr	nr
Y	nr	3	5	ND	nr	nr
Ni	166	330	332	297	312	454
Cr	244	1300	1651	873	928	1116

nr - not reported
ND - not detected
* - analyses recalculated on a volatile-free basis
[a] - average of only 4 samples

1. Boninitic lavas - Norman and Strong (1975), average of 5 samples
2. Boninitic lavas - Gale (1973), average of 9 samples
3. Boninitic lavas - this study, average of 7 samples
4. Boninitic lavas - Coish and Church (1979), average of 5 samples
5. Boninitic lavas - Upadhyay (1982), average of 2 samples
6. Komatiitic lavas - Upadhyay (1982), average of 5 samples

a coherent geochemical suite that spans the Baie Verte Belt and is clearly distinct from typical tholeiites by being enriched in MgO and depleted in TiO_2 (Table 6-6). These unusual lavas, herein equated with boninites, appear to form part of a broad spectrum of ophiolitic lavas in western Newfoundland, as first delineated by Church (1977). He noted that, along a west to east section across western Newfoundland, MgO increases whereas TiO_2 decreases within the ophiolitic lavas. This is evident in the Baie Verte Belt (Table 6-6); only a portion of the westerly Point Rousse Complex displays a boninitic character and certain oxides such as alumina display typical tholeiitic values, whereas the intervening Pacquet Harbour Group displays a strongly boninitic character and the easterly Betts Cove Complex contains the high magnesian portion of the spectrum (komatiites). This systematic geochemical change in the ophiolitic lavas across the Baie Verte Belt indicates that these units originally formed portions of a single slab of oceanic crust.

Regional geological relationships indicate that the Baie Verte Belt boninites formed along a consuming plate margin (see Chapter IX). This setting is similar to most modern boninite occurrences, which appear to be within the inner walls of trenches around the Pacific Ocean (Cameron et al., 1979, 1980). The generation of these unusual low TiO_2 lavas appears to require extensive melting of a depleted lherzolite, probably hydrous (Sun and Nesbitt, 1978). The geochemical spectrum across the Baie Verte Belt ophiolites thus appears

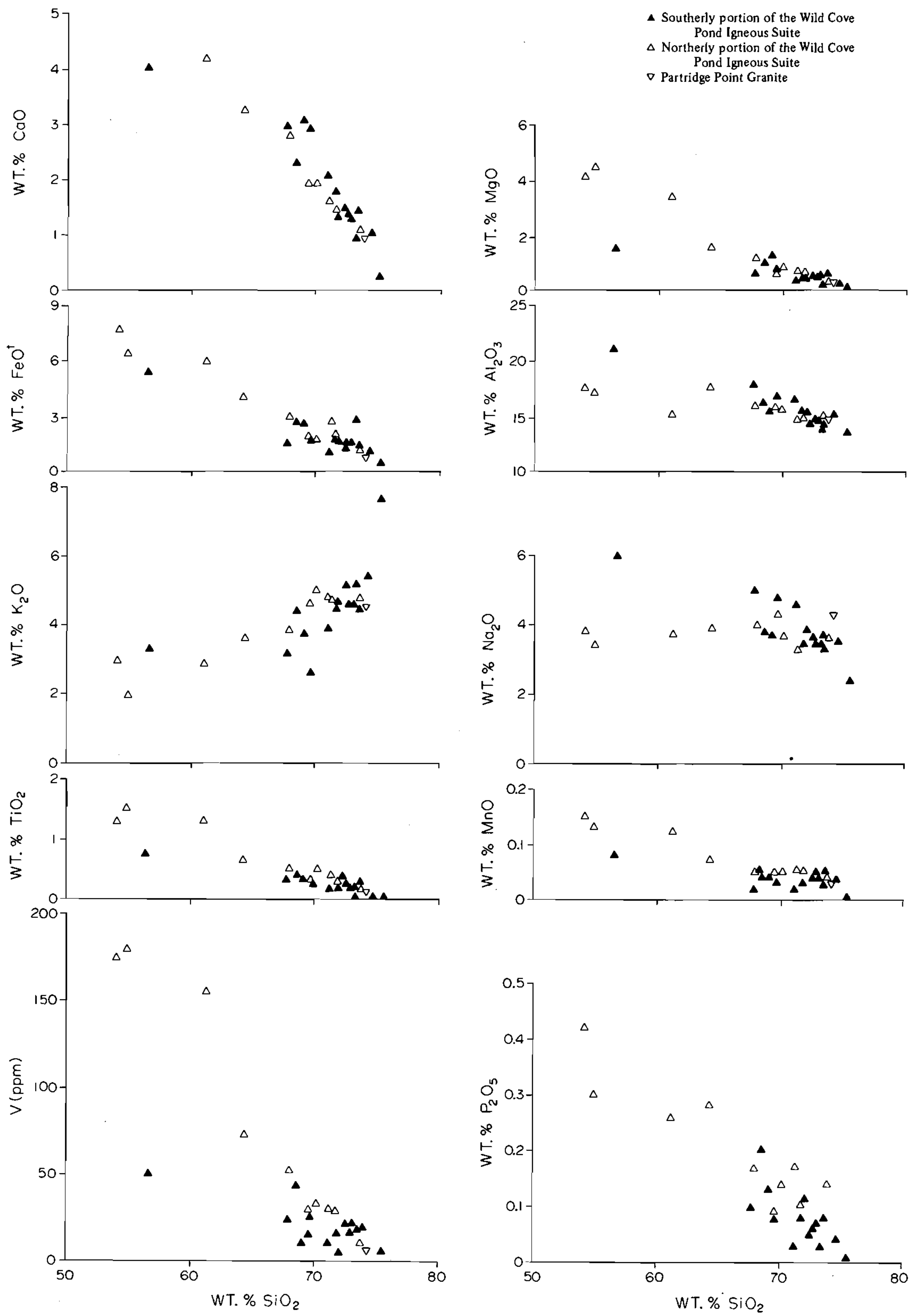


Figure 6-13: *Selected Harker diagrams for the Wild Cove Pond Igneous Suite and Partridge Point Granite.*

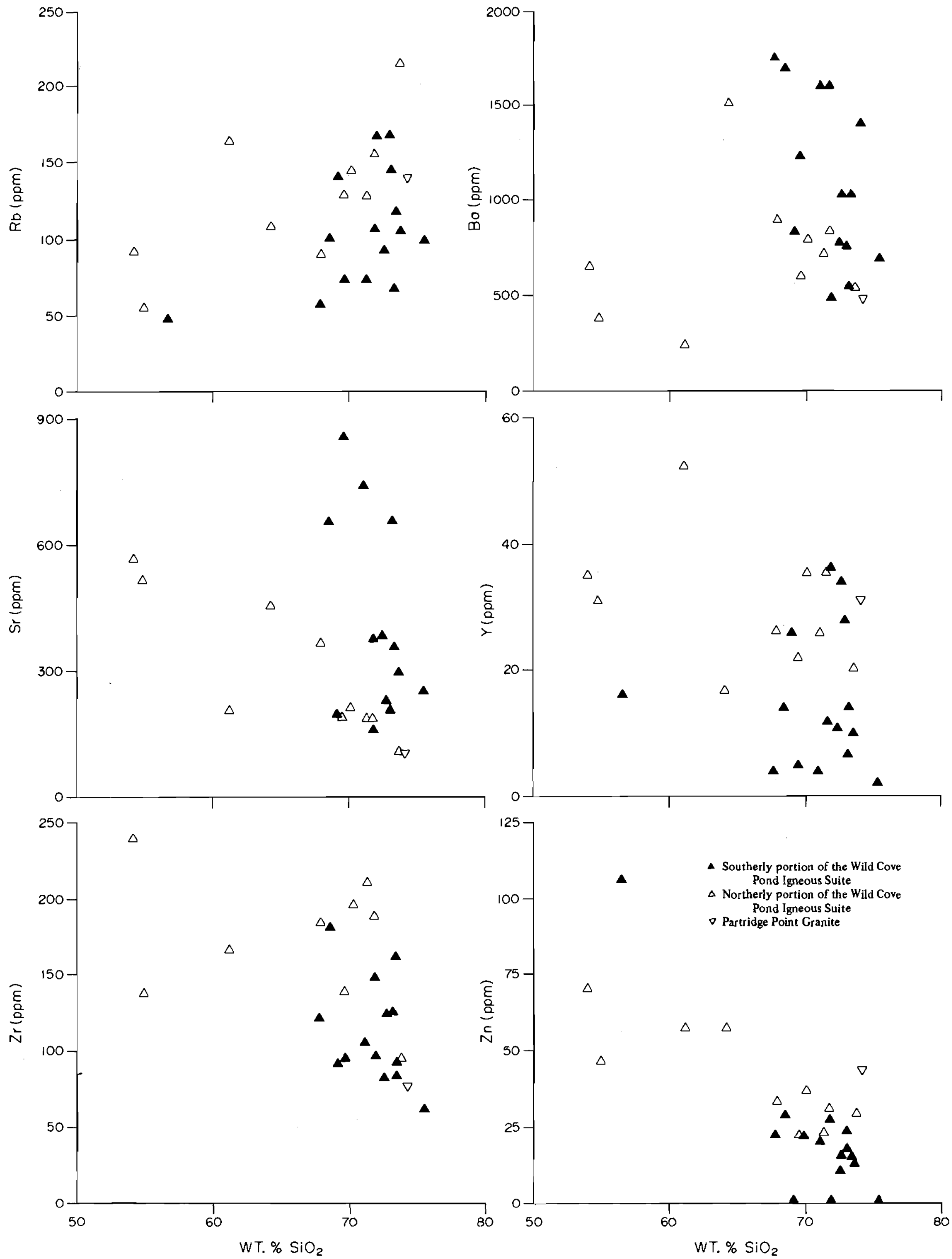


Figure 6-14: *Selected Harker diagrams for the Wild Cove Pond Igneous Suite and the Partridge Point Granite.*

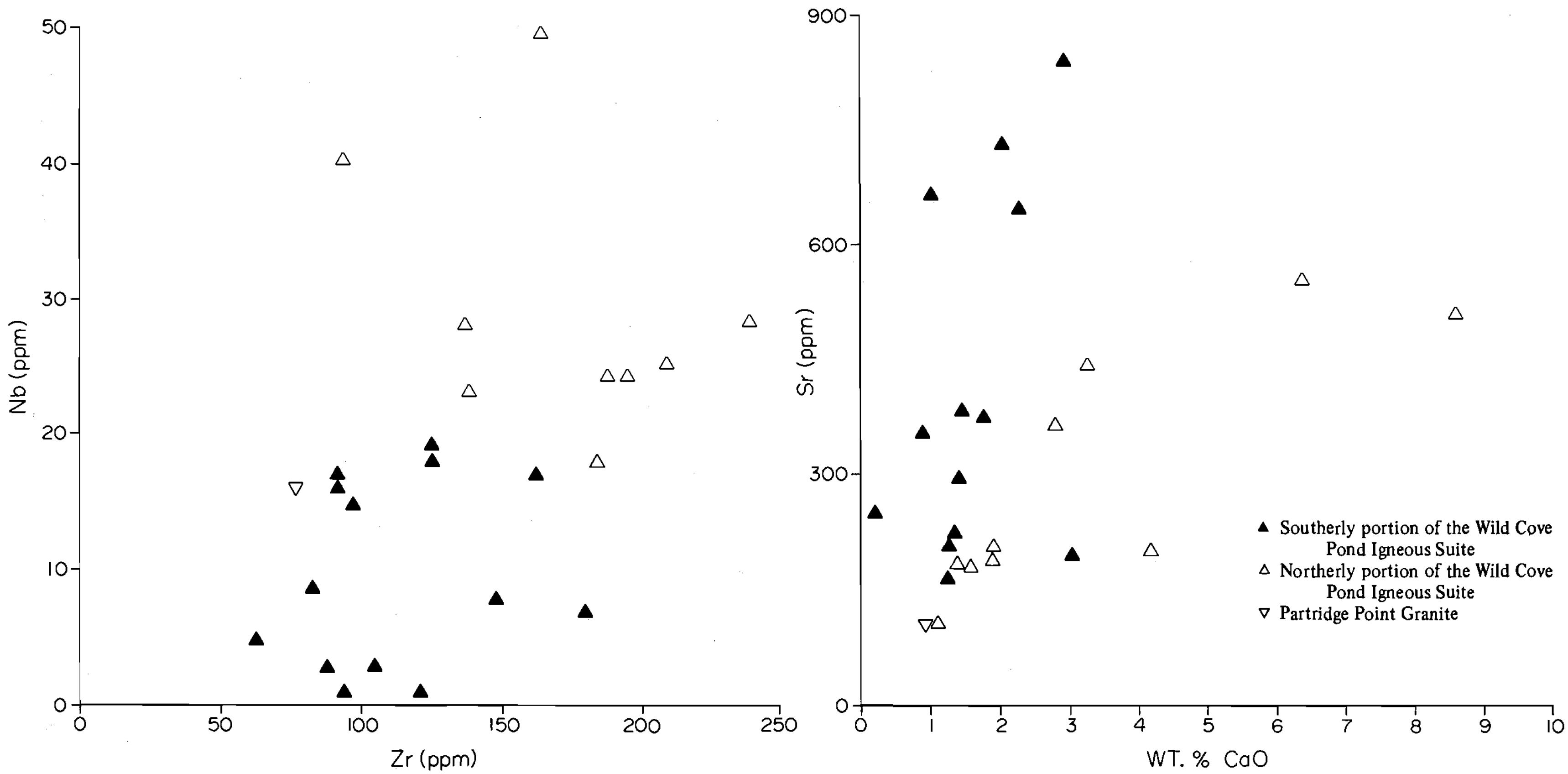


Figure 6-15: *Nb vs. Zr and Sr vs. CaO plots for the granitoid rocks analyzed.*

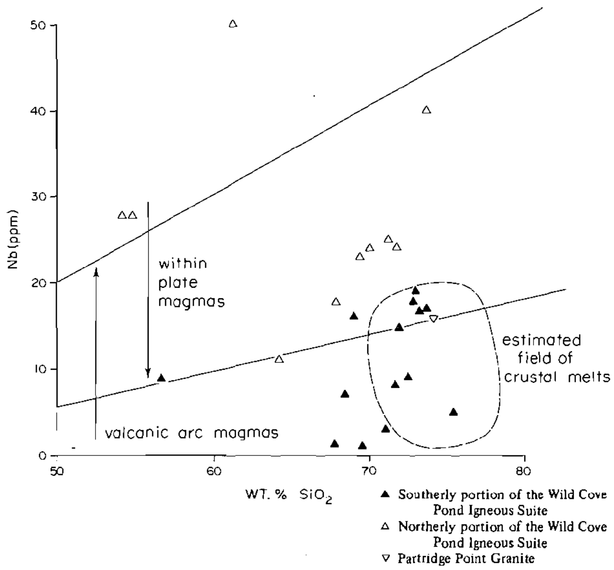


Figure 6-16: *Nb vs. SiO_2 diagram with fields as delineated by Pearce and Gale (1977).*

to reflect increased stages of mantle melting from west to east beneath the fore-arc area. Most significantly, it appears that the eruption of the lavas that reflect this spectrum was approximately synchronous with subduction and the ensuing regional obduction of ophiolites in the area (see Chapters V, VII and IX). These processes were marked by profound faults that extended into the mantle and could have acted as conduits for sea water to reach the mantle or could have introduced water-saturated sediments into the mantle. It appears that the boninitic lavas of the Baie Verte Belt reflect extreme mantle melting due to subduction and/or obduction processes, and that the geochemical spectrum across the belt reflects the stages of melting from earlier (Point Rousse lavas) to later stages (Pacquet Harbour and Betts Cove pillow members). The boninitic lavas may have formed at an oceanic ridge either near or at the subduction site, or within an extensional zone of the fore-arc area during ophiolite obduction. The origin of the typical tholeiites in these sequences remains enigmatic.

This unique magmatic regime, generated at the destructive plate margin, may have had lingering effects on subsequent volcanism; thus, the discrepancy between normal island arc geochemical traits and those of the tholeiites of Snooks Arm Group (Jenner, 1977; Jenner and Fryer, 1980) may reflect late stage effects of the obduction-related magmatism.

Typical calc-alkaline island arc trends of the Cape St. John Group (DeGrace et al., 1976) are in glaring contrast to the boninitic and tholeiitic obduction-related magmatism and indicate a major change in magma regimes. This change suggests a hiatus between older and younger units of the Baie Verte Belt, and is manifest in the Cape St. John - Snooks Arm Group unconformity; it also indicates that a significant break probably separated the Pacquet Harbour volcanism from that of the Cape St. John Group. The late stage calc-alkaline regime appears to have locally evolved into an alkaline to peralkaline system, as it is intruded by the alkaline Reddits Cove Gabbro (DeGrace et al., 1976). The gabbro appears to be associated with peralkaline syenite, porphyry and granite (W.R. Church, unpublished data; DeGrace et al., 1976; see Chapter V).

Late stage magmatism of the Fleur de Lys Belt appears to have been distinct from the alkaline-peralkaline regimes to the east, as the Wild Cove Pond Igneous Suite displays typical calc-alkaline trends. It appears, though, that the suite is not geochemically homogeneous, and that the northerly part of the suite was derived from a source different from that of the southerly part.

CHAPTER VII

STRUCTURE AND METAMORPHISM

INTRODUCTION

The Fleur de Lys and Baie Verte Belts represent the marginal portions of two larger tectonic realms, the Humber and Dunnage zones respectively, of the Newfoundland Appalachians (Williams, 1978a). The tectonic history of the Baie Verte Peninsula centers upon the juxtaposition and interaction of these two realms along the steep, polygenetic structural zone termed the Baie Verte Line (shortened from the Baie Verte - Brompton Line of St-Julien et al., 1976) (Figure 1-1). Both belts as well as the line display arcuate structural trends that swing from a north-northeasterly orientation in the southern portion of the peninsula, to an east-west orientation in the north. This major change in orientation has been termed the Baie Verte Flexure (Hibbard, 1982) (Figure 1-1). Both the Baie Verte Line and Flexure appear to be primordial features that have largely controlled the structural-metamorphic evolution of the peninsula. Generally, all of the rocks to the north and west of the line, i.e. the Fleur de Lys Belt, as well as all portions of the Baie Verte Belt near the line along the east limb of the flexure, have been inhomogeneously polydeformed and polymetamorphosed; the remainder of the peninsula displays features of a paratectonic belt (Dewey, 1969a). The line and the flexure represent composite structures and are discussed in more detail elsewhere in this chapter.

Structural domains locally crosscut the stratigraphic belts of the peninsula; hence, for the purposes of the subsequent description of the structure and metamorphism, the peninsula is separated into the following six tectonic blocks that are based upon tectonic style, age of tectonism, or geographic isolation (Figure 7-1):

- (I) the Western Orthotectonic Block displays dominantly Taconic polytectonism
- (II) the Transition Blocks display polydeformational and polymetamorphic style that stems from probably both Taconic and Acadian events
- (III) the Eastern Orthotectonic Block exhibits dominantly Acadian polytectonism
- (IV) the Paratectonic Block generally shows a single strong phase of penetrative deformation and metamorphism of either Taconic or Acadian age
- (V) the Horse Islands Block is geographically isolated and displays polytectonism of unknown age
- (VI) the Granby Island Block is also geographically isolated and exhibits a single penetrative fabric of unknown age

It is stressed, here, that these structural blocks are not necessarily coincident with the stratigraphic belts outlined in previous chapters (compare Figure 1-1 with Figure 7-1). The Fleur de Lys Belt encompasses the Western Orthotectonic, Horse Islands, and Granby Island Blocks as well as the northernmost part of the Eastern Orthotectonic Block. The Baie Verte Belt includes the Transition and Paratectonic Blocks and the southerly part of the Eastern Orthotectonic Block. The Eastern Orthotectonic Block thus straddles both of the stratigraphic belts.

Each structural block is described independently below with respect to its internal minor structures, major folds and faults, metamorphism, and timing of tectonism. The contiguous structural blocks are separated by major interblock faults, except for the southerly contact of the Eastern Orthotectonic Block with the Paratectonic Block, which is structurally gradational. These major faults, including faults of the Baie Verte Line and the Scrape Thrust and associated faults are described separately, along with structural correlation between blocks. The geometry of all of these structural elements is apparently controlled by the Baie Verte Flexure, which is discussed in a subsequent section. The Cabot and Green Bay Faults appear to be late faults that truncate the geology of the peninsula and are the final structures to be considered in this chapter. Structural correlation and the tectonic history of the peninsula is summarized and discussed at the close of the chapter.

STRUCTURAL AND METAMORPHIC NOTATION

Resolution of the regional tectonic history of an area as complex as the Baie Verte Peninsula depends on a cumulative deductive process, requiring the classification of local scale structures into discrete deformational phases and assessment of the regional significance of these phases. This is somewhat subjective, as noted by many workers (Park, 1969; Williams, 1970; Roberts, 1977; Williams and Zwart, 1977). The present study was carried out with insufficient detail in mapping for reliable correlation of minor structural elements and their assignment to a particular deformation phase. During this study, minor structures have been classified essentially on the basis of style. Thus, in individual outcrops, the most prominent penetrative foliation, i.e. the main foliation, is designated S_M and the folding associated with this fabric is labelled F_M; any secondary fabrics that are associated with folds affecting S_M, i.e. any late fabrics, are collectively termed S_L and related fold episodes are denoted F_L. Where the penetrative fabric is represented by a L>S fabric, it is referred to as L_M or L_L, according to generation. Rarely, a micaceous foliation is found in an outcrop that predates S_M; it is designated S_E, and rare associated folding is noted as F_E. Also, in rare cases in the Western Orthotectonic Block, a gneissic layering and complex fold interference patterns have been noted to predate S_M (see also de Wit, 1972, 1980); in this block these features are all assigned to a basement deformation, D_B, that predates D_E.

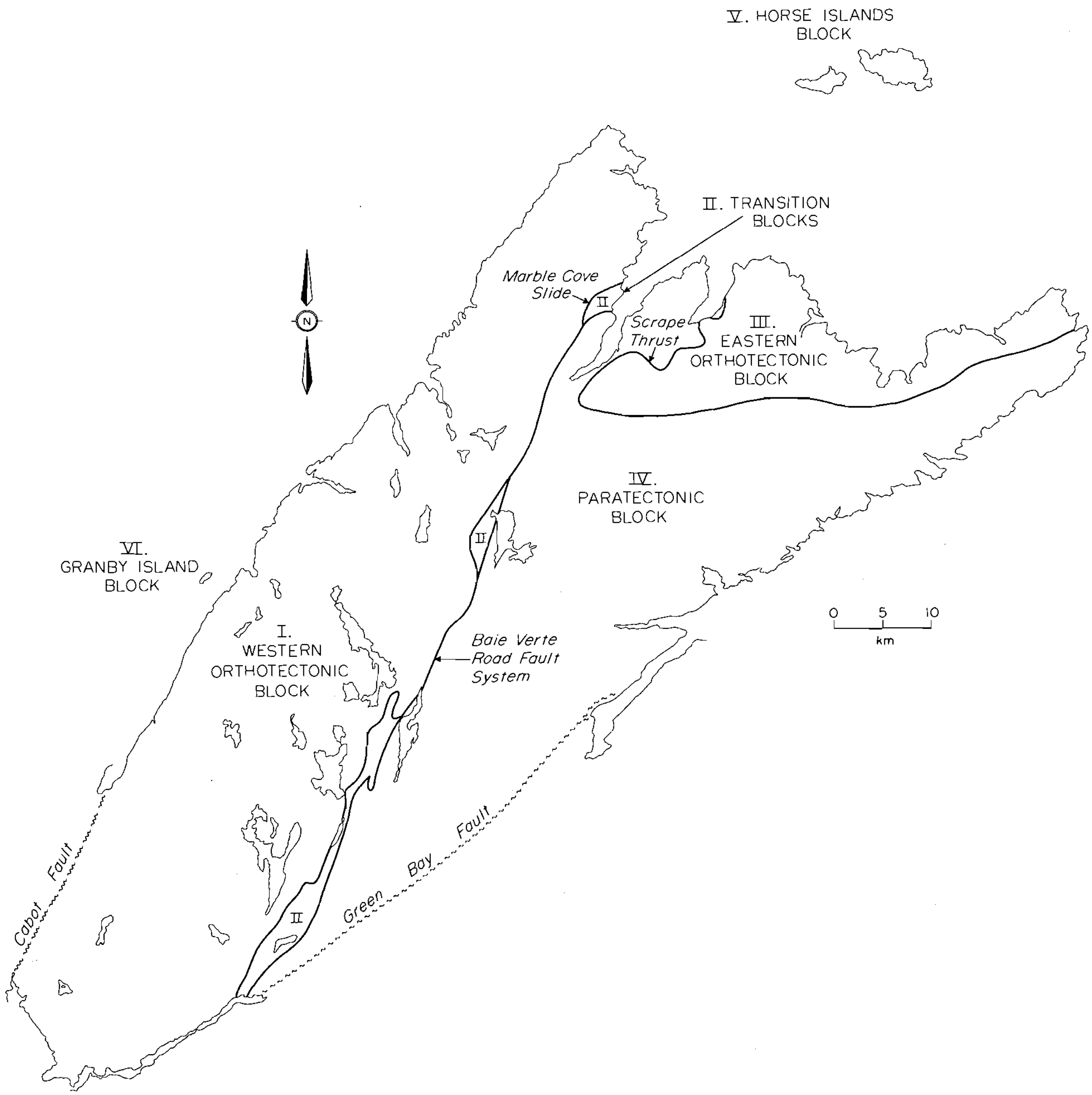


Figure 7-1: *Structural blocks of the Baie Verte Peninsula.*

Generally, D_B, D_E, D_M and D_L structures appear to be correlative over local areas, and in the following discussion I assume that they are correlative within the individual structural blocks except for local areas in the orthotectonic and transition blocks. In these structural blocks, polydeformation has produced complex structural sequences in which there is evidence of both inhomogeneous development of structural styles for individual deformation phases and coaxial deformation. Specific examples of these situations are outlined in more detail in the description of their respective blocks.

In the following descriptions, fabrics are classified according to Hobbs et al. (1976), whereas fold profiles are described according to Fleuty's (1964) classification of interlimb angle and Ramsay's (1967) fold class system based on dip isogons. Fold orientation is described according to Turner and Weiss (1963) and interference patterns resulting from the intersection of two fold systems are classified according to Ramsay (1967).

It has become increasingly evident during the past decade that orogenies are not single climactic events, but rather ongoing processes marked by many episodes of tectonism. The terms Taconic and Acadian are therefore used here in a broad sense. I use Taconic to denote the orogenic events that affected the western part of the Appalachian Orogen during the Ordovician; I use Acadian to refer to deformational and metamorphic events of approximately Middle Devonian age, that affected almost all of the Northern Appalachians.

As with the tectonic history, the assessment of the regional metamorphic history is a cumulative deductive process. The accurate portrayal of the metamorphic history depends upon detailed studies of mineral assemblages with respect to equilibrium and variations of the physical conditions of metamorphism. Because of the emphasis on structural aspects of metamorphic minerals in previous studies and the reconnaissance nature of the present study, a detailed investigation of the metamorphic evolution of the area was not undertaken. However, an effort to document the broad metamorphic evolution of the area with respect to deformational events is attempted. In the following descriptions of the metamorphism of each structural block, metamorphic events that are syntectonic with respect to a given deformation phase are denoted MS, i.e. MS_E, MS_M, etc., whereas metamorphic events that postdate deformation phases are labelled MP, i.e. MP_E, MP_M, etc.. As with structural notation above, this notation refers only to fabrics in a given outcrop; however, these fabrics are considered to be generally correlative throughout each block, except where otherwise noted.

I. WESTERN ORTHOTECTONIC BLOCK

This structural block essentially encompasses the main outcrop area of the Fleur de Lys Belt on the Baie Verte Peninsula. It is tectonically bounded by the Cabot Fault to the west, the Green Bay Fault to the south, and the Baie Verte Road Fault, the Advocate Fault and the Marble Cove Slide to the east (Figure 1-1). All of the rocks in this area are polydeformed and polymetamorphosed with the exception of the Wild Cove Pond Igneous Suite and the Partridge Point Granite. The overall structure of this block is anticlinorial, as first discerned by Murray (Murray and Howley, 1881) and later supported by Baird (1951), Neale and Nash (1963) and de Wit (1972, 1974, 1980). The East Pond Metamorphic Suite forms the core of this structure over which the Fleur de Lys Supergroup is draped. Major subsidiary folds affect the White Bay and Rattling Brook Groups on the flanks of the anticlinorium. Gneissic basement probably of Grenville age underlies the White Bay Group and the East Pond Metamorphic Suite (see below). Gravity studies (Miller and Deutsch, 1976) indicate that continental basement extends under other elements of this block at least as far north as the Little Lobster Harbour Fault (Figure 1-1). North of this fault, the basement to the structural block is uncertain, but it may be ophiolitic since tectonic slivers of serpentinized ultramafics and metagabbro are found in this area (see Chapter IV) and the easternmost unit in the block, the Birchy Complex, appears to be ophiolitic (see Baie Verte Belt, Chapter VI).

Numerous workers have conducted detailed structural analyses on portions of the Western Orthotectonic Block, and some have extrapolated their results throughout the whole block; their work is summarized in Table 7-1. Though there is a discrepancy in the number of deformational phases recognized by different workers, there is general concensus on the overall structural evolution of the area. Based on these studies and observations from the present study, four major tectonic phases are herein recognized for this block, including D_B, D_E, D_M and D_L. The earliest event, D_B, is related to the formation of gneissic structures in the basement, locally exposed in the block. Subsequent phases all relate to the juxtaposition of the Fleur de Lys and Baie Verte Belts; their individual roles during this event are discussed below with each phase.

In Table 7-1, the overlap of D_M with both D_E and D_L stems from the observations of both de Wit (1972, 1980) and Bursnall (personal communication, 1978), as well as evidence in the present study that the main observed fabric in an outcrop may be of different generations across the belt. Thus, it is possible that S_M at one location is actually genetically related to either S_E or S_L at other locations.

BASEMENT DEFORMATION - D_B

Complex structures ascribed to this deformation phase are found in very limited areas in the core of the East Pond Metamorphic Suite and at the western margin of the White Bay Group. They are manifested as a complex gneissic banding on the Westport road (de Wit, 1972, 1980), as migmatites at East Pond (de Wit, 1972, 1974, 1980) and as a gneissosity in the area of Hampden. These structures are significant in that they provide the critical evidence for the interpretation of a predeformed continental basement for at least a portion of this structural block (de Wit, 1972, 1974, 1980; see East Pond Metamorphic Suite).

In the East Pond Metamorphic Suite, the gneissic banding is exposed locally on the Westport road approximately 8 km west of Flat Water Pond. It is characterized by alternating, centimetre scale pale and dark bands that locally display complex intrafolial fold interference patterns as well as simple intrafolial folds. The complex banding has been intruded by a felsic dike that displays the main fabric of the area; hence, this complex banding here predates the main deformation. The migmatitic structures at East Pond have been documented above (see East Pond Metamorphic Suite, Chapter IV); they were clearly affected by the local main fabric at East Pond.

Other areas of migmatite within the East Pond Metamorphic Suite at Pine Pond and southwest of Gull Pond may contain structures that predate the local main fabric, though conclusive evidence is lacking. De Wit (1972, 1974, 1980) contended that banding throughout the psammitic and semipelitic schists of the East Pond Metamorphic Suite, as well as in some amphibolites that are interlayered with well banded metasediments on the White Bay coast, represents tectonic structures that predate the local main fabric. During the present study, no conclusive evidence was found in support of de Wit's ideas. The regular layering displayed by psammitic and semipelitic schists of the East Pond Metamorphic Suite appears to be devoid of intrafolial structures, and graded, coarse pebbly layers within these rocks at the south end of Gull Pond support my contention (see East Pond Metamorphic Suite, Chapter IV) that this layering represents bedding. The amphibolites on White Bay that de Wit considered to contain early structures (see Chapter IV) contain either a layering or a foliation that is askew to the foliation in the surrounding country rocks. Such features could be attributable to local high strain and rotational deformation or to an early phase of deformation; thus the status of these structures remains ambiguous.

Granitic gneiss apparently forms screens in the Oody Mountain Amphibolite (see Chapter IV), and is best exposed

Table 7-1: *Comparison and tentative correlation of detailed structural analyses of portions of the Western Orthotectonic Block.*

This Study (entire Block)	Church (1969) (entire block)	Kennedy (1971) (Fleur de Lys area)	de Wit (1972, 1974, 1980) (central portion of block)	Kidd (1974) (Flat Water-Micmac area)	Bursnall (1975, 1979) (area NW of Baie Verte)	Kennedy (1975a) (entire block)
	F_7 - Conjugate kink folds				Steep late faults δ_5 - Thrust and normal faults	
	F_6 - Close to open conjugate folds	Later faults, kink bands (?), and flat crenulations (?)		D_4 - Minor and large scale kinks; strain-slip fracture cleavage	δ_4 - Coarse S_4 crenulation cleavage, angular F_4 minor folds	D_4 - Minor tight to open F_4 folds, S_4 strain-slip fabric
D_L	F_5 - Mylonitic foliation F_4 - Large scale open folds	Baie Verte Fold and Baie Verte Road Fault	D_L - S_L strain-slip fabric, open to close F_L folds, minor slides, strong lineation	D_3 - Local open to close F_3 folds, poorly developed S_3 strain-slip fabric, local major folds, common lineation	δ_{3L} - Baie Verte Fold, slides and shear zones	
?	F_3 - Crenulation cleavage; tight to close folds	Later strain-slip fabrics and minor folds		D_2 - Tight to isoclinal major and minor folds, S_2 schistosity	δ_{3E} - Close to tight F_3 folds, conjugate S_3 schistosities	D_3 - Minor tight to open F_3 folds overturned to the north, S_3 strain-slip fabric
D_M	F_2 - Crenulation and axial plane transposition foliation; tight to isoclinal folds	Northward facing tight to isoclinal major and minor F_2 folds; S_2 schistosity	D_M - S_M schistosity, major upright F_M folds (open to isoclinal), major slides, high strain zone around the basement		δ_2 - Close to isoclinal major and minor F_2 folds, S_2 schistosity, slides	D_2 - Major NE to NW facing F_2 isoclinal folds, minor tight to isoclinal F_2 folds, S_2 schistosity or L-S fabric
		Minor crenulation of S_1 with open folding (?)				Crenulation of S_1
D_E	F_1 - Bedding foliation; tight to isoclinal folds	Minor tight to isoclinal F_1 folds, S_1 schistosity, D_1 slides	D_E - Small scale recumbent folding, local development of S_E fabric and slides, possible large cross folds	D_1 - Minor F_1 isoclinal folds, S_1 schistosity, slide development	δ_1 - Tight to isoclinal minor F_1 folds, S_1 schistosity, possible large scale cross folding, slides Slides and mélange genesis	D_1 - Minor tight to isoclinal F_1 folds, S_1 schistosity or L-S fabric, slides (?) Tectonic slides
D_B			Deformation and migmatization in basement only Possible very early deformation			

on Rocky Brook. The screens contain a complex gneissic banding that is crosscut by the Oody Mountain Amphibolite; the amphibolite has been deformed with the remainder of the Fleur de Lys Supergroup. Thus, the gneiss must represent a predeformed basement to the supergroup in this area. The White Bay gneisses are closely associated with a metaconglomerate that bears predeformed granitic gneiss clasts; this association is identical to that of the East Pond Metamorphic Suite, and suggests that the gneissic rocks in each area are equivalent (see Chapter IV).

EARLY DEFORMATION - D_E

Structures related to this deformation phase have generally been obliterated by subsequent tectonism, though major D_E structures are resolvable with detailed mapping (de Wit, 1972, 1980; Kidd, 1974; Bursnall, 1975, 1979). The D_E phase appears to have been dominated by slides (Hutton, 1979). These are of paramount importance in the tectonic history of the structural block, for their generation seems to mark the emplacement of ophiolitic rocks into the structural pile and the onset of ensuing tectonism (Bursnall, 1975; Williams et al., 1977; Williams, 1977a). The emplacement of ophiolitic rocks during D_E indicates westerly directed tectonic translation during this deformational phase (Bursnall, 1975).

Minor Structures

Evidence of an early deformational phase that predates D_M is uncommon in outcrop, but it is locally preserved either as a transposed fabric or in fold interference patterns involving F_M. The macroscopic S_E is a micaceous fabric either oriented at high angles to S_M or in F_M closures. S_E is best preserved in the Rattling Brook Group and Birchy Complex north of the Little Lobster Harbour Fault, where it is apparently common (Bursnall, 1975; Bursnall and Hibbard, 1980; Kennedy, 1971, 1975a). In the Birchy Complex, reported Bursnall (1975), S_E is locally well developed, whereas subsequent fabrics are only weakly represented. In the same manner, the S_E fabric is preserved in the Old House Cove Group on the Wild Cove road, between Big Head and Pound Head, west of Fish Point and in the Charity Hill area. S_E is also microscopically preserved in places, as inclusion trails of quartz and mica in garnet and plagioclase porphyroblasts (see also Kennedy, 1971; de Wit, 1972; Kidd, 1974; Bursnall, 1975).

F_E folds are rare throughout the area, though most commonly they are observed in fold interference patterns of types II and III with F_M folds (Kennedy, 1971; de Wit, 1972; Kidd, 1974; Bursnall, 1975). De Wit (1972) reported F_E folds at Southern Arm, in the area south of Crow Head, at the mouth of Middle Arm and on the Westport road. Bursnall (1975) reported them from Slaughter House Cove and Deep Cove, and I have seen these folds in Western Arm. In all cases, F_E folds appear to be isoclinal and approach a class 2 (similar) fold style.

Major Folds

Kennedy (1969, 1971) first speculated that major open folds, predating D_M, are present in the Fleur de Lys area and are responsible for major changes in orientation of F_M fold axes. His thin section studies produced evidence of a pre-D_M crenulation of S_E, which he associated with this major broad folding event. Subsequently, de Wit (1972) and Bursnall (1975) supported the idea that large scale, open folding occurred in association with D_E. On the basis of variation in the orientation of F_M fold axes and the vergence sense of these folds, these workers indicated that the broad, late D_E folds most likely have roughly northwest trending axes.

Major Slides[1]

A number of structural discontinuities predate D_M structures; because of the subsequent intense deformation in most areas, the original nature and attitude of these slides is uncertain. Many of these tectonic breaks are significant, since they are associated with serpentinized ultramafic pods of probable ophiolitic affinity (see Chapter IV) and are most likely related to the emplacement of these rocks. Within most of these zones, it appears that the original structural discontinuity and emplacement of the ultramafics predate S_E and that these structures were subsequently intensified during the formation of S_E. The slide zones containing ultramafic pods are all found in the east half of the structural block. Some of the better recognized slides and related zones are described below.

SLAUGHTER HOUSE SLIDE AND COACHMAN'S MELANGE

The Slaughter House Slide was first delineated by Bursnall (1975) between Slaughter House Cove and Coachman's Cove (Figure 7-2). The slide is marked by isolated serpentinized ultramafic bodies and is in close proximity to ophiolitic mélanges of the Birchy Complex (Bursnall, 1975; see Chapter IV) (Figure 7-2), informally termed the Coachman's mélange (Williams, 1977a). Bursnall (1975) described the aspect of the slide in the Slaughter House Cove area, using his own structural notation (see Table 7-1):

> It can be demonstrated that the zone traverses the local stratigraphy. A pre-S_1 age for this structure is difficult to demonstrate throughout its length since reactivation during subsequent deformation events has resulted in the partial obliteration of the primary relationships. In Slaughter House Cove, however, the zone is locally folded by F_1 and S_1 can be traced (across later structural breaks) from the enveloping schists, through an actinolite-bearing talc-carbonate reaction margin into the periphery of the serpentinite body and associated amphibolite (?pyroxenite). The small serpentinite to the north contains a pronounced and complexly folded S_1 planar schistosity.

Bursnall (1975) noted that the S_E fabric within this slide is very intense; thus the early break probably served as a locus for high strain during the formation of D_E minor structures.

The Coachman's mélange, as described above (see Birchy Complex, Chapter IV), predates all of the regional fabrics and contains large blocks of serpentinized ultramafic rock. The close association of the early slide containing Alpine-type

[1] The term "slide" as defined by Hutton (1979) is used in this report, i.e. "...a fault which forms in metamorphic rocks prior to or during a metamorphic event. It occurs within a zone of coeval penetrative (i.e. microscopic) deformation that represents an intensification of a more widespread, often regionally developed deformation phase. Within this zone of high strain slides may lie along and be sub-parallel to (although they will cross-cut on a large scale) the boundaries of lithological, tectonic and tectonic-metamorphic units."

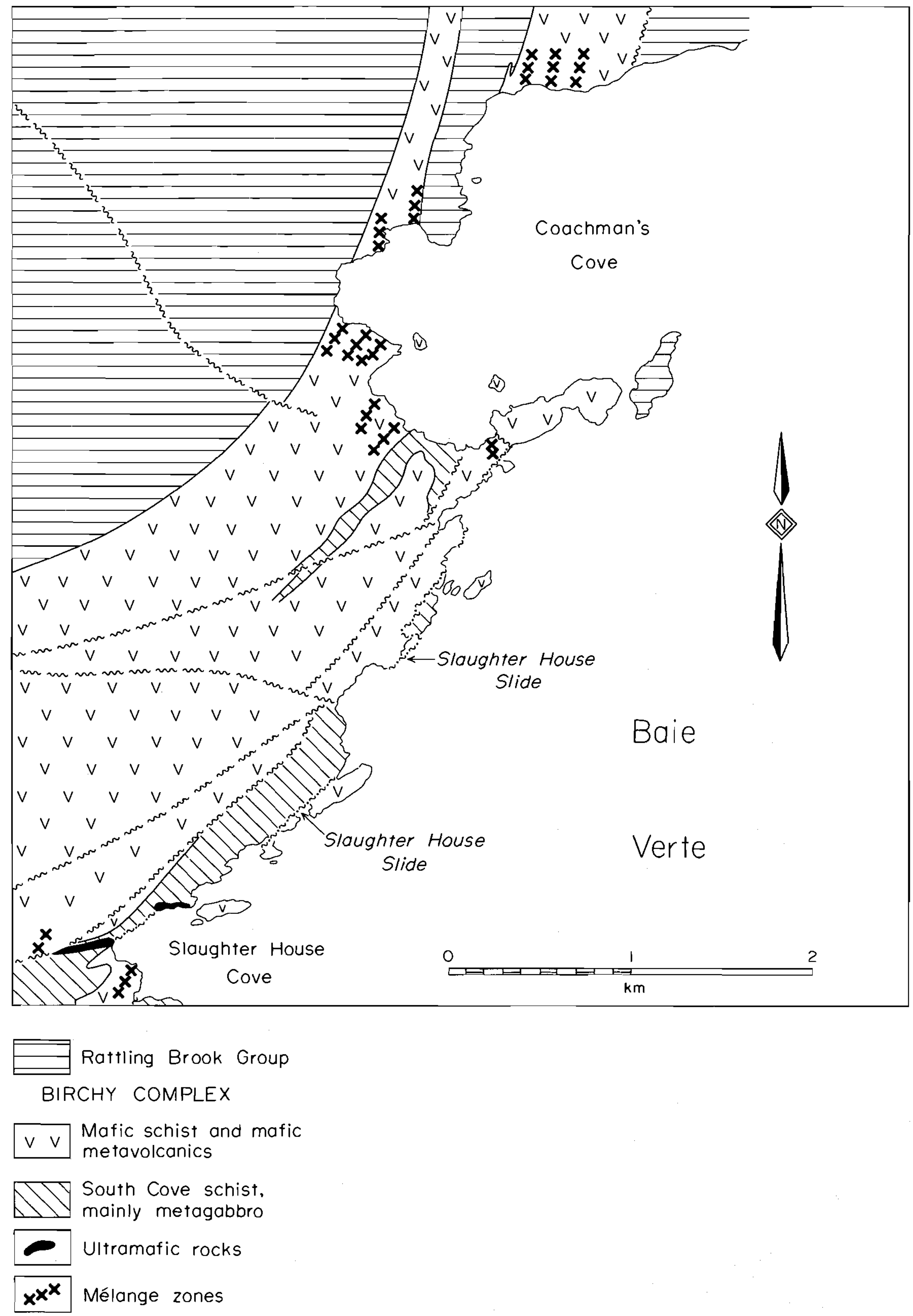


Figure 7-2: *Distribution of the Slaughterhouse Slide and D_E mélange zones (from Bursnall, 1975).*

ultramafics with the predeformational ophiolitic Coachman's mélange indicates that these structures are related and that they mark the early tectonic emplacement of ultramafic rocks into the local stratigraphy (Bursnall, 1975, 1979; Williams, 1977a; Williams et al., 1977). Bursnall (1975) speculated that the northward extension of the slide may be delineated by the Coachman's mélange in the Coachman's Cove area.

Bursnall (1975, 1979) also conjectured that to the southwest, the Slaughter House Slide may extend inland and be responsible for the local emplacement of larger ultramafic bodies west of the South Cove schist (Figure 1-1); however, poor exposure in this area precludes any definite interpretation.

BISHIE COVE SLIDE

This tectonic break was first noted by Kennedy (1969, 1971) on the east side of the marble exposed along the coast between Bishie Cove and Cook In Cove (Figure 1-1). It extends inland to the Lead Mine Fault (Figure 1-1) where it is offset to the west and continues southward along the west side of metagabbro and serpentinized ultramafic to Fleur de Lys Harbour; south of the harbor, it disappears beneath an east striking thrust fault. According to Kennedy (1971), the slide transects the local stratigraphy and is marked by a fine grained, flaggy quartzite. The quartzite displays discontinuous bands that resemble mylonitic banding, though mylonitic textures are absent (Kennedy, 1969, 1971); most likely these textures were obliterated by subsequent tectonization. Banding in the quartzite has been folded by F_M folds. The localization of mafic-ultramafic assemblages along this zone suggests that the slide is related to the emplacement of these bodies, similar to the setting of the Slaughter House Slide.

Bursnall (personal communication, 1980) mentioned the possiblity that the linear zone of ultramafic bodies in the central portion of the Fleur de Lys Peninsula is located along an extension of the Bishie Cove zone; further detailed work in this poorly exposed terrane is necessary to document this notion.

ROADSIDE SLIDE

Kidd (1974) recognized this slide along the west side of the Baie Verte highway between Micmac Lake and the granitic satellite of the Wild Cove Pond Igneous Suite (Figure 1-1). He described the zone using his own structural notation (Table 7-1):

> The slide zone is composed of strongly-foliated semipelitic schist, with subordinate graphitic semipelitic schist and mafic schist mixed in a zone about 100 metres wide. In outcrop, some parts of the zone have a gneissic appearance with regularly-spaced highly-segregated folia of quartz-plagioclase, and mica, including a large proportion of orange biotite relative to muscovite. In other outcrops on the margin of the slide zone, bands of monomineralic orange-biotite schist a few centimetres thick occur, best seen in the roadcut 0.6 km WNW of the south end of Slink Pond. However, the most striking, consistent, and diagnostic feature of this slide zone is the presence of abundant small lenticular bodies up to 1 metre long of tremolite-fuchsite rock.
>
> These bodies are usually strongly-foliated, but some are statically recrystallized with coarse sheaves and rosettes of acicular tremolite. These tectonic lozenges are seen in outcrop folded by F_2 minor folds. Also minor F_2 folds of the intense foliation elsewhere in this and other outcrops indicate that the development of the slide zone is of D_1 age, and the emplacement of the meta-ultramafic tremolite-fuchsite lenses is of this age or earlier. Also seen in outcrop is a large (30 x 5 metres) tectonic lozenge of metagabbro, and some of the mafic schist in this outcrop, especially in the leucocratic varieties, has been derived from this metagabbro.

Thus the Roadside Slide shows characteristics similar to the Slaughter House and Bishie Cove Slides.

OTHER D_E SLIDES

In addition to the early slide zones associated with ultramafic rocks, Kennedy (1969, 1971), de Wit (1972), Kidd (1974) and Bursnall (1975) recognized other D_E slides.

De Wit (1972) considered some whitish gray quartzites in the Old House Cove Group as representing D_E slides, since they locally display a discontinuous glassy mylonite-like banding where recrystallization has been minimal; this banding is folded by isoclinal F_M folds (de Wit, 1972). Where recrystallization is complete, these bands are characterized by clusters of porphyroblastic garnet. De Wit (1972) reported that in Western Arm, one of these bands contains a small amount of fuchsite. He also reported a similar D_E slide from the Penny Hills slide zone, near Penny Hills (see D_M, below).

Kidd (1974) interpreted the contact between rocks of the present East Pond Metamorphic Suite and the Rattling Brook Group as an unexposed slide, the Noseeum Slide. His evidence for the slide was the abrupt nature of the boundary, the intensely deformed nature of outcrops closest to this boundary, and the presence of an angular, mylonitic psammitic schist boulder on an outcrop very near this contact. Kidd (1974) proposed a D_E age for this slide based on the probable relative age of the fabric in the boulder. Considering the absence of the Old House Cove Group between the groups in this area, a tectonic contact is very likely.

Two D_E slides reported by Kennedy (1969, 1971), the Pardee Slide and the Birchy Slide, were sought during the present study, but both remain unrecognized. Kennedy (1969, 1971) reported the Pardee Slide from the cove immediately west of Lower Pardee Cove, and traced it discontinuously inland to the area west of Slaughter House Cove. He interpreted the Birchy Slide to be discontinuous and to separate the Birchy greenschists from the Flat Point Formation on the peninsula immediately north of Coachman's Harbour. Bursnall (1975), though, reported a D_E-slide-zone-like quartzite from the inland area west of South Cove and entertained the possibility that it represents an extension of the Pardee Slide.

MAIN DEFORMATION - D_M

The major structural framework of the Western Orthotectonic Block was formed during D_M, the main deformation. Following the D_E emplacement of apparent ophiolitic rocks into the structural pile of the block, the belt was folded on a major scale during D_M; this deformation imparted a dominantly northeast trending structural grain to the block. Tectonic translation during D_M, as during D_E, appears to have been from east to west (Kennedy, 1971; Bursnall, 1975).

Minor Structures

Typically, minor structures related to this phase are manifested as a main fabric, S_M, with associated minor F_M

folds, though more complicated features related to D_M have been noted (Kennedy, 1971; de Wit, 1972; Bursnall, 1975) and are best summarized by de Wit (1972):

> The recognition in several places of continuous refolding during D_M has led to the conclusion that rotational strain during the progressive deformation is at least locally of considerable importance. The probability arises that, during one period of deformation, discontinuous episodes of folding in one place may be continuous elsewhere. Locally, evidence of transposition of the main S-plane (S_M) caused by D_M, is found. Folds, related to the main deformation, commonly have simple relationships with this main schistosity, but sometimes, no regularity is seen even within one outcrop. It is clear that the main deformation is a complex event....

In addition, the main structures observed in outcrop in this structural block change in character from the East Pond Metamorphic Suite to the Fleur de Lys Supergroup; more subtle changes in the D_M minor structures are also evident across the strike of the supergroup.

In the East Pond Metamorphic Suite, D_M is expressed almost exclusively as a strong, steeply dipping, penetrative fabric, S_M; S_M varies dramatically in intensity and ranges from a pure S-fabric to a pure L-fabric. These variations in the main fabric are most readily noticed in the Middle Arm metaconglomerate. On the west side of Middle Arm, the metaconglomerate contains plump rounded clasts and displays only a weak schistosity; however, in places along the Bear Cove road, the clasts are so intensely flattened that a gneissic banding is produced, and at the northwest corner of East Pond, the clasts have been stretched so much that their shape resembles a cigar. Folds related to D_M are rare in the suite, and where seen they are tight to isoclinal class 2 and class 3 folds.

S_M is very intensely developed at the interface between the East Pond Metamorphic Suite and the Fleur de Lys Supergroup, and has resulted in the formation of tectonic schists. These schists (see Chapter IV) are apparently the result of intense flattening rather than shearing (see above) (de Wit 1972).

The main fabric in the Fleur de Lys Supergroup is generally a moderate to steeply dipping penetrative schistosity associated with tight to isoclinal F_M folds; it is commonly parallel or subparallel to compositional layering, and trends northeasterly. S_M is mostly a planar fabric, though typically in massive psammites of the Old House Cove Group it is represented by a L-S fabric with a mineral lineation of quartz and feldspar. In many places where S_E is present, particularly in the area north of the Little Lobster Harbour Fault, S_M is manifested as a crenulation cleavage. S_M appears to be heterogeneously developed throughout the block (see also de Wit, 1972; Bursnall, 1975).

As noted above, the regional S_M fabric is poorly developed in portions of the Birchy Complex, thus allowing the regional S_E fabric to become S_M in these areas (Bursnall, 1975). A similar change in generation of the main observable fabric is suggested by evidence in the present study and was postulated by de Wit (1972) and Bursnall (personal communication, 1978) for the extreme eastern portion of the structural block; here, however, the change involves S_M and S_L. Along the Wild Cove road, a late, spaced crenulation cleavage, S_L, becomes more intense and more penetrative in an easterly direction, and in the Rattling Brook Group it appears at least locally to transpose a very strong earlier fabric probably correlative with the main schistosity to the west. Thus, the S_L crenulation cleavage appears to become the main observable foliation in an easterly direction across the structural block. Similar observations were made by de Wit (1972) to the south of Wild Cove road and by Bursnall (personal communication, 1978) in the area north of Baie Verte. De Wit reported that, near the Baie Verte Road Fault, there is a single strong fabric in the Birchy Complex which he suggested is correlative with S_L to the west. Unfortunately, the extent and nature of this change in generation of S_M could not be traced reliably, due to poor exposure and the reconnaissance nature of this study. There is evidence in the Wild Cove road area, though, that this change is zonal, because in some places, a spaced S_L crenulation cleavage only mildly affects a strong S_M fabric without any trace of earlier fabrics.

In addition to changes in generation of S_M, it appears that S_M has been locally intensified after its formation and prior to D_L. In places, S_M has been tightened by later D_M movements, such that F_M folds are completely transposed along S_M surfaces and locally S_M itself appears to be transposed by S_M (Plate 7-1).

Minor folds associated with S_M are typically tight to isoclinal; many of these are class 2 folds, though all other classes are found (Plates 7-2, 7-3). Bursnall (1975) showed that the variation in these fold styles can be attributed to variations in folded rock types, ductility, position of the fold in the structural pile, and the inhomogeneity of D_M strain. In most areas, F_M minor folds are gently to moderately plunging (Kennedy, 1971; de Wit, 1972; Kidd, 1974; Bursnall, 1975), though in the area north of Baie Verte, these folds are steeply plunging due to their position on a D_L major structure (see below). Kennedy (1969, 1971) postulated the presence of north facing F_M folds based on a few younging directions in the folded strata. This model has remained unconfirmed in the present study. F_M folds unaffected by D_L are rare, and generally they form complex interference patterns with F_L folds; most common are type III patterns (Plate 7-4), though type I and II patterns have been noted.

Minor D_M tectonic structures that appear very similar to sedimentary structures have been documented by de Wit (1972); they occur at the interface of two layers of contrasting ductility and have an appearance like load casting and crossbeds (Plates 7-5, 7-6, 7-7). These structures are apparently formed by buckling at the interface during compressive stress parallel to layering; continued stress generates minor folds, i.e. generative or drag folds (Dewey, 1967; Ramsay, 1967) with curved axial planes (Plate 7-6) (de Wit, 1972). These structures appear to be a small detail of the D_M history, but are mentioned here in order to emphasize that they are tectonic structures. The misinterpretation of these structures as sedimentary features could lead to the erection of a false stratigraphic succession.

Boudinage also accompanied D_M, and is best represented by the disrupted amphibolite pods in the Old House Cove Group. This boudinage is two-dimensional in the S_M plane.

Major Folds

Most major scale folds depicted in the Western Orthotectonic Block in Figure 1-1 are related to D_M. All of these folds are defined primarily on the basis of the map pattern of rock units.

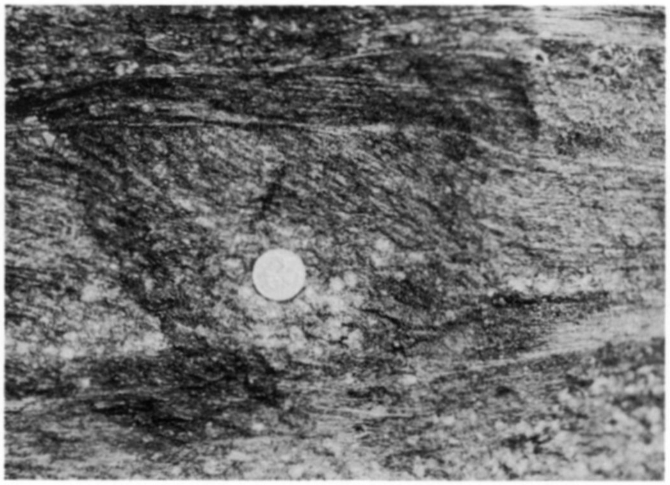


Plate 7-1: *S_M fabric refracted and transposed by S_M itself; semipelite of the Old House Cove Group at Hard Bay.*

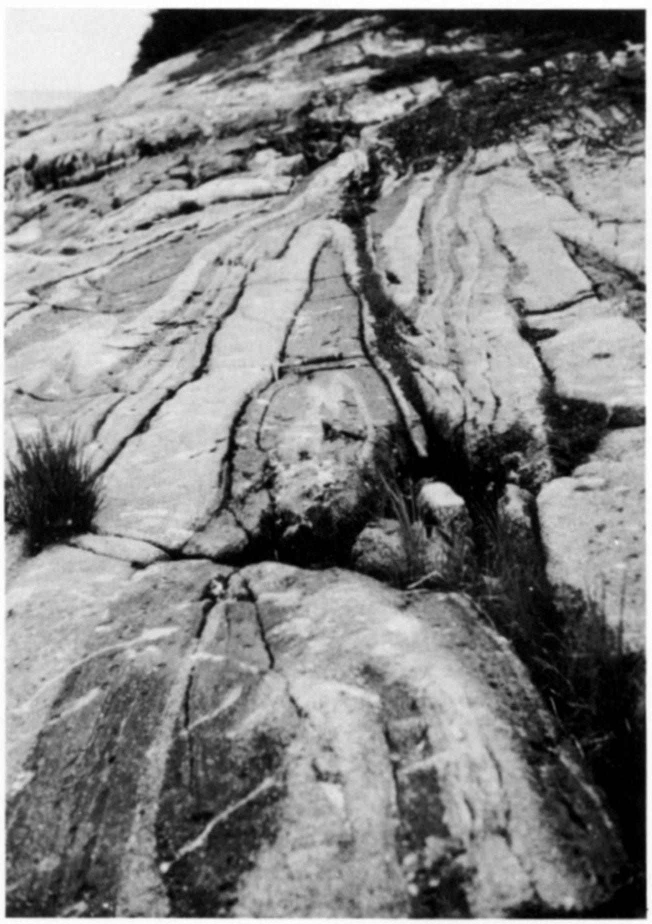


Plate 7-2: *Very tight F_M minor fold in the Old House Cove Group metaclastics near the mouth of Middle Arm, White Bay.*

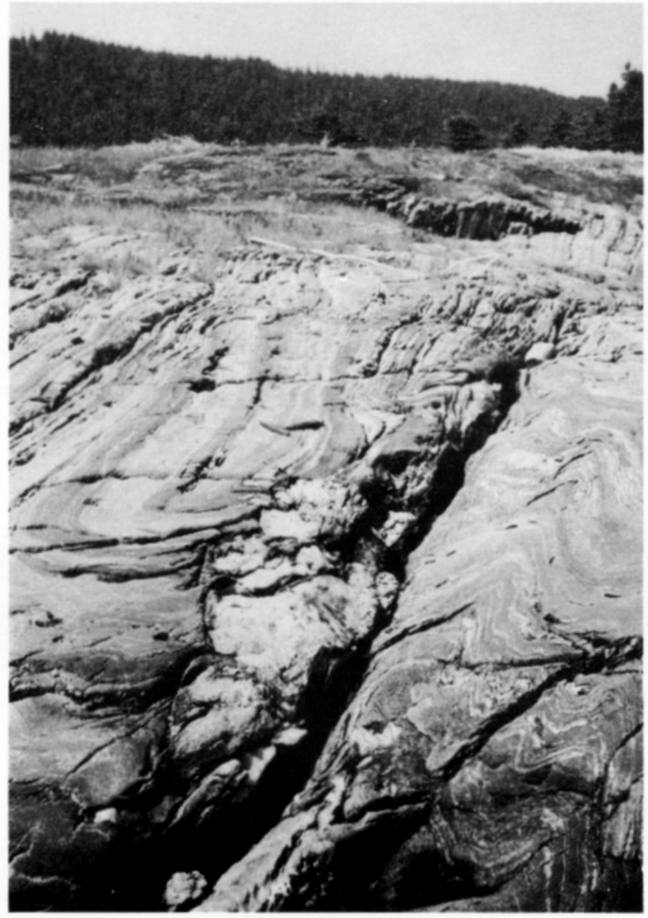


Plate 7-3: *Isoclinal F_M fold in psammite of the Pigeon Island Formation, White Bay Group in Western Arm; note F_L minor folds to right of hammer.*

Kennedy (1971) first noted megascopic, tight to isoclinal F_M folds in the Fleur de Lys Supergroup between Fleur de Lys and Coachman's Harbour. These folds are defined by the repetition of the Birchy Complex and the Rattling Brook Group. Their axes appear to vary in orientation over short distances, but overall the folds are reclined to vertical (or neutral). Bursnall (1975, 1979) identified these folds futher to the south in the Slaughter House Cove area; he noted their variable moderate to steep plunges in this area and attributed the changes to F_M folding of a previously folded surface.

The coastal synclinorium along the White Bay coast was first noted by de Wit (1972), and is defined by the repetition of the Old House Cove Group on either side of the White Bay Group; furthermore, the Garden Cove Formation and the carbonate unit within the White Bay Group are repeated in Western Arm and inland from Purbeck's Cove, though at the latter site the carbonate unit is too thin for map resolution (Figure 1-1). The axis of the synclinorium is virtually horizontal and the axial trace of the fold is within the Pigeon Island Formation. The axial plane of the fold appears to be defined by the regional west dipping S_M, and thus it is an inclined horizontal fold. Sparse stratigraphic younging indi-

Plate 7-4: *Ramsay Type III fold interference pattern involving F_M and F_L minor folds; in psammite and semipelite of the Pigeon Island Formation, White Bay Group, near Herbert Point, Western Arm.*

cators in Western Arm and along White Bay indicate that the fold is a syncline with a slightly overturned west limb.

De Wit (1972, 1974, 1980) was also the first to recognize two major anticlinoria represented by the outcrop areas of the East Pond Metamorphic Suite and an intervening synclinorium in the Gull Pond area. The traces of the anticlinoria are poorly defined, whereas the synclinorium is well defined as a north trending, nearly horizontal, tight, upright structure. The trend of these structures is parallel to S_M and the regional S_L fabric forms augen about the East Pond anticlinoria; they are thus interpreted as D_M structures. De Wit (1972, 1974, 1980) also inferred medium scale, upright, almost isoclinal F_M folds within the East Pond Metamorphic Suite, based on schistosity-layering intersections; these folds, he suggested, have amplitudes as large as 1500 m. This interpretation has not been substantiated by the present work, though it is consistent with the style of F_M folds in the area.

Major F_M structures have been reported throughout the eastern portion of the Western Orthotectonic Block (Kennedy, 1971; Kidd, 1974; Bursnall, 1975, 1979; Hibbard, 1979). In the area of Jacks Pond, Kidd (1974) interpreted the Birchy greenschists as being pinched in a tight, upright F_M synform, with the adjacent Rattling Brook semipelites to the east occupying a complementary antiform. Based on minor D_M structures in the area, Kidd (1974) indicated a gentle southerly plunge for these larger folds.

I have delineated a major fold in the Rattling Brook Group in the area of Wild Cove Brook (Figure 1-1) (Hibbard, 1979). The fold is defined primarily upon the closure of a graphitic schist unit in the area (Figure 1-1), but also on the local structural trends and the closure of amphibolite around the outermost portion of the fold nose. Its axial plane dips moderately to steeply eastward. Although I depicted it as a north plunging antiform of D_L generation on a recent map of the area (Hibbard and Gagnon, 1980), I here revert to my earlier interpretation that the structure is a south plunging F_M synform (Hibbard, 1979), mainly because it is more compatible with the regional stratigraphy and structural styles and also because it affords the simplest structural model for the data available. Its scale and style are very similar to other D_M major folds in the block whereas they are completely out of character with D_L structures (see below). Though the main penetrative fabric through the fold may locally correlate with S_L to the west, this appears to be zonal and the regional S_M is the penetrative fabric in the fold area (see above, S_M). Stratigraphic relationships between the Rattling Brook and Old House Cove Groups (see Chapter IV) suggest that the fold is a south plunging syncline.

The interpretation of the Wild Cove Brook fold as an F_M syncline is compatible with major structures outlined above for this portion of the structural block. Collectively, these structures can be used to construct a possible tectonic picture of the central portion of the block following D_M folding (Figure 7-3); slide zones are left out of this schematic diagram. Most significantly, this model indicates that the Birchy Complex is the highest structural unit in the supergroup. This is in accord with the mechanism of deformation and metamorphism in the structural block (see below).

Major Slides

Slides that formed during D_M appear to be associated with zones of major F_M folding. D_M slides have been recognized in association with the coastal synclinorium, the Wild Cove Brook fold, and the zone of folding near Slaughter House Cove, and are described below.

PENNY HILLS SLIDE ZONE

This slide zone, first recognized by de Wit (1972) in the Penny Hills area, extends from the Penny Hills south to the area inland of Little Pumbly Cove; if the slide extends further, it remains unrecognized in the area further south. The slide separates the inboard sequence from the outboard sequence of the White Bay Group on the west limb of the coastal synclinorium, but the sense of movement and significance of the fault are unknown. It is generally marked by a strong S_M fabric and a disruption of layering for up to 150 m across strike. In the area of the Penny Hills, this D_M slide combines with a D_E slide and another D_M slide in Bear Cove, thus forming a slide zone (de Wit, 1972). Near Westport and Wiseman's Coves, numerous small faults appear to splay from the Penny Hills Slide; these are marked by an intense S_L fabric and are mostly D_L faults.

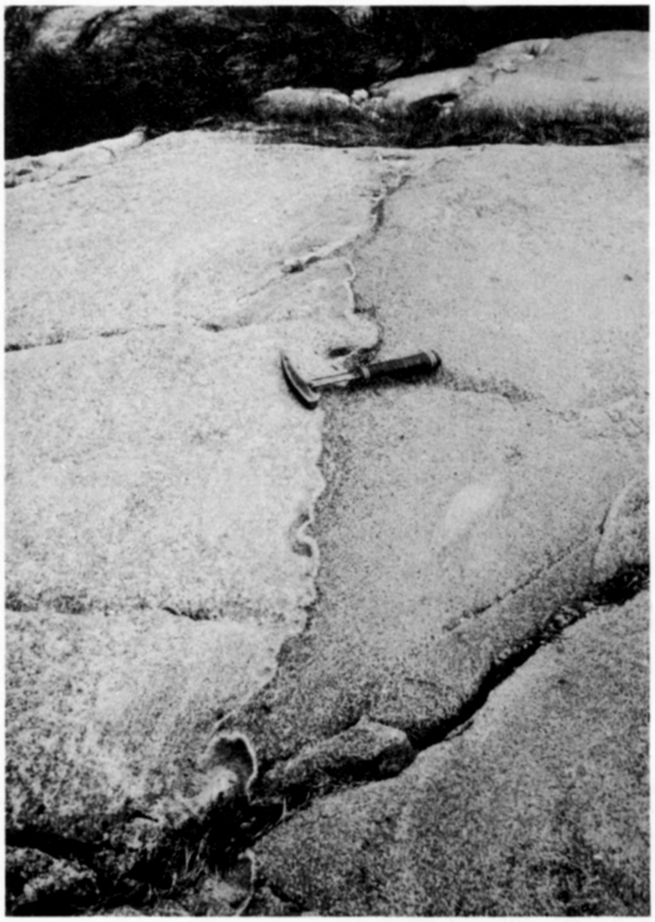


Plate 7-5: *Tectonic interface structures between psammitic layers with different mica contents in the Old House Cove Group; from the White Bay coast.*

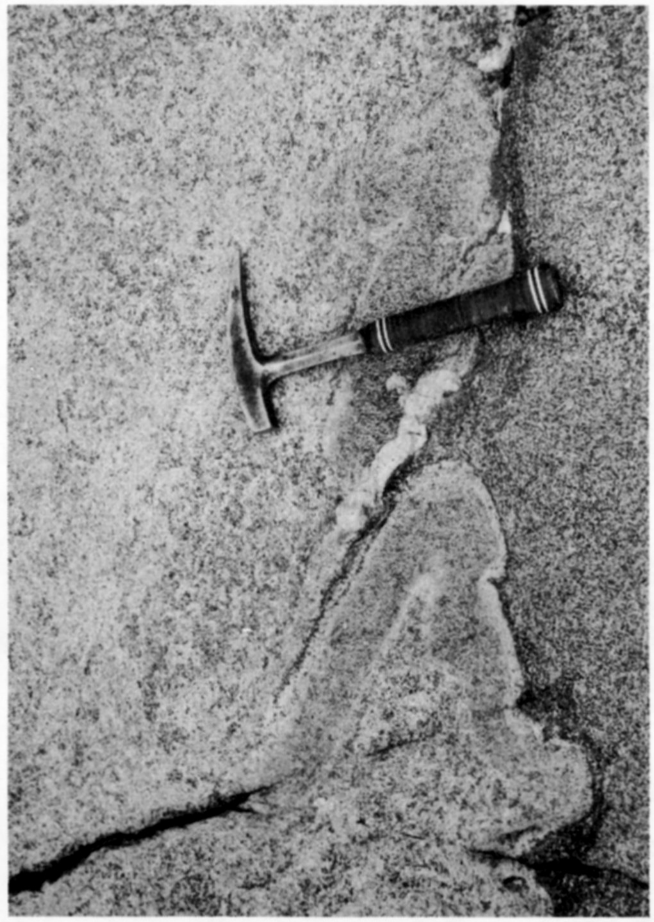


Plate 7-6: *Close-up of minor structure directly above hammer in Plate 7-5; note drag folding of layering.*

CARROL HILL SLIDE

This poorly exposed slide is largely inferred for the length of the White Bay Group south to the area of Little Chouse Brook. Where exposed, the slide is marked by an intense development of S_M and a disruption of layering for up to 100 m across strike. In addition, the zone appears to have formed a locus for later structures, as F_L folds are common in the area of this slide. The Carrol Hill Slide on the east limb of the coastal synclinorium separates the White Bay and Old House Cove Groups over a long distance, though to the north the slide appears to cut into the White Bay Group stratigraphy. This is particularly evident in the area between Purbeck's Pond and the Devonshire Valley, where the slide crosses over from the east side of the Garden Cove Formation at the pond, to the west side of the formation to the north. The sense and amount of displacement along the slide is uncertain, though it has definitely offset the Garden Cove Formation.

WILD COVE BROOK SLIDE

This slide is best defined by a strong linear feature appearing on aerial photographs in the area immediately east of Wild Cove Brook. In outcrop, the zone is marked by strongly foliated muscovite and biotite schists that are locally associated with platy, strongly schistose quartz-feldspar psammite and quartzite layers up to 2 m wide. The fabric in these rocks appears to be the regional S_M. At one locale, the mylonite-like psammites contain irregularly shaped clasts of quartz and feldspar that may be porphyroclasts. The slide appears to truncate the graphitic schist unit along the east limb of the Wild Cove Brook fold (Figure 1-1). As with other D_M slides above, its sense and amount of displacement is uncertain. A similar D_M slide has been mapped by Bursnall (1979) in the area immediately north of the Little Lobster Harbour Fault and just east of the linear zone of serpentinized ultramafics (Figure 1-1).

EASTERN SLIDE ZONE

In the area west and northeast of Flat Water Pond, de Wit noted that the layering in the Rattling Brook Group is extremely deformed and discontinuous on a centimetre to millimetre scale. He described the characteristics of the zone:

> The recognition within this belt of consistent bands or horizons is impossible and on small-scale, only slivers of representative lithologies are recognizable. The boundary in the north, from the recognizable lithological units into the slide zone, is gradational and arbitrary. Along the Westport

Plate 7-7: *Apparent transposition of semi-pelitic layer along S_M; from the Old House Cove Group along the Wild Cove Road.*

Road, a 1.5 km section reveals most of the lithological characteristics of the metasediments. The zone is not well enough exposed for detailed work, and no inland mapping along the contact was attempted. The slide zone has some characteristics resembling the tectonic schists [at the margins of the East Pond Metamorphic Suite], but the attenuatious nature of the banding and the extreme regional thinning of the metasediments suggests a large amount of the succession is missing, and so the zone is preferably classified as a slide.

BLACK BEAR SLIDE ZONE

This name was given to the tectonic break at the west side of the South Cove schist and equivalent metagabbro to the south by Bursnall (1975). The slide extends from Deep Cove south to the Marble Cove Slide. Bursnall (1975) described the zone using his structural notation (Table 7-1):

> It is marked by slides of S_2 age throughout much of its length. The western boundary of the Slaughter House metagabbro contains well-developed planar S_2 fabrics, somewhat mylonitic in places, and marks the southern extension of the Black Bear slide zone along which the metagabbro was emplaced in its present structural level during the S_2 event. Further north, abundant discrete mylonitic S_2 shear zones occur... and apparently partially truncate the western limb of a large F_2 antiform within the South Cove Formation [i.e. schist]: strongly deformed tectonic inclusions of metagabbro occur here and broadly mark the northward continuation of the S_2 elements of the slide zone.

LATE DEFORMATION - D_L

A major change in structural style is evident between D_M and D_L in the Western Orthotectonic Block, wherein D_L structures are generally less penetrative and more brittle in aspect than the D_M structures. Minor D_L features are fairly homogeneous throughout the block. D_L major structures are largely limited to the area north of the Little Lobster Harbour Fault. Here, a complex late fault system has been developed and a major late homoclinal fold has formed (Kennedy, 1971; Bursnall, 1975). The area south of the fault appears to be devoid of any D_L features that significantly contribute to the major structure of the area. Thus the Little Lobster Harbour Fault appears to be of major significance in the late structural development of the block. Bursnall (1975) related the D_L phase to a reversal of the polarity of tectonic translation from the westerly bulk movement during D_E and D_M, to easterly movement during D_L; this is particularly well shown by the steepening of structures toward Baie Verte, concomitant with an intensification of D_L structures.

Minor Structures

D_L is manifested throughout most of the belt by late fabrics (S_L) and/or late folds (F_L). Numerous distinct late sets of these structures have been distinguished by previous workers (see Table 7-1), but in the present discussion they are collectively termed D_L features. The development of these structures varies in intensity throughout the belt; D_L structures are either rare or absent in the area south of Big Chouse Brook and in the East Pond Metamorphic Suite, whereas they appear to be intense in zones near the Baie Verte Line (see D_M above), and locally along the White Bay coast in the area of Westport.

Most commonly throughout the belt, S_L is a subpenetrative crenulation cleavage that transposes S_M to varying degrees (Plate 7-8). Generally, recrystallization along the S_L fabric is weak and confined to mica growth parallel to the fabric. Locally, along the White Bay coast, as many as three S_L fabrics have been observed, though two of these appear to be of only local significance. Likewise, up to three later fabrics were reported by Kennedy (1971) and Bursnall (1975, 1979) for the area between Baie Verte and Fleur de Lys, though only one of these appears to be regionally persistent. Bursnall (1975) suggested that two of these late fabrics developed simultaneously as a conjugate couple. De Wit (1972, 1976a) reported that, in many places, S_M fabric planes were remobilized due to shearing during the D_L event; this is evident where S_L veers from its typical oblique course and merges with S_M, forming a composite fabric.

Commonly, S_L is axial planar to F_L minor folds that range from open to tight, and are generally plunging, inclined and asymmetric. Most commonly these are flexural folds of class 1C (Plates 7-3, 7-4, 7-9 and 7-10), though locally they approach similar style folds of class 2. North of the Westport road, these folds are generally moderately north plunging, whereas south of here, the plunges vary in direction and amount (Figure 1-1). In some cases, the folds display curved axes. The maximum observed amplitude of these folds is approximately 100 m along the White Bay coast. Bursnall (1975) mapped F_L folds of up to 300 m amplitude in the area north of Baie Verte. De Wit (1972) noted that F_L folds are commonly closely

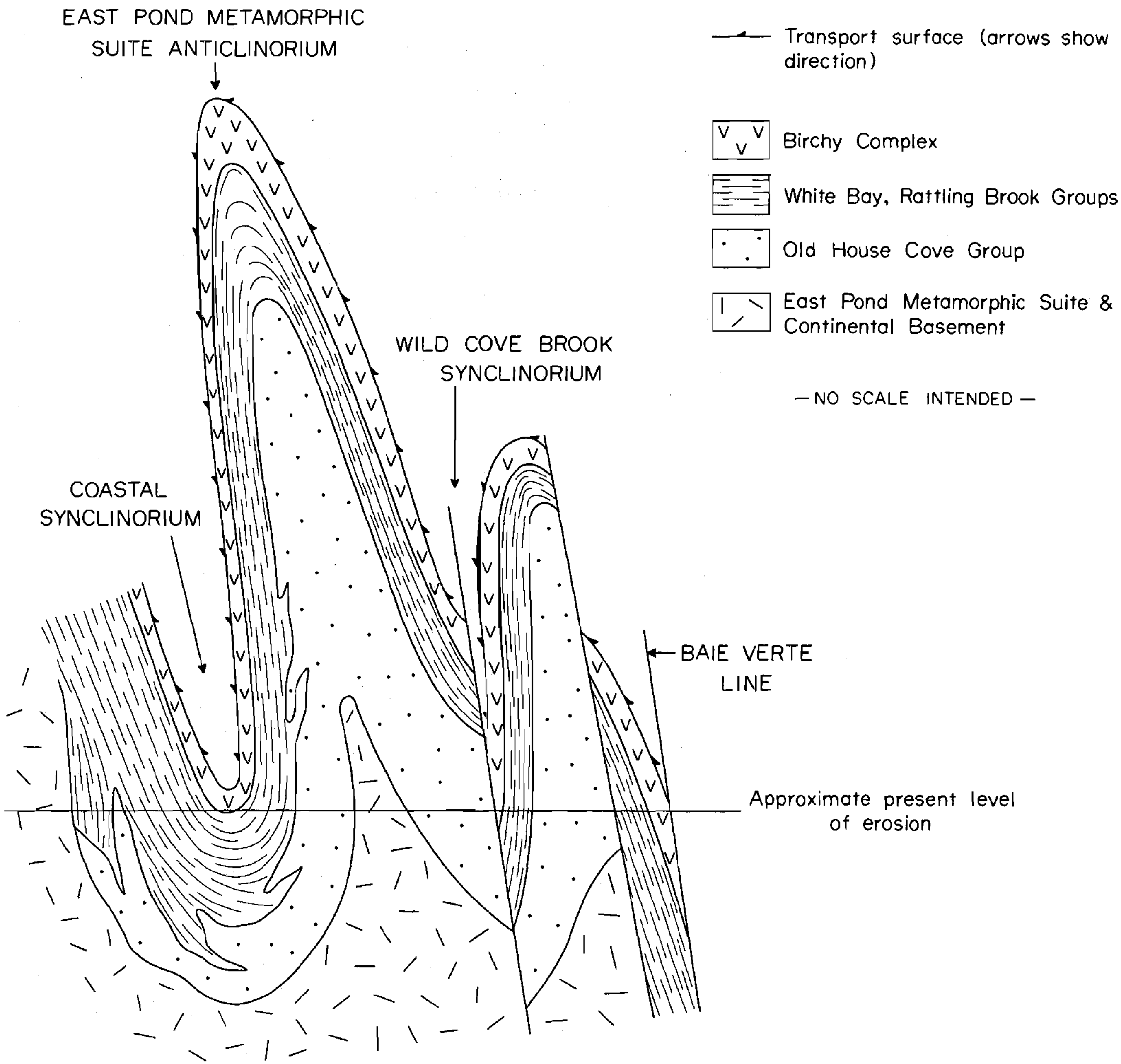


Figure 7-3: *Possible schematic section through central part of the block following* D_M.

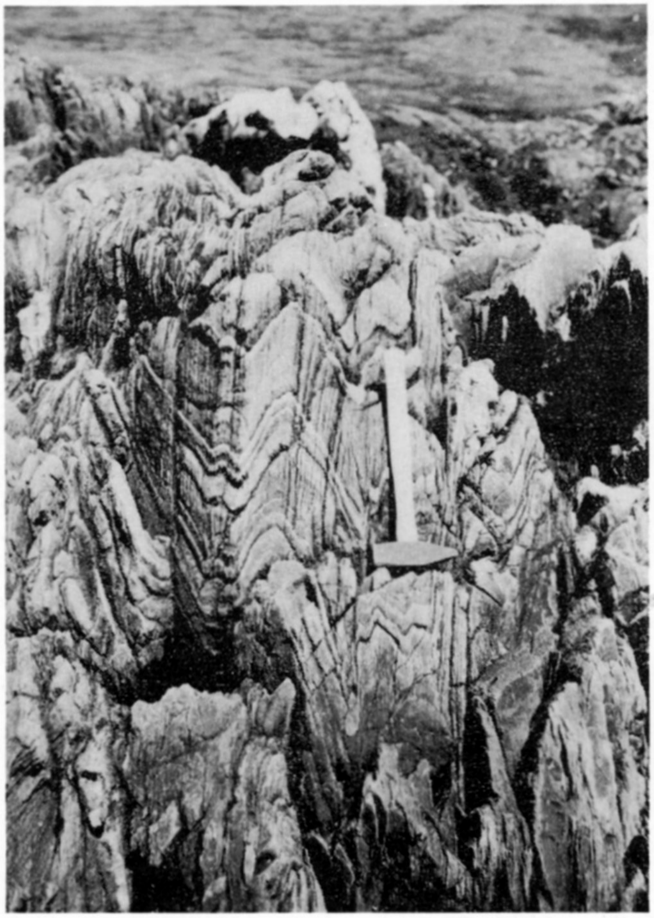


Plate 7-8: *S_L spaced crenulation cleavage within carbonate-rich semipelite of the White Bay Group on White Bay.*

Plate 7-9: *Open F_L fold within Old House Cove Group metaclastics in Southern Arm, White Bay.*

Plate 7-10: *F_L minor folds of amphibolite boudins and Old House Cove Group metaclastics in Southern Arm of White Bay.*

associated with amphibolite boudins that are of D_M generation (Plate 7-10); these folds occur near the amphibolites even in areas where F_L folds are uncommon (de Wit, 1972).

Throughout most of the area, a lineation (L_L) is commonly developed in association with D_L structures. It is manifest as an intersection lineation, mineral lineation, and clast alignment in pebbly psammites. The lineation is usually parallel to F_L fold axes.

Major Folds

Only one major F_L fold set has been noted in the terrane south of the Little Lobster Harbour Fault. Kidd (1975) recognized two south plunging synclines and an anticline in the Muddy Pond - Gull Pond area. These late folds may account for the unusual outcrop pattern in the region of Gull Pond (Figure 1-1). Other major late folds may occur in this terrane, but they are unrecognized.

North of the Little Lobster Harbour Fault, all D_E, D_M and most D_L structures were affected by a late major homocline, termed the Baie Verte Fold by Kennedy (1971). Essentially, this fold is defined by a steepening of structures toward Baie Verte (Kennedy, 1971; Bursnall, 1975, 1979).

Faults

During D_L, many pre-existing major slides were reactivated, e.g. Slaughter House and Black Bear Slides (Bursnall, 1975), the Penny Hills Slide (de Wit, 1972) and the Carrol Hill Slide. In addition, a complex fault system formed during this phase, and appears to have been best developed in the area between Baie Verte and Fleur de Lys. The system is composed of at least three sets of steep faults, each set with a distinct orientation; locally, thrust faults have developed in conjunction with these faults (Figure 1-1). The oldest set appears to be northeast trending and its faults are offset by later northwest and west trending fault sets. The relative age relationship between the latter faults is uncertain, though locally, just east of Cape Etat, and south of British Point, Hard Bay, it appears that the generally west trending Lead Mine Fault (Fuller, 1941) (Figure 1-1) truncates minor northwest trending faults. This east-west fault is described in more detail in the next chapter. The most important of these D_L faults is a northwest trending high angle fault emanating from and named after Little Lobster Harbour (Figure 1-1). On the basis of regional offset of stratigraphic units along the fault, it appears to have normal offset with the upthrown side to the south; there may also have been left-lateral strike-slip movement along the fault. To the southeast, near Advocate Mines, the fault is offset approximately 1 km southward along a steep, north trending fault; the short southeasterly extension of the fault is termed the Advocate Fault, and truncates the Baie Verte Road Fault to cross into the Paratectonic Block. This is the only major fault that extends outside of the Western Orthotectonic Block.

The Black Lake Fault (Figure 1-1), a steep fault parallel to the Little Lobster Harbour Fault, transects a portion of the structural block near Black Lake and appears to be of the same generation as the Little Lobster Harbour Fault. The Black Lake Fault displaces the western boundary of the Wild Cove Pond Igneous Suite, but appears to die out in the pluton (see also Wild Cove Pond Igneous Suite); no evidence was found for its extension across the Baie Verte Road Fault as depicted by previous workers (Neale, 1959a; Neale and Nash, 1963; Neale and Kennedy, 1967).

METAMORPHISM

The whole of the Western Orthotectonic Block underwent polyphase regional metamorphism; locally, this was overprinted by thermal metamorphic effects in the areas near postkinematic granitoids. Detailed analysis of the metamorphic evolution of portions of the block were presented by Kennedy (1969, 1971, 1975 a,b), de Wit (1972, 1976a,b, 1980) and Bursnall (1975); Kidd (1974) briefly outlined the metamorphic history of the Kidney Pond area. The results of these studies, as well as data from the present reconnaissance study serve as a basis for the following discussion. It should be noted here that any metamorphism related to D_B, the basement deformation, appears to have been completely obliterated by subsequent recrystallization during D_E, D_M and D_L.

Regional Metamorphic Mineral Assemblages and Textures

Overall, the metamorphic rocks of the block are characterized by upper greenschist to lower amphibolite facies, low to medium pressure mineral assemblages (Miyashiro, 1973). Metamorphic grade in the block generally decreases to greenschist facies toward the Baie Verte Line (Kennedy, 1969, 1971; de Wit, 1972; Bursnall, 1975). The following assemblages are common in the area:

WEST AND CENTRAL AREAS

pelitic and semipelitic schists:

quartz – plagioclase (albite or oligoclase) – muscovite – biotite ± garnet ± epidote ± chlorite

amphibolites and mafic schists:

amphibole (hornblende or actinolite) – plagioclase (albite or oligoclase) – epidote minerals ± sphene ± garnet ± quartz ± biotite ± chlorite

ultramafic rocks:

carbonate – talc – antigorite ± actinolite ± fuchsite

NEAR BAIE VERTE LINE

pelitic and semipelitic schists:

quartz – albite – muscovite – chlorite – epidote ± biotite ± garnet

mafic schists:

actinolite – albite – chlorite – epidote ± muscovite ± calcite

In addition to these common regional metamorphic assemblages, local occurrences of chloritoid (de Wit, 1972), staurolite and kyanite (Kennedy, 1969, 1971; Bursnall, 1975) have been recorded in semipelitic rocks and two occurrences of regional metamorphic andalusite in the East Pond Metamorphic Suite (de Wit, 1972) and cordierite in the Birchy Complex (Bursnall, 1975) have been noted (Figure 7-4). Accessory minerals throughout the area include tourmaline, calcite, clinozoisite, apatite, rutile, sphene, graphite and magnetite. Bursnall (1975) also reported riebeckite and stilpnomelane from coticules in the Birchy Complex.

Eclogitic assemblages of omphacite – garnet ± quartz have been reported from amphibolite pods in the East Pond Metamorphic Suite (Church, 1969; de Wit, 1972; de Wit and Strong, 1975). De Wit and Strong (1975) concluded that this assemblage resulted from a dry metamorphic environment during medium pressure greenschist to amphibolite grade metamorphic conditions. The eclogitic amphibolites were a significant factor in de Wit's (1972, 1974, 1980) interpretation of most of the present East Pond Metamorphic Suite representing a predeformed basement (see Chapter IV); but as noted before in this report, this interpretation appears invalid since the eclogitic amphibolite pods also occur in the Middle Arm metaconglomerate (Church, personal communication, 1979), which de Wit (1972, 1974) considered as cover on his predeformed basement.

Though there are only minor apparent differences in metamorphic mineral assemblages in the Fleur de Lys Supergroup and East Pond Metamorphic Suite, there is a major difference in metamorphic textures between the two units (de Wit, 1972, 1976a,b, 1980). Rocks of the supergroup are characterized by the widespread development of the porphyroblasts, particularly of plagioclase and garnet, whereas porphyroblast growth appears to have been less important in the metamorphic history of the suite; instead, the latter rocks are marked by chaotic textures that in general are only evident microscopically.

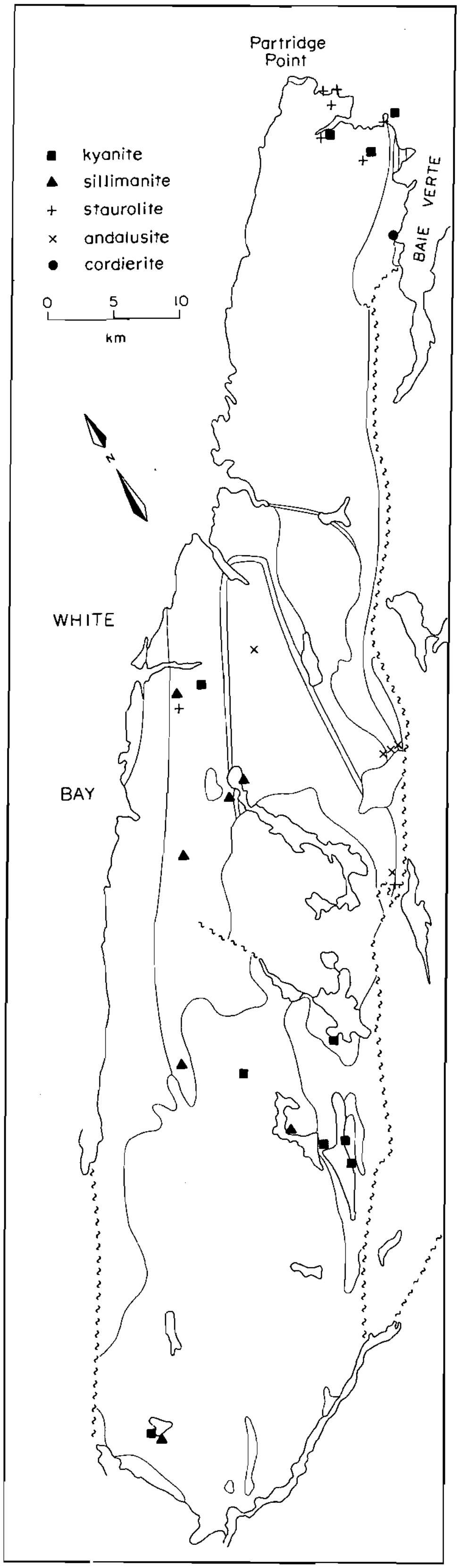


Figure 7-4: *Distribution of aluminosilicate metamorphic minerals in the Western Orthotectonic Block.*

Nearly all of the metamorphic minerals in the Fleur de Lys Supergroup occur as porphyroblasts somewhere in the belt, though plagioclase, quartz and garnet most commonly display this habit. Plagioclase is the most abundant porphyroblastic mineral, whereas garnets form the most striking porphyroblasts. In the metasediments, these porphyroblasts are typically imbedded in an oriented matrix of platy minerals (muscovite, biotite, chlorite) and quartz. In the amphibolites and mafic schists, plagioclase porphyroblasts predominate and are set in an oriented matrix of amphibole and epidote group minerals ± biotite ± chlorite. Typically, inclusion trails are preserved in the porphyroblasts. On Pigeon Island near Bear Cove, garnets form spectacular porphyroblasts up to 3 cm in diameter that display mesoscopic inclusion trails of quartz. In many places plagioclase porphyroblasts are studded with small garnets.

De Wit (1972, 1976a,b) noted that porphyroblast growth appears to be related to interaction of S_M and S_L (see below); in the present study, it was noticed that porphyroblasts are absent from areas where S_L is lacking, such as in the area of Hampden, where S_M is weak and where the regional S_L appears to become the main observable fabric (see Main Deformation above), such as the eastern edge of the block.

In contrast to the common porphyroblastic and oriented crystal textures of the supergroup, textures in rocks of the East Pond Metamorphic Suite are unusual, as described by de Wit (1980):

> The textures are the result of complex mineral intergrowth which is, at least partly, dependent on the host rock composition. The intergrowths vary from vermicular to myrmekitic, dendritic, granophyric, and to widman-statten-like: a transition from completely random symplectic intergrowth to crystallographic or lattice-controlled patterns. Commonly, minerals occur as randomly orientated, complexly intertangled poikiloblasts initiated through skeletal or dendritic growth. They are in fact coarse symplectites. Macroscopically, such textures are displayed in coarse recrystallized patches, such as in the layered amphibolites around Middle Arm Pond [East Pond]. Here a strongly layered/foliated section can be traced into swells up to several metres across, in which total random recrystallization and new growth is evident. Little trace of the original banding/foliation remains, being replaced by an ophitic-like texture.

De Wit (1972, 1980) attributed these unusual textures to a very quick release of D_M strain by the initiation of the D_L strain system in the block.

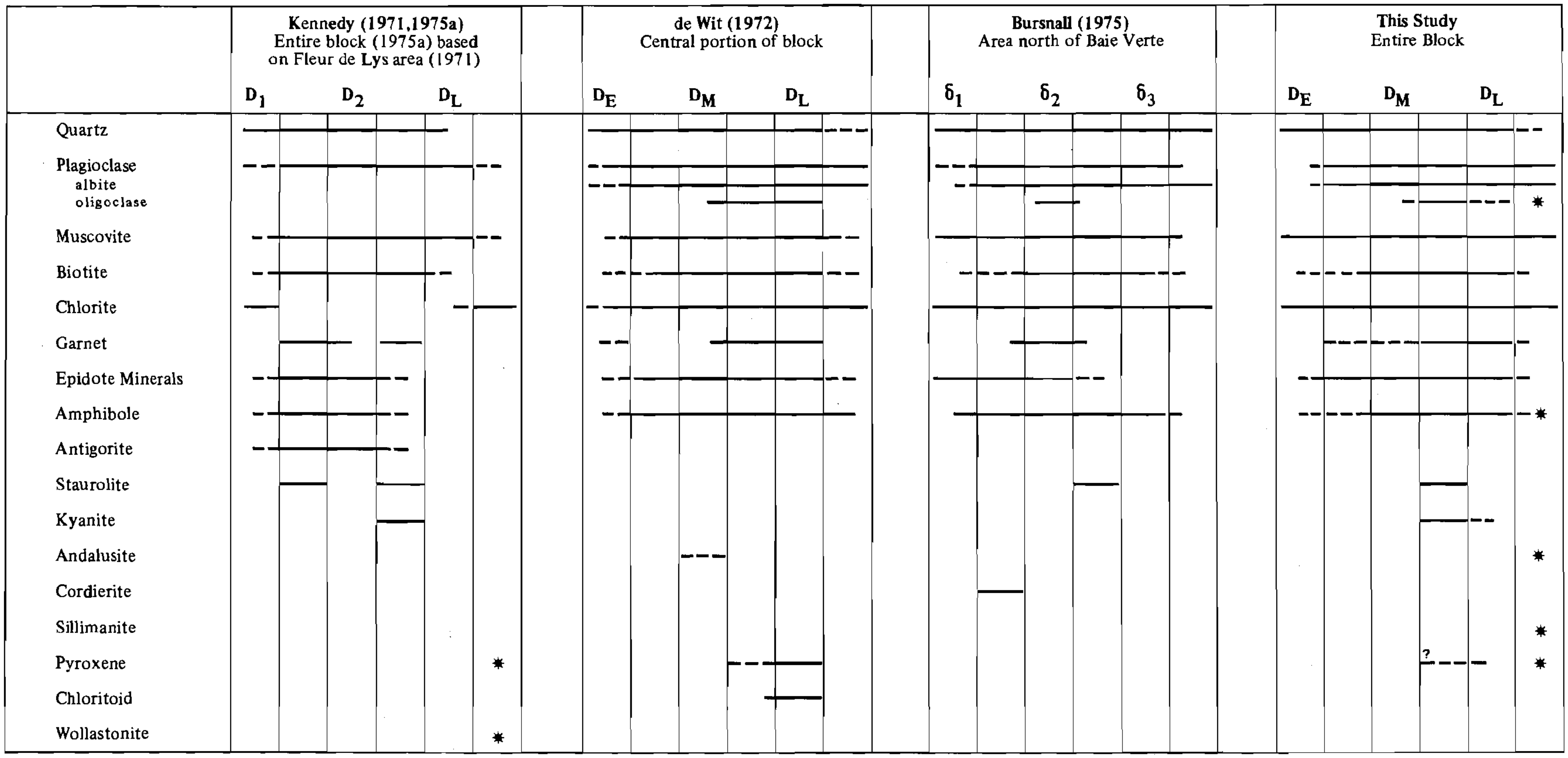
Regional Metamorphic Evolution

The prevailing metamorphic assemblages that formed either during or between any deformational phases can be determined by the relationship between metamorphic minerals and fabrics observed in thin section. In this manner, the metamorphic history of portions of the block was deduced in detail by Kennedy (1971, 1975a), de Wit (1972, 1976a,b, 1980) and Bursnall (1975); those studies are compared with the results of the present reconnaissance investigation in Table 7-2.

In the Western Orthotectonic Block, there appears to have been essentially continuous metamorphism from D_E to post-D_L, though at any given locality, all phases are not necessarily represented. The following description of the metamorphic evolution is summarized in Table 7-2.

The character of MS_E mineral growth is difficult to discern in the area due to later recrystallization events. From

Table 7-2: *Summary of detailed metamorphic analyses for the Western Orthotectonic Block.*



—— major regional growth
– – – inferred or minor regional growth
* growth due to hornfelsing

minerals preserved both between S_M planes and in inclusion trails within MP_E porphyroblasts, it appears that the MS_E metamorphism was in the greenschist facies (Table 7-2). Locally, it appears that there were areas of elevated temperature during MS_E; Kennedy (1969, 1971) and de Wit (1972) noted the growth of MS_E hornblende in a few amphibolite pods near Fleur de Lys and in the central portion of the block, respectively. The Slaughter House Slide also appears to have attained higher metamorphic grade, as Bursnall (1975) reported possible MS_E hornblende near this fault. Locally, in the central portion of the block, de Wit (1972) reported minor MS_E garnet growth.

Metamorphic conditions during MP_E appear to have been similar to those during MS_E with a few exceptions. Kennedy (1969, 1971) and Bursnall (1975) reported MP_E garnet porphyroblast growth in the area to the north and northeast of the Little Lobster Harbour Fault; he also noted that hornblende may have crystallized locally during this event. The Slaughter House Slide appears to have been an area of elevated temperature, as Bursnall (1975) reported MP_E cordierite porphyroblasts there. Kennedy (1969, 1971) reported minor MP_E staurolite from the area between Starboard Point and Cape Crapaud.

MS_M metamorphism represents the main tectonizing metamorphism in the block, that is, the most significant phase of regional metamorphism during which the minerals developed a preferred orientation due to deformation (de Wit, 1972). It is marked in this structural block by the extensive development of aligned phylloblastic minerals. Garnet porphyroblast growth continued during this event in the area north of the Little Lobster Harbour Fault (Kennedy, 1969, 1971; Bursnall, 1975) and oligoclase formed in the more westerly portions of this region (Bursnall, 1975). It appears that oligoclase also started growth in the central portion of the block, and de Wit (1972) reported MS_M andalusite from the East Pond Metamorphic Suite. Bursnall (1975) reported major MS_M plagioclase porphyroblast growth for the Slaughter House Cove region.

A major phase of porphyroblast development was initiated during MP_M over most of the block, except for the Hampden area, the region along the Baie Verte Line, and the Slaughter House Cove area. It involved the extensive growth of plagioclase (both albite and oligoclase), garnet and quartz and the sporadic development of biotite, chloritoid (de Wit, 1972), kyanite, staurolite and pyroxene. The distribution of kyanite and staurolite throughout the area is depicted in Figure 7-4. The timing of this major porphyroblastic event in relation to the deformational scheme is a moot point; some workers indicated a MP_M timing for this event (Kennedy, 1969, 1971; Bursnall, 1975), whereas de Wit (1972, 1976a,b) postulated this metamorphism to be MS_L. According to orthodox interpretations of fabric - metamorphic mineral relationships (Spry, 1969), the porphyroblast growth appears to be mainly MP_M. However, de Wit (1972, 1976a,b) noted an inconsistency in age relations of both garnet and feldspar porphyroblasts to fabrics over very small areas, and demonstrated how this inconsistency is related to D_L minor structures (de Wit, 1972, 1976a,b). In a study of garnet porphyroblasts, de Wit (1972, 1976a,b) indicated that the D_L strain regime was oriented at a high angle to that of D_M, thus causing buckling and bowing out of S_M on a small scale. De Wit speculated that during this process the porphyroblasts developed in these spaces, and with continued D_L rotational strain along S_M, complex inclusion fabrics formed in the porphyroblasts. Thus, de Wit (1972, 1976a,b) concluded that these essentially syntectonic porphyroblasts display complex inclusion trails that, if they were interpreted in the orthodox sense of fabric-mineral relationships, would indicate many generations of porphyroblast growth. He proposed that this porphyroblast growth corresponds to the formation of disequilibrium textures in the East Pond Metamorphic Suite (see above). Further detailed study is necessary to test these hypotheses and the extent to which they apply to the whole structural block; the universal application of this model for porphyroblast growth has been challenged by Schoneveld (1978).

Following this main porphyroblastic event, MP_L metamorphism is marked mainly by the growth of chlorite, muscovite and plagioclase. The plagioclase is typically a growth rim on previously formed plagioclase porphyroblasts. The phylloblastic minerals indicate retrogressive metamorphism. Late phase, prehnite-pumpellyite-bearing veins were noted locally by Kidd (1974) and Bursnall (1975).

Discussion and Summary of Regional Metamorphism

On a broad scale, the metamorphic evolution of this block appears fairly homogeneous, with only local variations. The regional metamorphic growth through MS_M appears to have been characterized by greenschist assemblages, though local "hot spots" are evident. In the area immediately northeast of the Little Lobster Harbour Fault, MP_E upper greenschist assemblages with garnet appear to mark the local metamorphic peak. Greenschist metamorphism also appears to have prevailed southward in the Birchy Complex, along the Baie Verte Line. The Slaughter House Slide marks an area of higher grade metamorphism in this greenschist grade terrane; MP_E hornblende and cordierite in the slide area suggest low pressure amphibolite facies metamorphism (Miyashiro, 1973). Further to the north and west, medium pressure lower amphibolite facies prevailed during MP_E as indicated by the presence of MP_E hornblende and staurolite (Kennedy, 1969, 1971). Peak metamorphism continued through to MP_M as indicated by MS_M oligoclase and MP_M kyanite and staurolite near Fleur de Lys (Kennedy, 1969, 1971; Bursnall, 1975). This peak metamorphism appears to have coincided with that in the central and southern portions of the block, where it is marked by MP_M kyanite, staurolite, oligoclase and pyroxene (in the East Pond Metamorphic Suite). Eclogitic assemblages in the suite appear to have formed concomitant with medium pressure amphibolite grade metamorphism in the Fleur de Lys Supergroup (de Wit, 1972). Near Hampden, it appears that greenschist was the highest grade achieved. Following peak metamorphism in the block, greenschist metamorphic conditions prevailed.

In summary, peak metamorphism varied from medium pressure, lower amphibolite facies metamorphism in the west and central portions of the block to the upper greenschist facies near Hampden and along the Baie Verte Line. In the area north of the Little Lobster Harbour Fault, it appears that peak metamorphism started earlier, during MP_E, than the MP_M initiation of peak metamorphism to the south.

Contact Metamorphism

Thermal aureoles formed in the areas near the postkinematic granitoids of the block. Generally, hornfelsing related to these plutons is limited to the area less than 1 km from the intrusive bodies. Around the perimeter of the Wild Cove Pond Igneous Suite, the following metamorphic minerals have been observed: oligoclase, microcline, andalusite and sillimanite in metasediments, biotite and hornblende in mafic rocks and diopside in marbles. Kidd (1974) reported contact metamorphic andalusite locally replaced by fibrolite from the area west of Kidney Pond and also east of Wild Cove Pond. At the latter locality, he reported, the andalusite replaces MP_M staurolite. Sillimanite appears to be the most common mineral in the pelitic rocks of the East Pond Metamorphic Suite and Old House Cove Group along the northern and western borders of the Wild Cove Pond Igneous Suite. Oligoclase is fairly common in these rocks and microcline is rarely developed. Diopside was observed only in the marbles at the perimeter of the pluton south of Big Chouse Brook.

In the area immediately surrounding the Partridge Point Granite, Kennedy (1969, 1971) reported extensive recrystallization of biotite and diopside-vesuvianite-wollastonite assemblages in calcareous rocks abutting the stock.

AGE AND MECHANISM OF DEFORMATION AND METAMORPHISM

The age and mechanism of deformation and metamorphism of the area here termed the Western Orthotectonic Block have been controversial issues on the Baie Verte Peninsula. Early tectonic analyses indicated a pre-Early Ordovician age of tectonism for the area based on the observations of Church (1969). He correlated rocks in the present Eastern Orthotectonic Block with those of the Fleur de Lys Supergroup and found deformed detritus from his "easterly Fleur de Lys" in the Arenigian Snooks Arm Group and the former Baie Verte Group. This concept of a pre-Early Ordovician deformation led to models for the area that were incompatible with regional relationships (e.g. Dewey and Bird, 1971; Kennedy, 1973, 1975a; Kidd, 1974, 1977). Most glaringly, the models called for a pre-Early Ordovician age for the Cape St. John Group; however, the group unconformably overlies the Arenigian Snooks Arm Group. This made many other workers in the area suspicious of such a concept.

De Wit (1972) questioned the validity of this age assignment for the Fleur de Lys tectonism, and noted that the tectonism could be as young as Middle Ordovician, based on the correlation of Fleur de Lys stratigraphy and some structures with those of the less deformed portions of the Humber Zone. He also recognized a predeformed basement in the terrane and assigned it a Grenvillian age (de Wit, 1972, 1974, 1980); this age is accepted here, but on a more limited range of rocks than that envisaged by de Wit (Chapter IV).

The polyphase deformation and metamorphism of the main Fleur de Lys outcrop belt has more recently been related to the ophiolitic rocks along the Baie Verte Line (Bursnall, 1975; Bursnall and de Wit, 1975; Kennedy, 1975a; Williams et al., 1977; Hibbard, 1982). The Fleur de Lys Supergroup and East Pond Metamorphic Suite collectively represent sediments and continental basement originally at the eastern edge of ancient North America (Dewey, 1969b; de Wit, 1972, 1974, 1980; Williams and Stevens, 1974; see Chapter IV). Ophiolitic complexes of the Baie Verte Belt, including those along the Baie Verte Line, have been related to a single cycle of Eocambrian to Early Ordovician ocean floor generation (Church and Stevens, 1971; Bursnall and de Wit, 1975; Williams et al., 1977). The earliest deformation and metamorphism of the Fleur de Lys Supergroup have been attributed to late Early to Middle Ordovician obduction of this oceanic lithosphere over the clastic rocks at the margin of ancient North America (Bursnall, 1975; Bursnall and de Wit, 1975; Williams et al., 1977). Originally, the oceanic crust was rooted to ophiolitic complexes along the Baie Verte Line, but is now preserved as klippen at Hare Bay and the Bay of Islands.

Bursnall (1975) related D_E and D_M of the present report to the emplacement of ophiolites to the west. Particularly compelling for this interpretation is the presence of pre-D_E ophiolitic mélanges (Bursnall, 1975; Williams et al., 1977; Williams, 1977a) and the emplacement of ultramafic rocks along D_E slide zones that locally appear to have had elevated temperatures (Slaughter House Slide, Bursnall, 1975; see above). If the relation of the structures in the block to the western Newfoundland allochthonous ophiolites, as outlined above, is accepted, then Early Ordovician age dates from dynamothermal aureoles at the base of the western ophiolites most likely mark the inception of deformation in the Western Orthotectonic Block. These $^{40}Ar/^{39}Ar$ dates of 490 $\pm$ 5 Ma and 469 $\pm$ 5 Ma (Dallmeyer, 1977) denote either a Llanvirn to Late Caradoc (Ross et al., 1978) or a Tremadoc to Late Arenig (Churkin et al., 1977) age for D_E of the block.

De Wit (1972, 1976a,b, 1980) related the D_L deformation of the block to stresses that released the intense D_M stress system, and Bursnall (1975) related these late structures to a reversal in tectonic polarity from westerly to easterly. Essentially, D_L appears to be related to uplift following obduction. The MP_M timing of the peak metamorphism throughout most of the belt appears to correspond very reasonably with this interpretation, as one would expect maximum depth and temperature to be attained immediately prior to uplift. Lower grade metamorphism of rocks during MP_E and MS_M may indicate that rocks involved in these events were situated high in the structural pile. Likewise, low grade rocks along the Baie Verte Line suggest that rocks in this position remained high in the structural pile throughout all metamorphic events. This notion appears to be confirmed by a decrease in intensity of D_M toward the line and by the tectonics of the Transition Blocks (see below).

Incremental $^{40}Ar/^{39}Ar$ ages of the main Fleur de Lys outcrop area indicate that regional metamorphic temperatures had dropped below approximately 300°C by the Middle Silurian (Dallmeyer, 1977a). These dates (Figure 1-1) mark the cooling ages for hornblende and biotite and most likely reflect the subsidence of either MP_M or MS_L metamorphism in the area. North of the Little Lobster Harbour Fault, $^{40}Ar/^{39}Ar$ ages appear to be slightly younger and may reflect later uplift of this block along the fault. Alternatively, if left-lateral movement occurred along the Little Lobster Harbour Fault, then the northerly part of the Western Orthotectonic Block may have originally lain further to the east and may have been influenced by Acadian tectonism in that area (see Eastern Orthotectonic Block). However, polytectonism in most of the Western Orthotectonic Block can be broadly ascribed to Taconic events.

II. TRANSITION BLOCKS

These are three distinct blocks that appear to have a common structural history and are interconnected along the Baie Verte Road Fault system (Figure 7-1). The northern Transition Block is bounded to the north by the Marble Cove Slide, which represents a portion of the Baie Verte Line (see below), and to the south by the probable unconformity of the Advocate cover sequence on the Advocate ophiolite (see Chapter V), as well as by southeastward directed thrust faults (Bursnall, 1975) (Figure 1-1). The block, here, is composed of two steeply dipping, imbricate ophiolite slices of the Advocate Complex, separated by a major slide zone, the Advocate Western Boundary Slide (Bursnall, 1975). The southerly two blocks are composed mainly of ultramafic rocks situated along the southern extent of the Baie Verte Line and are separated from surrounding structural blocks by the Baie Verte Road Fault system.

The informal name of these blocks derives from their position relative to the Western Orthotectonic and the Paratectonic Blocks; within the blocks, there appears to be a transition from dominantly Taconic structures of the Western Orthotectonic Block (see above) to Acadian structures of the Paratectonic Block (see below). The tectonic history of the Transition Blocks is extremely complex and heterogeneous; that of the northern block has been documented in detail by Bursnall (1975). Structures in the southern ultramafic blocks are poorly developed and have been largely neglected in previous detailed descriptions. Scant available descriptions (de Wit, 1972; Kidd, 1974) indicate that the structures developed in these blocks are similar to those in the northern block; hence, the following discussion focuses on Bursnall's (1975) work to the north. Structures throughout the blocks are critical in structural correlation across the Baie Verte Line (see Baie Verte Line).

STRUCTURE

Bursnall (1975) recognized five major phases of deformation in the northern block. In the present study, these are condensed into three broad phases, D_E, D_M, and D_L. D_E and D_M are roughly correlative with S_1 and S_2, respectively, of Bursnall (1975), whereas his later phases are collectively referred to as D_L in this report. In the block, D_M structures generally appear to decrease in intensity toward the southeast concomitant with an intensification of D_L structures; the Marble Cove sequence is generally more highly tectonized than rocks to the southeast, though both areas contain structures attributable to the three deformation phases.

Fabrics have been developed during all of the deformation phases in the block, but due to their heterogeneous development, temporal correlation of individual fabrics throughout the block is confusing and unreliable. For example, Bursnall (1975) noted that locally in the cover rocks to the Marble Cove sequence (see Chapter V), "...S_2 [S_M] may be easily confused with a coarser, less penetrative later crenulation of probable S_3 [D_L] age..." In addition, the recognition and distinction of S_E and S_M in gabbroic portions of the block is clouded by the local presence of a predeformational nonpenetrative "high temperature" foliation (see Chapter IV). Thus, any structural succession erected for the area is tentative at best.

The earliest schistosity, S_E, is uncommon and appears as a relict fabric between S_M foliation planes and in F_M fold noses; southeast of the Marble Cove sequence, an early fabric that may be temporally equivalent to S_E of the sequence is ill defined and observable only in thin section. The succeeding foliation, S_M, varies from a regional, subpenetrative, phacoidal schistosity to a penetrative schistosity in local mylonitic zones. S_M appears to decrease in intensity toward the southeastern portion of the block. Bursnall (1975) inferred that in places throughout the block, the main observable fabric may be temporally equivalent to one of the S_L fabrics in other parts of the block (see below); thus, in these places, S_M may either represent a fabric completely separate from that elsewhere in the block, or record strong D_L remobilization of pre-existing S_M planes. Minor F_M folds are apparently uncommon in this block, although, on the basis of the variation of F_L fold axes in the Marble Cove sequence, Bursnall (1975) postulated the presence of large scale F_M folds.

D_L structures of this study embrace at least three phases of structures that, in general, are zonally developed. The fabrics associated with these phases range from a fracture cleavage, through a spaced crenulation cleavage, to a subpenetrative mylonitic fabric. Generally, a S_L crenulation cleavage is the most pronounced and widespread of these fabrics in the northern Transition Block. It appears to be most intensely developed to the southeast, though Bursnall (1975) noted that D_L structures in the area of the Marble Cove Slide are exceedingly complex. The zonal interplay of these fabrics with other fabics in the block causes confusion in temporal correlations of foliations in the block, such as outlined above for S_M. F_L minor folds are generally open to tight class 1 folds. Larger scale F_L folds, up to 400 m amplitude, appear to be present in the block and are best defined along the contacts of the major rock divisions (Figure 1-1); these larger folds are also prevalent in the area of the Marble Cove Slide.

Numerous faults and shear zones dissect rocks in the northern Transition Block; most of these structures are related to D_L and are of minor regional significance. However, there are zones of major D_M movement and a late D_L regional thrusting event. Large portions of the Marble Cove cover sequence are protomylonitic, and Bursnall (1975) related the protomylonite fabric to D_M. Similarly, he interpreted the major fault that bisects the block, the Advocate Western Boundary Slide, to be as old as D_M if not older. The slide zone, which separates the Marble Cove sequence from other Advocate Complex rocks to the south, is largely composed of exceedingly fine grained, black to dark green, mafic cataclastic schists that contain tectonic lenses of highly altered gabbro, serpentinite, mafic volcanic rock, and quartz albitite; in addition, these schists appear to grade locally into hornblende amphibolites and a zone of garnet and pyroxene amphibolites near the Advocate ultramafic body (see Chapter V). Essentially, the zone represents a tectonic mélange. Bursnall (1975) described the overall structural character of the zone:

> An approximately 100 m wide zone of cataclastites, protocataclastites, recrystallized mylonites, and other highly deformed schistose rocks outcrops within the central part of the western boundary zone.... The zone is truncated to the northeast and south by later faulting in the Advocate Mine area but reappears south of the Advocate Fault. The earliest structural elements within the zone are at least as old as late-D_P [i.e. D_M] but have been considerably modified by subsequent D_S [i.e. D_L] events.

Composite planar tectonic fabrics are common and tectonic slicing and interleaving of lithological units on all scales into relatively thin but tenuously persistent lenticular areas is common. Late-D_S [i.e. D_L] high angle reverse faulting and thrusting from the northwest has partially obliterated the earlier tectonic relationships....

Bursnall (1975) indicated that the higher grade amphibolites were transported into their present position during early tectonic movements along the slide zone, and that the fabric in these rocks is most likely correlative with the regional S_M. He interpreted the Western Boundary Slide as marking the early emplacement of ophiolitic rocks of the Advocate area onto ophiolitic rocks of the Marble Cove sequence (Bursnall, 1975). This slide subsequently acted as a major zone of D_L strain. This interpretation of the slide is significant in the correlation of structural events between this block and those of the Western Orthotectonic Block (see below).

The most significant D_L faulting appears to be the southeastward thrusts that occurred along the Western Boundary Slide and along the south margin of the block. Displacement appears to have been significant in some cases, as Bursnall (1975) interpreted the large gabbro bodies just south of the block and west of Shark Point as detached, southerly thrusted portions of the Transition Blocks.

METAMORPHISM

The whole of the northern Transition Block appears to have been metamorphosed in the lower greenschist facies, with the exception of garnet-amphibolite ± pyroxene assemblages along the Advocate Western Boundary Slide and local areas in the Marble Cove sequence. Bursnall (1975) reported the following assemblages from the block:

metagabbros:

zoisite/epidote + albite + quartz ± chlorite ± actinolite/tremolite

mafic schists:

actinolite + chlorite + epidote + albite + quartz ± sphene ± muscovite

The growth of these minerals as related to deformational phases in the northern part of the block is summarized in Table 7-3 from Bursnall (1975). Locally in the Marble Cove sequence, Bursnall (1975) noted hornblende-like cores to actinolitic amphiboles. He tentatively included this higher grade metamorphism with early MP_E growth of garnet and pyroxene along the Western Boundary Slide (Table 7-3). Subsequent metamorphism is lower greenschist and subgreenschist grade.

AGE AND MECHANISM OF DEFORMATION AND METAMORPHISM

The Transition Blocks are a key area in the tectonic framework of the Baie Verte Peninsula, because the transition from dominantly Taconic structures in the Western Orthotectonic Block to the Acadian structures of the Paratectonic Block (see below) must occur either within or at the margins of these blocks.

Bursnall (1975) correlated D_E and D_M structures in both the Western Orthotectonic and northern Transition Blocks and considered them to be the result of Early to Middle Ordovician obduction of ophiolites in the area (see Western Orthotectonic Block). Further compelling evidence for this interpretation is found in the northern Transition Block, where garnet and pyroxene amphibolites are found at the base of ophiolitic rocks along the Western Boundary Slide; these rocks are very similar to those forming the dynamothermal aureoles at the base of transported ophiolites to the west, in the Humber Zone (Williams and Smyth, 1973), and are probably of similar origin. The decrease in intensity of D_E and D_M structures to the southeast most likely reflects the higher structural position of the southeasterly rocks during these events.

The significance and origin of D_L fabrics in the block is less clear than that of the earlier fabrics. Bursnall (1975) tentatively correlated early D_L structures with both D_L structures of the Western Orthotectonic Block and D_M structures of the Paratectonic Block (see below). This appears to be inconsistent with regional data since it infers a correlation of D_L of the Western Orthotectonic Block with D_M of the Paratectonic Block; this is unlikely because the late fabric of the orthotectonic block is probably about 420 to 400 Ma old (see Western Orthotectonic Block), whereas deposition of the Micmac Lake Group, which is affected by S_M of the Paratectonic Block, was in Late Silurian to Early Devonian times (see Chapter V). In addition, Kidd (1974) indicated that early D_L structures of the Orthotectonic Block are truncated by the Baie Verte Road Fault system, which formed during D_M of the Paratectonic Block. There are at least three solutions to this apparent problem:

(i) the fabrics in question are not correlative, but are apparently inseparable within the Transition Blocks due to either merging of the fabrics (coaxial deformation) or their physical similarity;

(ii) early D_L fabrics within the Western Orthotectonic Block are not temporally equivalent throughout the area;

(iii) there is more than one age of fabric development in the Paratectonic Block.

Considering the nature of the Transition block as outlined above, I prefer solution (i), though clearly more detailed work is necessary before this problem can be resolved.

III. EASTERN ORTHOTECTONIC BLOCK

This orthotectonic block encompasses the northern part of the Cape St. John Peninsula (Figure 7-1) and is bounded to the north by the ocean and to the northwest by the Scrape Thrust; to the south, it is structurally gradational with the Paratectonic Block. All of the rocks in this area are polydeformed and polymetamorphosed. The Baie Verte Line (see below) strikes eastward through the block; hence the block straddles both the Fleur de Lys and Baie Verte stratigraphic belts.

The basement to most of the block is ophiolitic; that of the Ming's Bight Group is uncertain. The Pacquet Harbour Group is herein interpreted as ophiolitic (see Chapter V and VI); the Cape St. John Group unconformably overlies the Betts Cove Ophiolite, and ophiolitic blocks are common in the Cape Brulé porphyry. Also, gravity studies (Miller and Deutsch, 1976) indicate a dense subsurface body, most likely ophiolitic, beneath the southern part of this block. The eastward exten-

Table 7-3: *Relationship of metamorphic mineral growth and structural sequence for the Transition Blocks (from Bursnall, 1975).*

METAMORPHIC MINERAL	D_E		D_M		D_L	
Chlorite	———	———	———	———	– –	
Muscovite	———	———	———	———	– –	
Actinolite/Tremolite	———	———	—— –	– – –	– –	
Dark green amphibole	———	——				
Albite	———	———	———	– – –	– –	
Quartz	———	———	———	———	– – —	—
Zoisite/Epidote	———	———	——			
Calcite			–	– – –	———	—
Stilpnomelane			– –	– – –	– –	
Pumpellyite				– – –	– –	
Prehnite		– – –	– – –	– – –	– –	
Garnet	—	———				
Pyroxene	—	———				

Observed growth ———

Inferred growth – – – –

sion of the Baie Verte Line (see below) severed the Ming's Bight Group from the remainder of the block, and basement to the group is unexposed.

Many workers have undertaken structural and metamorphic analyses of this block and their work is summarized in Tables 7-4 and 7-5. There is by no means a consensus on the structure of the area (see Table 7-4), but the tectonic history can be broadly viewed, at this time, as three tectonic pulses, namely D_E, D_M and D_L. Previous workers have indicated that all of the rock units in the block underwent the same polydeformational history (Church, 1965a; Kennedy et al., 1973; DeGrace et al., 1976). I have undertaken only minor reconnaissance studies in the northwest and northeast portions of the Pacquet Harbour Group; hence, the following descriptions are largely derived from previous workers.

It is my impression, from compiling the previous analyses, that D_M and D_L structures are commonly confused in northern portions of the block; also from previous work, personal observations, and observations of others (T. Calon, P. Williams, personal communications, 1978), it appears that structures in the Pacquet Harbour area are more complex than elsewhere in the block. Thus, the following review is intended to serve only as a broad, tentative overview of the block; clearly, there is room for further structural analysis.

EARLY DEFORMATION - D_E

The earliest deformation phase in the block is manifested as a relict fabric preserved either between S_M foliation planes or in the noses of F_M minor folds. I have also observed S_E within a mafic xenolith in the Cape Brulé porphyry (Plate 7-11); it was crenulated by S_M in the porphyry. This S_E fabric has been interpreted by some workers as a relict bedding plane foliation (Church, 1969; DeGrace et al., 1976), whereas others have considered it a tectonic fabric because of its presence in pillow lava and massive lavas (Kennedy, 1975a; Tuach, 1976; Tuach and Kennedy, 1978). I consider S_E as a tectonic fabric in the Pacquet Harbour area because here it appears to be associated with slide zones, as outlined below. Neither Coates (1970) nor Gale (1971) recognized the relict D_E fabric (Table 7-4).

Table 7-4: *Tentative structural correlation chart for the Eastern Orthotectonic Block.*

This study Entire Block	Church (1969) Entire Block	Coates (1970) Area Southeast of Pacquet Harbour	Gale (1971) Rambler Mines Area	Kennedy (1975a) Entire Block	DeGrace et al. (1976) Entire Block	Tuach (1976) Rambler Mines Area
	F_6 - Close to open conjugate folds				D_6 - Large open F_6 warps	
	F_5 - Open folds				D_5 - Open to close F_5 recumbent folds; local S_5 fracture cleavage	
D_L	F_4 - Open to close folds	D_3 - Local crenulations and open concentric folds	D_3 - S_3 crenulation cleavage, recumbent F_3 minor folds	D_4 - S_4 strain-slip fabric, open F_4 minor folds	D_4 - Broad, open F_4 folds	D_4 - S_4 strain-slip fabric, open upright F_4 minor folds
	F_3 - Crenulation cleavage, tight to close folds	D_2 - Subpenetrative transpositional fabric; open to tight, upright to recumbent F_2 folds; major F_2 recumbent folds	D_2 - Local vertical crenulation cleavage and F_2 minor folds	D_3 - S_3 strain-slip fabric; open to isoclinal F_3 folds overturned to the south	D_3 - S_3 strain-slip fabric, recumbent south facing F_3 folds	D_3 - S_3 strain-slip fabric, minor northeast plunging folds
D_M	F_2 - Crenulation and axial plane transposition and foliation, tight to isoclinal folds	D_1 - Penetrative L-S fabric with pronounced L component and local F_1 minor folds	D_1 - L >S penetrative fabric	D_2 - S_2 schistosity of L-S fabric, locally a crenulation fabric; major south facing F_2 isoclinal folds, minor close to isoclinal F_2 folds	D_2 - S_2 schistosity, major and minor tight, upright F_2 folds	D_2 - Penetrative L-S fabric to strain-slip fabric, major tight upright F_2 folds; prominent L_2 lineation
D_E	F_1 - Bedding plane foliation, tight to isoclinal folds			D_1 - S_1 schistosity or L-S fabric, local major F_1 folds, minor tight to isoclinal F_1 folds; tectonic slides	D_1 - S_1 bedding plane schistosity (?)	D_1 - Relict L-S fabric

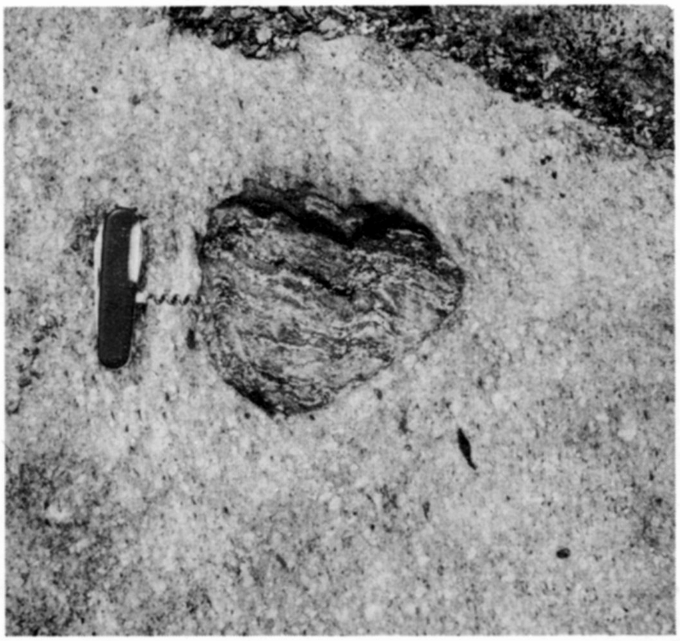


Plate 7-11: *S_E within a layered amphibolite xenolith in the Cape Brulé porphyry; note that S_M trends approximately parallel to pen-knife and S_E parallels corkscrew blade. Along the La Scie highway.*

Kennedy (1975a) has been the only worker to recognize F_E folds. He recorded tight to isoclinal F_E minor folds and suggested that downward facing F_E major folds are present between Ming's Bight and Pacquet Harbour.

Along the southern margin of the Ming's Bight Group, ophiolitic mélanges, metaclastics and greenschists containing ultramafic fragments appear to mark an early zone of D_E or pre-D_E ophiolite emplacement (see Ming's Bight Group). Formation of the mélanges and incorporation of clasts in the greenschists predate D_M, but their relationship with D_E is uncertain at this time. At Pacquet Harbour, these rocks appear to be associated with possible D_E slide zones. Along the brook immediately west of Pacquet Harbour, locally known as Big Brook, platy quartzitic rocks reminiscent of a slide zone blastomylonite (Plate 7-12) occur along strike from and in close association with the rocks bearing ophiolite detritus. The quartzitic rocks locally contain a strong early fabric, transposed by the local main fabric. Because the zone served as a locus for intense D_M and D_L strain, relationships beween fabrics are complex and the early fabric in the quartzitic rocks can only tentatively be identified as S_E. This zone is also distinct in that it served as a locus for later higher grade metamorphism (see Metamorphism, below). Thus, I interpret a major D_E fault zone (see Contact Relationships, Ming's Bight Group, Chapter IV) marking the emplacement of ophiolitic slivers; I informally term the section along Big Brook as the Big Brook slide zone. This interpretation is strengthened by both the position of this zone at the boundary of the Fleur de Lys and Baie Verte Belts and the correlation of this zone with identical zones in the Birchy Complex (see Ming's Bight Group, Chapter IV, and Western Orthotectonic Block).

Plate 7-12: *Band of platy quartz-kyanite crosscutting coarse biotite schist of Ming's Bight Group on Big Brook, Pacquet Harbour. Highly strained aspect of these rocks strongly suggests that this represents a slide zone.*

MAIN DEFORMATION - D_M

D_M is characterized by a strong L-S fabric throughout the area, which ranges from a S>L fabric to an almost pure L fabric. The S_M fabric ranges from penetrative to subpenetrative and it is a crenulation cleavage where S_E is present. S_M is present but inhomogeneously developed in the larger intrusive bodies of the structural block. The L>S fabric is common in the Pacquet Harbour Group and is featured as an intense mineral and clast lineation; it is particularly striking in outcrops along the La Scie highway (Plates 7-13, 7-14, 5-12). Coates (1970) reported intense L_M fabrics in Grand Cove, and J.R. DeGrace (personal notes, 1975) noted a strong L_M hornblende lineation along the coast east of Brents Cove (Plate 7-15). The ore bodies at Consolidated Rambler Mines are parallel to the L_M lineation. Kennedy (1975a) noted that D_M structures decrease in intensity southward through the structural block. I am uncertain if the intensity of D_M decreases to the south, but clearly metamorphism along the S_M fabric is of lower grade in this direction.

F_M minor folds are uncommon in the area, though all workers have noted that they are generally upright, gently

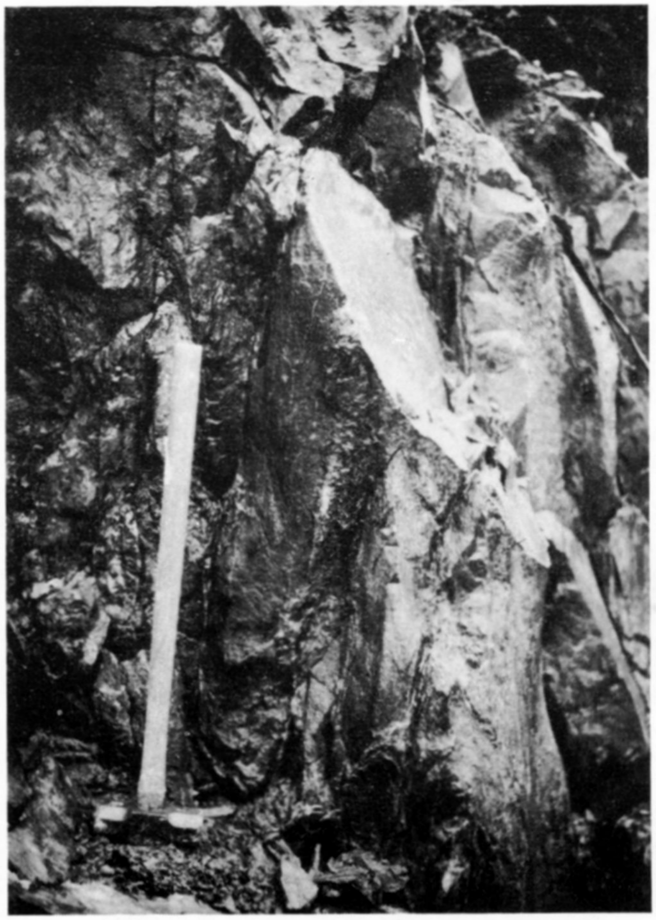


Plate 7-13: *Strong L_M elongation parallel to hammer handle in pillow lava of the Pacquet Harbour Group on the La Scie highway, just east of Side Pond.*

to moderately northeast plunging, close to isoclinal folds. Tuach (1976) noted that the L_M lineation fabric is generally parallel to F_M minor fold axes. I have observed only a few F_M folds within the Pacquet Harbour Group; they are tight, upright, near horizontal class 2 and class 3 folds that clearly display transposition along S_M planes.

Major F_M folds have been noted during the present study and reported from the area by Kennedy (1975a), Tuach (1976) and DeGrace et al. (1976), though there appears to be a discrepancy between workers as to the orientation of these folds. Kennedy (1975a) reported major recumbent isoclinal F_M structures between Ming's Bight and Pacquet Harbour and in Confusion Bay; he indicated that the "limb lengths" of these folds are approximately 6 km and that they are all south facing structures. In contrast, only upright major F_M folds have been reported by other workers. Tuach (1976) reported a local, tight F_M fold in the Rambler area with a wavelength of approximately 700 m; the structure appears to be a steeply inclined syncline with an overturned south limb. This structure appears to complement a similar scale F_M anticline that I have interpreted as lying immediately to the northeast, based on stratigraphic younging directions, cleavage bedding intersections, and the apparent repetition

Plate 7-14: *Birds-eye view of volcanic conglomerate outcrop shown in Plate 5-12; note intense L_M fabric not evident in Plate 5-12.*

Plate 7-15: *Hornblende lineation in Cape St. John intermediate rocks along the coast east of Brent's Cove (photograph by J.R. DeGrace).*

of lithic units (Hibbard and Gagnon, 1980); DeGrace et al. (1976) traced this anticline eastward to Gooseberry Cove. They also indicated the presence of upright to inclined F_M folds to the north of this fold and a major F_M syncline to the southeast (Figure 1-1). These are largely defined on bedding-cleavage intersections, minor F_M fold asymmetry, and a few stratigraphic younging directions. The best defined of these is the syncline to the southeast, the axis of which forms the southern boundary to this structural block; a profile of this fold is exposed at its easternmost exposure, near South Bill (Figure 1-1). Based on this interpretation, almost all of the Cape St. John Group in this block occupies the north limb of this upright structure. The apparent discrepancy between Kennedy's (1975a) major F_M recumbent structures and the upright F_M folds of later workers may actually reflect the change in intensity of D_L from north to south (see Late Deformation below). Thus, in the north, where Kennedy (1975a) carried out most of his work, the orientations of D_M minor structures were greatly influenced by intense D_L vertical shortening and they appear as recumbent structures, whereas southward, where D_L is less intense, D_M minor structures tend to retain their original upright orientation. Even so, there remains a discrepancy in the Confusion Bay area, where DeGrace et al. (1976) depicted upright folds and Kennedy (1975a) interpreted major F_M recumbent structures. This difference in interpretation may be a reflection of the complexity of polyphase structures in the block.

Faulting appears to have been insignificant during the main deformation, though S_M along the Big Brook slide zone appears more intense than elsewhere, possibly indicating movement along the zone during D_M.

LATE DEFORMATION - D_L

Late deformational structures are inhomogeneously overprinted on earlier structures; D_L is most intense to the north and its effects dramatically decrease southward. The southern boundary of the Eastern Orthotectonic Block is most readily defined by the dying out of D_L structures. DeGrace et al. (1976) summarized D_L as involving vertical shortening of earlier structures accompanied by at least two sets of later, local structures.

The S_L fabric ranges from a subpenetrative, tightly spaced crenulation cleavage in the north (Plate 7-16) to a widely spaced fracture cleavage to the south; generally it dips shallowly to the north. In places, it is a L-S fabric defined best by the linear alignment of amphibole. Locally Gale (1971), Tuach (1976) and I have noted a late, steep, northeast trending crenulation cleavage in the Pacquet Harbour Group. Gale (1971) interpreted this fabric to predate the gently north dipping fabric; Tuach (1976) and I have observed this steep fabric affecting the gently dipping fabric and we conclude that the steeper fabric is younger.

Minor F_L folds are common in the area and appear to vary dramatically in amplitude and style from the north to the south (Plates 7-17 and 7-18). In the Ming's Bight Group, F_L minor folds appear to be tight to isoclinal class 3 folds (DeGrace et al., 1976) (Plate 7-17). Southward at Pelée Point (Plate 7-18), F_L folds are tight to close class 2 folds; along Ming's South Brook these folds are close to open class 1B to 1C folds. Concomitant with this overall change in style is an apparent dwindling in size of folds southward. At Brent Cove Head, F_L folds attain amplitudes up to 600 m, whereas to the south at Brent Cove, they are merely metre scale folds (DeGrace et al., 1976). Tuach (1976) reported F_L recumbent folds with amplitudes up to 20 m at the Ming Mine. Most workers have interpreted the F_L minor folds to be southward facing structures.

Coates (1970) indicated the presence of major F_L folds with amplitudes on the order of kilometres between Cape Brulé and Cape Canis. These folds have not been substantiated by most subsequent workers. In contrast to Coates (1970), Kennedy (1975a) suggested the presence of similar scale F_M folds in this area (see F_M above). As with the discrepancy in the interpretation of F_M major folds in the same area, the difference between Coates' (1970) and Kennedy's (1975a) interpretation of major fold structures may well arise from the structural complexity of the area.

Coates (1970) also suggested the presence of a major D_L monocline that steepened all of the structures to the south of the block. DeGrace et al. (1976) indicated that this monocline is only an apparent effect related to the dying out of nearly flat lying D_L structures and the revelation of upright D_M structures in a southward traverse of the structural block.

Previous workers have reported numerous other late structures dominated by local open folds, crenulations, and kink bands (Table 7-4). The most regionally significant of these is the set of broad northeast trending warps that affect the block (Figure 1-1); these flexures plunge steeply north-northeast and affect all structures outlined above (DeGrace et al., 1976).

The Big Brook slide zone (see D_E above) appears to have been a wide zone of intense D_L strain; here, S_L is a very strongly developed, subpenetrative composite fabric and F_L folds are nearly isoclinal. Coarse grained mica schists formed apparently during both D_M and D_L, and kyanite and staurolite also formed in the zone (see Metamorphism, below). At present, displacement during D_L cannot be demonstrated; hence, in Figure 1-1, the slide zone is depicted as a high strain zone.

There is a conspicuous lack of D_L faults in other reports of the block. I suspect that more D_L faults are present, but due to the lack of marker units and the effects of late recrystallization to the north, they are either effectively obscured or obliterated.

METAMORPHISM

Rocks of the Eastern Orthotectonic Block have been subjected to polyphase regional metamorphism; peak assemblages to the north appear to have been formed in the lower amphibolite facies, whereas to the south they attained upper greenschist facies. The change in metamorphic grade through the belt is transitional.

The metamorphic evolution of the block is not well documented; it was studied locally by Coates (1970) and Tuach (1976), and contributions have been made by Gale (1971) and Kennedy (1975a). Coates' (1970) and Tuach's (1976) results, compared in Table 7-5, are different, as they record a different range for the peak metamorphism (note that D_1 of Coates = D_2 of Tuach; see Structure, above); essentially, it appears that peak metamorphism was sustained longer in the north. This discrepancy may reflect the inhomogeneous metamorphic pattern of the block.

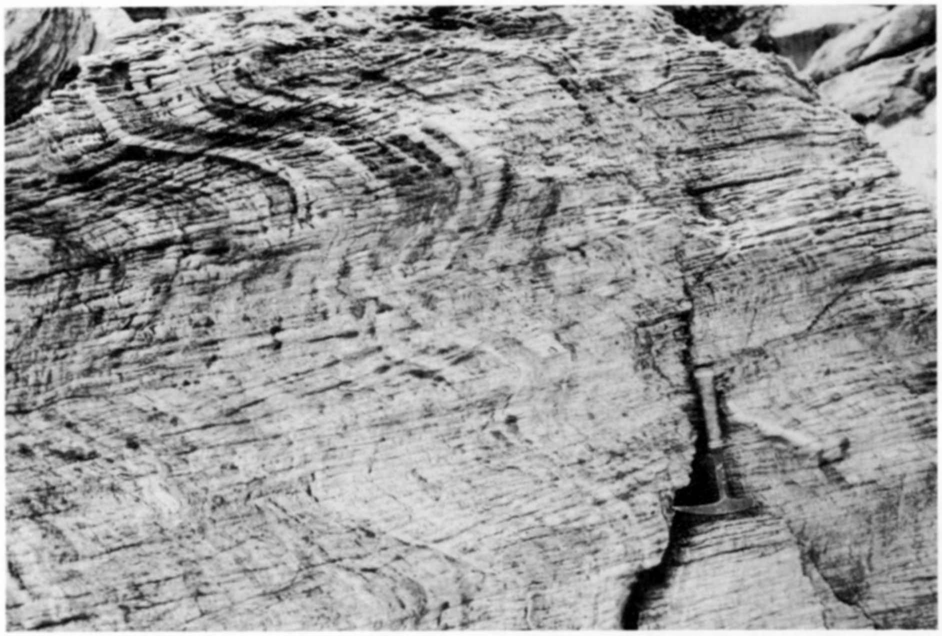


Plate 7-16: *Intense S_L crenulation cleavage affecting both layering and S_M in Cape St. John Group felsic volcaniclastics along coast west of Reddits Cove (photograph by J.R. DeGrace).*

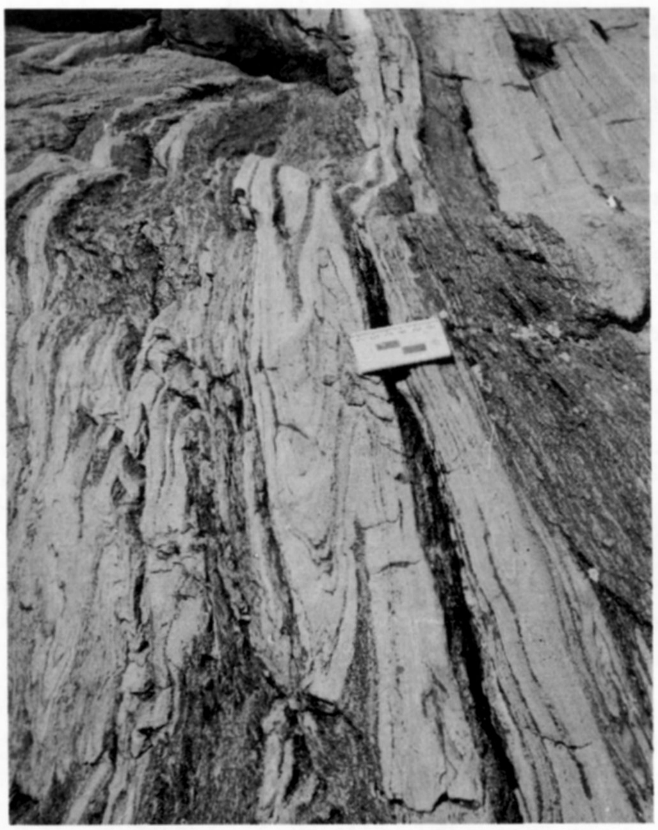


Plate 7-17: *Tight to isoclinal F_L minor folds in Ming's Bight Group metaclastics, just northwest of Pelée Point (photograph by J.R. DeGrace).*

Plate 7-18: *Tight to open F_L minor fold in dike of Dunamagon Granite, just southwest of Pelée Point. M.J. Kennedy scales the fold.*

Available data indicate that peak metamorphism was reached throughout the block by either MS_M or MP_M. The following metamorphic assemblages are common in the area:

Ming's Bight Group schists:

quartz – plagioclase – biotite – muscovite ± garnet ± chlorite ± epidote

northerly mafic lavas:

hornblende – albite – epidote ± quartz ± actinolite ± biotite ± muscovite ± chlorite

southerly mafic lavas

actinolite – albite ± epidote ± quartz

felsic volcanic rocks:

quartz – albite – muscovite ± biotite ± chlorite ± epidote

Table 7-5: *Summary of detailed metamorphic analyses for the Eastern Orthotectonic Block.*

	Coates (1970) Pacquet Harbour-Confusion Bay Area							Tuach (1976) Rambler Area					
Metamorphic Minerals	D_1		D_2		D_3			D_1		D_2		D_L	
Quartz	- - -	- - -	- - -	—				—	- - -	—	- - -	—	- - -
Plagioclase	- - -	—	- - -	—				—	- - -	—			
Biotite	—	- - -	—	—	- -			—	- - -	—	—	—	
Muscovite	—	- - -	- - -	- - -	- -			—	- - -	—	- - -	—	
Chlorite					—	—		—	- - -	—	- - -	—	- - -
Epidote	—	—	—	—	—	—		—	- - -	—			
Actinolite								—	- - -	—	—	- - -	
Hornblende	—	- -	—	—	- -				-	—	—	-	
Garnet	- - -	—	- - -	—						- - -	—		

Most commonly, amphibole defines the two major fabrics S_M and S_L in the mafic rocks; in many places, spectacular, random MP_M amphiboles up to 4 cm long pattern the volcanic-derived rocks in the block.

In addition to these common assemblages, Tuach (1976) and Coates (1970) reported garnet from felsic rocks in the Pacquet Harbour Group and Coates (1970) also noted it in the Cape Brulé porphyry. Locally, I have noted oligoclase in the metavolcanic rocks (optical methods), though Gale (1971) indicated that all of the feldspar in the Rambler area is albite, based on microprobe data. Kennedy (1975a) reported kyanite and staurolite from the Ming's Bight Group at Pacquet Harbour; I located this occurrence on the coast at Pacquet Harbour and found that it extends inland along the Big Brook slide zone. Along the brook, kyanite and staurolite occur as well formed blades and prisms greater than 4 cm long; both are found in highly strained mica schists, but large blue blades of kyanite are more commonly associated with quartz pods in the schists. Kyanite is also a major constituent of the quartzitic slide-like rock type on the brook (see Structure, above, and Plate 7-12). In all cases observed, it appears that both kyanite and staurolite formed after the local main deformation and before the late fabric, and have been localized in the high strain zone associated with the presumed D_E Big Brook slide zone. Neither of the aluminosilicate minerals has been reported from elsewhere in the block; it appears that the Big Brook slide zone was the site of elevated metamorphic conditions with respect to surrounding areas during the hiatus between D_M and D_L. One possible explanation for this pattern is that the Big Brook slide zone acted as a hydrothermal conduit during MP_M peak metamorphism, providing aluminum-rich fluids necessary for local kyanite and staurolite nucleation. Gresens (1971) proposed such an origin for larger scale kyanite deposits in New Mexico.

AGE AND MECHANISM OF DEFORMATION AND METAMORPHISM

Earlier tectonic analyses of the block indicated a pre-Early Ordovician age of deformation for all of the rocks (Church, 1969); this interpretation was based mainly upon deformed detritus, supposedly derived from the block, in the Arenigian Snooks Arm Group. This concept was incompatible with regional field relationships (Neale, 1957; Neale et al., 1975; DeGrace et al., 1976; Williams et al., 1977). In particular, it called for a pre-Early Ordovician age for the Cape St. John Group, which unconformably overlies the Snooks Arm Group. Subsequently, DeGrace et al. (1976) rejected this concept and considered the polydeformation and metamorphism of the block to be Acadian. In the present study, I view the tectonism of the block to be of two ages, including an earlier event, previously unrecognized, followed by later polydeformation.

The zone of ophiolitic blocks at the southern margin of the Ming's Bight Group, in conjunction with the Big Brook slide zone, defines an early structural zone that most likely records the emplacement of oceanic crust over the Ming's Bight Group; hence, it juxtaposes the group against the ophiolitic Pacquet Harbour Group. This event must be Early to Middle Ordovician or older because both the Ming's Bight and Pacquet Harbour Groups were intruded by the Dunamagon Granite at 460 ± 12 Ma. Thus, I view the formation of this zone to be the same early Taconic (D_E) tectonism that formed similar structural zones in Fleur de Lys rocks in the Western Orthotectonic Block. In the Eastern Orthotectonic Block, this event was restricted to the Ming's Bight and Pacquet Harbour Groups; it is impossible for this event to have affected the Siluro-Devonian Cape Brulé porphyry and Cape St. John Group. S_E in the latter units is probably a bedding plane foliation (Church, 1969; DeGrace et al., 1976). The extent of Taconic effects in the older units, outside of the D_E slide zones, is difficult to assess due to the intense, later deformation that may have been coaxial with the earlier event (see Paratectonic Block). Comparison with portions of the Paratectonic Belt that appear to have undergone Taconic deformation (Snooks Arm Group; see below) suggests that the Ming's Bight and Pacquet Harbour Groups were characterized by a slaty cleavage and lower greenschist grade metamorphism following the D_E event.

Subsequent polydeformation of the whole block appears to be Acadian in age, since it affected the Siluro-Devonian Cape Brulé porphyry and Cape St. John Group. In addition, $^{40}Ar/^{39}Ar$ age spectra for metamorphic minerals in the Pacquet Harbour and Ming's Bight Groups (Dallmeyer, 1977; also new data, see Appendix IV) indicate rapid Devonian post-metamorphic cooling of these rocks. The mechanism of this late deformation and metamorphism in uncertain, though it was possibly related to a combination of the following: (i) northward strike-slip movement of the Baie Verte Belt along the north trending segment of the Baie Verte Line and the Green Bay Fault, possibly during the formation of a global scale Variscan shear zone, as outlined by Arthaud and Matte (1977), (ii) elevated heat flow in the Siluro-Devonian volcanic-plutonic terrane, and (iii) southerly overthrusting of rocks from the north due to uplift of the Fleur de Lys Supergroup.

IV. PARATECTONIC BLOCK

The Paratectonic Block embraces all of the Baie Verte Belt except for the parts of the Advocate Complex (see Transition Block), Pacquet Harbour and Cape St. John Groups, Cape Brulé porphyry, and all of the La Scie intrusive suite (see Eastern Orthotectonic Block). Most of the sedimentary and volcanic rocks in the block display a single penetrative fabric, and are at either lower greenschist or subgreenschist metamorphic grade; locally, a second foliation is present. The Burlington Granodiorite and the Middle Arm Ridge Suite appear to be devoid of any significant, regionally developed penetrative foliation; these rocks form a buttress that divides the block into an eastern and a western half. The eastern half of the block is in transitional structural contact with the Eastern Orthotectonic Block; the western half is in tectonic contact with both of the orthotectonic blocks and a Transition Block.

FABRIC AND FOLDING

The main fabric in the block, S_M, is generally a slaty cleavage. Commonly, it is parallel or subparallel to bedding, except in the Betts Cove - Snooks Arm area. Here, it is subparallel to bedding in the eastern areas and crosscuts bedding to the west (Neale, 1957; Upadhyay, 1973). In the west half of the block, S_M is most intense near the Baie Verte Line, and in this area, north of Flat Water Pond, it is accompanied by a clast elongation, L_M. In the area between Sisters Cove and Advocate Mines (Figure 1-1), according to Bursnall (1975), S_M locally transposes a weak earlier fabric. In the east half of the block, S_M is variably developed. It is best developed in the Snooks Arm Group and only weakly developed in the immediately adjacent Cape St. John Group (Neale et al., 1975; Strong, 1980). However, northward, away from the Snooks Arm Group, the fabric becomes pervasively developed in the Cape St. John Group rocks (Neale et al., 1975; DeGrace et al., 1976). In both the Pacquet Harbour Group and Cape Brulé porphyry, the main fabric is best developed in northern parts of the block. Throughout the area, the S_M cleavage is generally best developed in the sedimentary and volcaniclastic rocks and less well developed in pillow lavas, massive lavas and dikes. The S_M fabric appears to be of different ages in the block (see Age, below).

Minor folds (F_M) associated with the S_M fabric have been reported from only local areas. Those in the Flat Water Pond Group are isoclinal, upright folds with axes parallel to L_M (Kidd, 1974) (Plate 7-19). Upadhyay (1973) reported that, in the eastern area, F_M minor folds are best developed in the Bobby Cove Formation; here, they are upright structures that plunge moderately to the east and have wavelengths of approximately 500 m.

Plate 7-19: *Upright F_M minor folds within Flat Water Pond Group volcaniclastic rocks on the north shore of Flat Water Pond.*

Major F_M folds have been delineated in both the western and eastern halves of the block. Neale and Kennedy (1967) outlined a major F_M syncline in the Flat Water Pond and Micmac Lake Groups; the fold, slightly overturned to the east, was subsequently modified by the Flat Water - Micmac Fault (see below), which passes approximately through the hinge of the fold parallel to its axial plane. The fold is defined by the east facing Flat Water Pond Group and slivers of the Micmac Lake Group on the overturned west limb, and the west facing Micmac Lake Group on the east limb (Figure 1-1, cross-section). The tight fold defined by the map pattern of the Kidney Pond conglomerate at Flat Water Pond appears to be a F_M parasitic fold on the larger regional structure.

In the Point Rousse Complex, Norman (1973) first noted the opposing younging directions between the northern and southern portions of the cover sequence on the shores of Ming's Bight. Subsequently, a nearly east trending major syncline was outlined in these rocks (Hibbard and Gagnon, 1980); it is a close to tight F_M fold overturned slightly to the southeast. The syncline is best defined by facing directions in the cover sequence and the disposal of ophiolitic rocks on both the northwest and southeast sides of the cover sequence. Recent work by Noranda Exploration Company Limited (P. Dimmel, personal communication, 1980; Fitzpatrick, 1981) has confirmed the location of the axial trace of the fold (Figure 1-1).

The Snooks Arm Group and Betts Cove Complex occupy the steep to overturned northern limb of a major F_M syncline (Neale, 1957; Upadhyay, 1973). This fold is overturned slightly to the southeast and plunges gently to moderately to the east. The nose of the fold lies near Betts Bight (Upadhyay, 1973) and its axial trace passes through the area of Indian Burying Ground Cove (Neale, 1957). Immediately to the north of this area, DeGrace et al. (1976) delineated a major upright F_M syncline within the Cape St. John Group and Cape Brulé porphyry (see Eastern Orthotectonic Block, above). The axis of the northerly syncline reaches the coast near South Bill (Figure 1-1) where the syncline was first recognized by Baird (1951). Rocks of the Paratectonic Block appear to be confined to the southern limb of the fold, and the axis of the syncline approximately marks the transitional boundary from the Eastern Orthotectonic Block to the Paratectonic Block in the Cape St. John Group. No anticline has been recognized in the intervening area between the synclines described above. If one did exist, it was probably eliminated by faults. Alternatively, the synclines may be of different generation, although roughly coaxial, with no complementary anticline between them (see Age, below).

Locally throughout the block, the S_M fabric was affected by a later deformation, D_L. This late phase is manifested as a S_L crenulation fabric and F_L minor folds in some areas, whereas in others, only F_L minor folds are present. In the western half of the block, S_L is most strongly developed in the Flat Water Pond Group near the Baie Verte Line; typically it is a spaced subvertical crenulation cleavage (Plate 7-20). Locally, along the north shore of Flat Water Pond, moderately to steeply plunging F_L minor folds affect S_M and layering. Church (1969) noted large scale D_L warping of D_M structures in this area. The S_L cleavage here is weaker in the Micmac Lake Group and dips moderately westward (Neale and Kennedy, 1967). Norman (1973) reported that the S_L cleavage is absent from the Point Rousse Complex at Ming's Bight, though F_L crenulations of the main fabric are common.

Church (1969) is the only worker to have reported more than one fabric in the Snooks Arm Group. He indicated that F_M minor folds are crosscut by a later, regionally nonpenetrative foliation. Neale et al. (1975) and DeGrace et al. (1976) reported that further north, where S_M is developed in the Cape St. John Group, it is locally affected by a subhorizontal crenulation cleavage. This fabric is also locally evident within the Pacquet Harbour Group and Cape Brulé porphyry in more northerly parts of the block.

Plate 7-20: *Spaced, subvertical S_L crenulation fabric transposing S_M within Flat Water Pond Group gritty siltstone on the northwest shore of Flat Water Pond.*

FAULTS

Faults are common throughout the Paratectonic Block and are generally coeval with or later than the main fabric. The major faults and fault systems of the block are depicted in Figure 1-1 and described below. The major faults bounding the block along the Baie Verte Line and the Scrape Thrust are described later in this chapter.

Flat Water - Micmac Fault

Neale et al. (1969) and Neale (1962) first recognized and named this fault that traverses the western half of the block from Pittmans Pond to Flat Water Pond. Over most of its length, the fault separates the Flat Water Pond Group from the Micmac Lake Group, though locally, north of Fox Pond, it juxtaposes the main part of the Micmac Lake Group against east facing slivers of the same group (Kidd, 1974). To the south, the fault appears to join the Baie Verte Line; to the northeast of Flat Water Pond (Figure 1-1), the fault feeds into a major thrust zone bordering the east side of the Burlington Granodiorite. Neale (1962) described the fault in the King's Point (12H/9) map area:

> Evidence for the fault consists of steeply dipping shear zones in several localities between the two groups [Micmac Lake and Flat Water Pond] and truncation of structures in a few places - although, generally, the fault appears to be roughly parallel to the strike of bedding, flows and schistosity in both groups. In a few places interpretation of the fault is complicated by the presence of minor intrusions.
>
> Shear zones associated with the fault dip 50 to 70 degrees to the west and this is taken to indicate the dip of the fault zone. Drag folds within the zone of shearing suggest that the west moved upward relative to the east side...

Flat Water - Baie Verte Fault System

A number of faults emanate from the northeast end of Flat Water Pond and trend toward Baie Verte; approximately 6 km north of the pond, this zone narrows and becomes a single, low angle reverse fault that continues northward to the mouth of South Brook on Baie Verte. Here, it is uncertain if the fault either continues out to sea or links with the Scrape Thrust, which continues east from the area. In the Flat Water Pond area, small reverse faults repeatedly imbricate the Burlington Granodiorite with the Micmac Lake Group (Kidd, 1974); further west, the fault places Flat Water Pond rocks over both the granodiorite and the Micmac Lake Group. To the north of the pond, the single fault is marked by a sharply incised gully. Based on drill core data in this area (M.J. Collins, personal communication, 1979), the fault appears to dip moderately to the west. At the La Scie highway, intensely cleaved chloritic schist outcrops on the west side of the depression. In the area of Aspy Cove Brook, the fault is marked by chloritic schist as well as brecciated and finely comminuted granodiorite.

Advocate Complex Fault System

The fault system within this area is complex and has not been analyzed in detail. The earliest faults appear to be northeast trending, layer-parallel faults that bound the major ophiolitic slices in the complex. Bursnall (1975) noted the possibility that a broad mylonitic zone opposite the entrance to Advocate Mines represents the southern extension of the Advocate Western Boundary slide (see Transition Blocks) that was displaced southeastward along the Advocate Fault. Other layer-parallel faults may be related to southeast directed thrust faults that postdated S_M to the north of the complex (Bursnall, 1975). These latter faults were subsequently offset by at least three sets of steep faults that trend generally northeast, east and north (Figure 1-1). The youngest of these appears to be the north trending set, which seems to offset the other fault sets (Figure 1-1).

Point Rousse Complex Fault System

The structural pattern of the Point Rousse Complex is dominated by reverse faults and steeply dipping faults of uncertain displacement sense. Major faults in the Point Rousse area were described by Norman (1973) and Norman and Strong (1975). Many of these faults appear in the field to dip steeply to the northwest and are interpreted as high angle thrust faults (Figure 1-1, cross-section), since they place the deeper seated rocks of the ophiolite suite over higher stratigraphic members. These thrust faults appear to be mainly coeval with the formation of S_M in the area, although Kidd et al. (1977) noted that the moderately northwest dipping thrust planes at Eastern Point gently truncate the main fabric in the greenschists beneath the fault. Locally, ultramafic rock involved along the thrusts is extensively brecciated, for example in the area near Red Point.

A similar though more complex structural pattern occurs in the area of Three Corner Pond. The strip of serpentinized ultramafic rocks immediately north of the pond appears to have been emplaced along a gently to moderately west dipping thrust fault. Greenschist immediately underlying the thrust is intensely cleaved and contorted. At Kidney Pond, it generally dips gently to the west. At Three Corner Pond, identical schists near the fault appear to dip more steeply to the west, though these are highly contorted. The greenschists on either side of this ultramafic slab are intensely foliated and fine grained, but locally less deformed pods of diabase and gabbro are interspersed through the schists. These wide areas of greenschist are most likely highly tectonized gabbro and diabase (see Point Rousse Complex, Chapter V) that have been deformed during the formation of the local fault system. The faults that trend subparallel to the Ming's Bight road may also be thrust faults, since in the Ming's Bight area they are marked by moderately northwest dipping shear zones similar to thrusts on the Point Rousse Peninsula.

Many of the faults in the Three Corner Pond region appear to be high angle faults along which the sense of displacement is uncertain. The best examples are the north-northwest trending faults that bound the ultramafic bodies to the east and northeast of Three Corner Pond. Other minor, steep faults occur throughout the ultramafic body south of Ming's Bight.

Rambler Brook Fault

At the northern boundary of the structural block, the Rambler Brook Fault (Gale, 1971) extends from Rambler Pond southeastward toward the headwaters of Rambler Brook; southeast of the headwaters, exposure is poor and the fault has not been traced further. Along its entire length, the fault is marked by up to 3 m of chlorite schist and chlorite-sericite schist that are intensely foliated parallel to the trend of the zone. At its southeast extremity, the fault affects the Cape Brulé porphyry and the zone is more quartz-rich than to the northwest. Tuach (1976) noted the fault in drill core and indicated that it dips shallowly (approximately 30°) to the north-northeast. The fault apparently truncates the local main fabric on a regional scale and disrupts the local lithofacies pattern (Tuach, 1976). Tuach (1976) suggested that it is a late normal fault. His stratigraphic evidence is not compelling; I suggest that the fault represents a southeasterly directed thrust fault. The Rambler Brook Fault is low angle, post-D_M, and roughly parallels a major thrust fault, the Scrape Thrust, to the north. Thus, I envisage the Rambler Brook Fault either as a splay, or part of a series of thrust faults related to the Scrape Thrust (see below); I have invoked a similar interpretation for the emplacement of ultramafic bodies in the hanging wall at the Ming Mine (see Pacquet Harbour Group).

Betts Cove - Tilt Cove Fault System

Faulting in this area was described by Upadhyay (1973):

> Faulting of the Snooks Arm Group is most intense in the Betts Cove and Tilt Cove regions, and near Tilt Cove faults exert a major control on the outcrop pattern. There are two main orientations of these faults: (i) those parallel to or nearly parallel to the stratigraphic trends of the Snooks Arm Group, and (ii) those oblique to such trends. These have been referred to as "tangential" and "radial" faults respectively, by Snelgrove (1931). The tangential types are very extensive and constitute over 60 percent of Pittman Bight to Beaver Cove and mark the contact between the Snooks Arm Group [and Betts Cove Complex] and the adjacent... [Cape St. John]... terrain.
>
> Another fault extends parallel to this from Betts Big Pond to Long Pond and brings the ultramafic rocks and the pillow lavas of the Betts Cove Ophiolite into direct contact with each other. Faults and lineaments with similar trends also occur in the vicinity of Betts Cove mine. The evidence for the dips of these faults is generally lacking but wherever available, e.g. the cliff faces at Tilt Cove, and Betts Cove, they dip steeply either to the north or to the south. Low angle thrust faults, although rare, occur in Wild Bight (Snelgrove, 1931). The "tangential" faults and the cleavage in the Snooks Arm Group show crude parallelism to each other, indicating some possible genetic relationship...
>
> ...The "radial" faults are most common in the Tilt Cove area. They transect the stratigraphic trends and frequently terminate the geological formations. In the Betts Cove area the fault along the Kitty Pond - Betts Cove trail is a "radial" type.
>
> The faults are generally marked by narrow valleys or topographic depressions. Most of them contain mylonitized zones and, where passing through mafic rocks, chlorite schist. The lineaments, like faults, stand out as narrow linear features on air photos but do not show much evidence of mylonitization.

In addition, Strong (1981) suggested that the fault zone at the south margin of the ultramafic rocks in Tilt Cove is most likely a southeast directed thrust fault since the ultramafic sequence appears to overstep various units in the area. He also cited numerous faults that crosscut the Betts Cove Complex but do not affect the Snooks Arm Group; these faults, he suggested, may be early ocean floor faults (Strong, 1980).

Stocking Harbour Fault

The Stocking Harbour fault is but one of many faults that emanate from the Middle Arm Ridge Suite area; it was first noted by Maclean (1947) and subsequently described along its full length by Baird (1951) and in the King's Point area by Neale (1962). Baird (1951) gave the most complete account of the fault:

> The Stocking Harbour fault is the most prominent fault along the coast of Notre Dame Bay. It is marked along its entire length by a sharply delineated depression, which is most conspicuous in the Stocking Harbour - Rouge Harbour district. It has been traced from a point just north of the head of Middle Arm to Green Bay northeastward to a point northeast of Betts Cove, a distance of more than 20 miles.
>
> In the region south of the White Hills of Burlington, a gulch along this fault is followed for several miles by the stream that enters Burlington Harbour from the southwest. The gulch is steep walled and narrow, and the rocks in the walls show fracturing and shearing but little other direct evidence of the fault. The fault follows the two arms of Burlington Harbour, the elongated arms of Stocking Harbour, and the southwest arm of Rouge Harbour. Sheared and slickensided volcanic rocks were observed in the fault zone near the copper prospect at Rouge Harbour. At the head of a small cove, three-quarters of a mile southwest of Walsh Cove, the fault is exposed in a cliff and displays a vertical sheared and mashed zone about 30 feet wide, but from which the direction of movement cannot be determined. The fault trace passes northeastward under the isthmus at Walsh Cove and across the mouth of Nippers Harbour...

Though Baird (1951) believed that the fault continued inland from Pittman Bight to the Betts Cove area, Neale (1957) demonstrated that the fault intersects and coincides with faults bounding the Betts Cove serpentinite belt at Pittman Bight. Displacement along the fault appears to have involved an uplift of rocks to the south with respect to those to the north since the older Burlington Granodiorite is juxtaposed against the younger Cape St. John Group rocks just to the north of Middle Arm (Figure 1-1).

Middle Arm Fault System

Neale (1962) described this fault in the Green Bay area:

> An east-northeast fault zone, located in the linear stretches of Middle Arm Brook, curves eastward near the head of Middle Arm and strikes easterly parallel to and slightly south of the south shore of the Arm, emerging at the coastline in a little cove opposite the east end of Middle Arm village. The fault has produced cataclasis and mylonitization in syenite, rhyolite and quartz-feldspar porphyry and basic volcanic inclusions within these rocks all along the bed of Middle Arm Brook. Shear zones within these rocks dip very steeply to either the north or south. South of Middle Arm, where outcrops are few, the trace of the fault is delineated by a topographic linear near the base of a steep, wooded scarp. Westward, the fault was not traced beyond the headwaters of Middle Arm Brook...

He noted that an unrecognized southwestward extension of the fault may be responsible for the offset at the base of the Micmac Lake Group near Fox Pond (Figure 1-1). If so, this would indicate a right-lateral sense of displacement along the fault.

During reconnaissance work in the area, I have recognized a fault breccia along the north side of the Middle Arm Valley; the trend of the steep north wall of the valley suggests that this fault may trend parallel to the Middle Arm Fault and may also be the northeastward extension of a fault that offsets the base of the Micmac Lake Group north of Fox Pond. The apparent parallelism and similar displacement sense on these faults indicates that they are probably related to the Middle Arm Fault.

Other Faults

Numerous other faults cut rocks of the Paratectonic Block, although the only other significant set is a group of faults at the north end of the Middle Arm Ridge Suite (Figure 1-1). These faults radiate away from the suite, in a northeast direction, and one of them (Figure 1-1) can be traced from the area near Burlington to the area north of Burtons Big Pond. Since all of the Siluro-Devonian rocks involved in the faulting appear to have been related to a Paleozoic caldera, these faults may mark the collapse of the caldera.

METAMORPHISM

All of the rocks in the Paratectonic Belt are at greenschist or lower metamorphic grade, with the exception of rocks at the perimeters of major plutons which have been hornfelsed locally to lower amphibolite grade. All of the mafic rocks in the area were subjected to lower greenschist grade regional metamorphism and typically contain the following assemblage (Norman, 1973; Upadhyay, 1973; Kidd, 1974):

actinolite – albite – chlorite ± epidote group minerals – sphene ± calcite ± sericite ± quartz

Equivalent grades were attained locally in the Micmac Lake and southerly Cape St. John Group (Kidd, 1974; DeGrace et al., 1976) though recrystallization is inhomogeneous in these units.

Upadhyay (1973) recognized an increase in the metamorphic gradient down section, through the Snooks Arm Group and into the Betts Cove Complex; he originally attributed this to burial metamorphism. Regional greenschist metamorphism associated with S_M overprinted the earlier event. Coish (1977a,b) later demonstrated that the earlier metamorphism was an ocean floor metamorphism, and documented the detailed geochemical changes involved in this event.

Contact metamorphism was developed in the Pacquet Harbour Group within approximately 1 km of the Burlington Granodiorite. Mafic rocks in the Burlington - Middle Arm area, here considered as the southeast extension of the Betts Cove lavas, are also hornfelsed. The Pacquet Harbour rocks attained lower amphibolite grade metamorphism; the grade of the other mafic rocks is uncertain, though, based on their similarity to the contact metamorphosed Pacquet Harbour rocks, they are also probably at lower amphibolite grade.

AGE AND MECHANISM OF DEFORMATION AND METAMORPHISM

Rocks in the block on either side of the Middle Arm Ridge - Burlington Granodiorite structural divide appear to have undergone separate tectonic histories. Units in the block on the west half of the divide all appear to have been affected by the same D_M event. Since this event affected the Micmac Lake Group, it had to be Silurian or younger in age. Nearly flat lying Carboniferous rock in the block near Indian Pond sets the upper age limit of this tectonism and indicates that deformation in this west half is Acadian.

At a cursory level, it appears that the eastern half of the block has the same tectonic history as the western half, as it appears that one fabric with associated major upright folds affects all these units. DeGrace et al. (1976) related these features to Acadian deformation. However, observations by Strong (1980), Church (1969), Neale et al. (1975) and Maclean (1947) suggest that the tectonic history in this area is more complex than that to the west. Most cogently, Strong (1980; personal communication, 1981) observed quartz-feldspar porphyry cutting the Betts Cove Complex and feeding relatively undeformed Cape St. John pyroclastic rocks that unconformably overlie severely deformed and altered ultramafic rocks of the complex. He concluded that the metamorphism and deformation of the Snooks Arm Group and Betts Cove Complex predated deposition of the Cape St. John Group, though he noted that later zones of intense deformation locally affected all of these units. Neale et al. (1975) reported that Cape St. John Group rocks immediately north of the Snooks Arm Group have only a weak fabric, whereas the Snooks Arm rocks and Cape St. John rocks further to the north are all well cleaved. Church (1969) locally noted a second, weak fabric in the Snooks Arm Group. All of these relationships indicate that the Snooks Arm Group probably underwent a post-Arenig, pre-Cape-St.-John deformational episode. Following the deposition of the Cape St. John Group, Acadian deformation inhomogeneously affected the area; its influence was minor in southern parts of the Cape St. John Group and the Snooks Arm Group. Thus, the main deformation of the Snooks Arm Group and Betts Cove Complex is here considered, in a broad sense, as Taconic, whereas deformation of the Cape St. John Group is Acadian. The weak second fabric reported by Church (1969) from the Snooks Arm Group may be an Acadian fabric. The age of deformation of the Pacquet Harbour Group in the block is uncertain, though, since it is correlative with the upper parts of the Betts Cove Complex (see Pacquet Harbour Group, Chapter V), it may also be Taconic.

Significantly, Maclean (1947) outlined a similar deformation history for nearby correlative sequences on the southeast side of Green Bay. Here, he indicated, the Early Ordovician Lushs Bight volcanics were strongly folded and sheared during Taconic events whereas the overlying Silurian(?) Springdale Group was gently folded during a later, possibly Acadian, event. Thus, there is regional evidence for Taconic deformation in the eastern part of the Paratectonic Block.

V. HORSE ISLANDS BLOCK

On the basis of their geographic isolation from the Baie Verte Peninsula, the Horse Islands are assigned to a distinct structural block. The nature of the basement to the block is uncertain.

STRUCTURE

The Horse Islands Group underwent at least three major phases of deformation that are recognized primarily by cross-cutting fabric relationships, systematic variation of orientation data and, to a lesser extent, the superposition of folds. There is a distinct contrast in the development of structures between the two islands; structures related to the main deformation phase (D_M) are relatively simple and prominent on the Western Island, whereas structures due to later phases of deformation (D_L) strongly overprint D_M structures on the Eastern Island. Evidence of an early deformation, D_E, which predates D_M, is present locally on both islands. The following structural history of the Horse Islands Group is preliminary and tentative.

Layering and the main penetrative fabric (S_M), on both islands, are generally parallel and steeply dipping; they trend consistently to the northeast on Western Island, whereas on Eastern Island their trend is variable. S_M is of variable intensity and is composed of a quartz grain elongation (predominantly L-S fabric) within psammites and a well developed micaceous schistosity within pelites and semipelites. Minor folds associated with S_M are generally tight to isoclinal (Plate 7-21), characteristically have curved hinge lines, and in places fold an earlier fabric (S_E) defined by platy minerals. In thin section, S_E is well preserved as inclusion trails within helicitic plagioclase porphyroblasts; the trails are defined by chlorite, epidote and plagioclase. Locally, it appears that F_E minor folds have been refolded by F_M folds (Plate 7-22). Large scale fold structures related to S_E and S_M have not

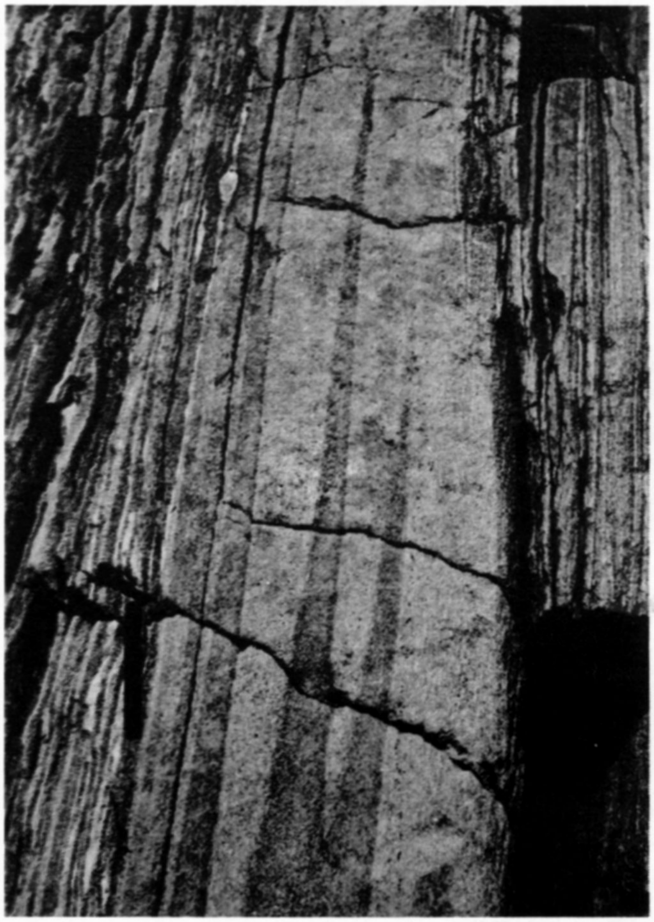


Plate 7-21: *Isoclinal F_M minor fold in psammite and semipelite of the Hit or Miss Point Formation, Horse Islands Group, on Eastern Island.*

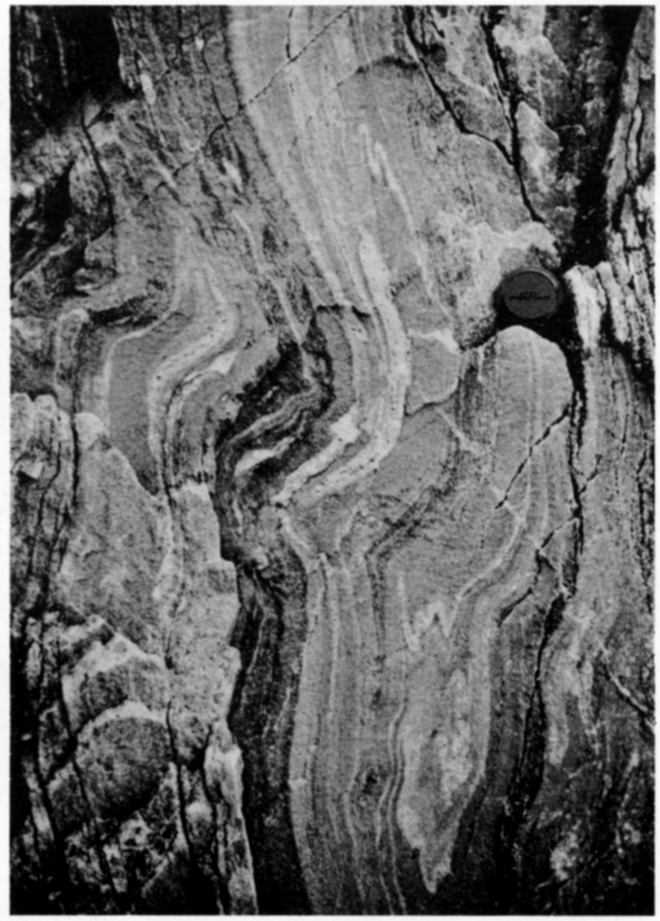


Plate 7-22: *Prominent F_M and F_L folds within semipelitic schist of the Eastern Island Formation, Horse Islands Group, on Eastern Island. Note fold interference pattern, at lower right of photograph, involving F_M folds and a possible F_E fold closure.*

been recognized on the island, although detailed structural mapping and the distribution of mafic and pelitic schists near the Horse Islands community indicate that tight D_M folds with amplitudes greater than 500 m may occur. Such closures have been inferred on Figure 1-1 for the abrupt termination of mafic schist units along the southern coast of Eastern Island.

Post-D_M structures have yet to be fully separated and are considered collectively as the late deformational phase (D_L). On the Western Island, D_M structures are overprinted by a single coarse crenulation cleavage (S_L) that consistently trends to the north-northeast and is associated with open to tight folds. These late folds do not regionally deflect D_M structural trends on this island. On Eastern Island, three distinct phases of late structures have been identified. A pronouncedly heterogeneous deformation event resulted in the development of 160° striking shear zones along S_M and layering; a weakly developed cleavage outside these zones is probably related to this event. These structures are deformed by two coarse crenulation cleavage sets and associated steeply inclined chevron folds. Major deflections in the strike of layering and S_M is probably the result of folding about the more consistent, northeast trending set. Conflicting evidence regarding the relative age of the two cleavages indicates that they may be genetically related (conjugate pair) or that further deformation has disrupted any original relationship. The diabase dikes at the Horse Islands community have been gently warped by at least one of these later deformations.

METAMORPHISM

Rocks of the islands have been polymetamorphosed and retain an upper greenschist facies metamorphic assemblage. Limited petrographic study has resulted in the compilation of a tentative metamorphic history for the Horse Islands Group. The earliest metamorphic event appears to have been synkinematic with D_E and included the growth of muscovite, chlorite, epidote, quartz and plagioclase. Growth of these minerals also appears to have occurred during D_M, but the main porphyroblast growth was post-D_M. Straight and warped S_M inclusion trails within plagioclase porphyroblasts indicate that they grew during at least part of D_L. Garnet porphyroblast growth post-dates at least one of the later deformations. The presence of actinolite and chlorite in the late diabase dikes, as well as chlorite pseudomorphs after garnet in the metafelsite layers and abundant chlorite in the schists throughout the area may indicate a late, low grade metamorphic event.

AGE AND MECHANISM OF DEFORMATION AND METAMORPHISM

Direct evidence for the age of tectonism in this block is lacking, though deformation and metamorphism of the Horse Islands Group is most reasonably viewed as being related to that of the other portions of the Fleur de Lys terrane. As in the rest of the supergroup, the Horse Islands Group was probably subjected to Early Ordovician obduction events (see Western and Eastern Orthotectonic Blocks). It is uncertain if the ensuing polydeformation of the group is related to other Taconic events, as in the main Fleur de Lys outcrop belt, or to Acadian orogenesis, as in the Ming's Bight Group. The mafic dikes on the Eastern Island may be significant in this problem; if they are correlative with mafic dikes that intrude the Eastern Orthotectonic Block (see mafic dikes), their relatively undeformed state as compared with the latter, polytectonized rocks would suggest that the tectonic events on the Horse Islands were mainly Taconic. This interpretation is tentatively accepted in the present report, in lieu of more compelling evidence.

VI. GRANBY ISLAND BLOCK

The Granby Island structural block is confined to Granby Island, White Bay, and is distinguished from the nearby Western Orthotectonic Block by its less intense deformation and metamorphism. The contact between these structural blocks is unexposed, but it is probably faulted along a splay of the Cabot Fault Zone, which enters White Bay approximately 24 km south of the island. The basement to this block is unexposed, but it is most likely Grenville gneisses since it is flanked by the Grenvillian Long Range inlier to the west and probable Grenvillian gneisses of the East Pond Metamorphic Suite to the east (see Western Orthotectonic Block).

Aside from primary depositional features, the main observable structural feature of the island is a steep, west dipping, northeast trending slaty cleavage that varies in intensity. In some of the coarser, carbonate-rich feldspathic sandstones on the north coast of the island, this fabric is better termed a schistosity. Transportation of bedding is common along the main cleavage (Plate 7-23); locally, on the west side of the island, the slaty cleavage grades into a crenulation cleavage wherein it transposes an earlier fabric that is defined in thin section by very fine grained mica and quartz domains. Along the north coast of the island, the main cleavage was affected by two sets of kink bands (070° and 340°, with steep westerly dips) that may represent a conjugate couple.

The main fabric of the block is axial planar to close, gently northeast plunging, steeply inclined folds that have an amplitude on the order of tens of metres. In profile they appear to be most like Ramsay's class 1C (nearly parallel). These folds appear to control the geometry of strata in this structural block (Figure 1-1). Locally, as at the southwest corner of Granby Island, a clast elongation lineation is developed in the coarser sediments and is roughly parallel to the axes of these folds.

Plate 7-23: *Transposition of bedding (directly below hammer) in slate of the Granby Island Formation, north shore of Granby Island.*

The Granby Island Formation has been subjected to greenschist grade metamorphism as indicated by the prevalence of muscovite and the common occurrence of biotite.

The age of tectonism in the Granby Island Block is uncertain; stratigraphic correlation (see Granby Island Formation, Chapter IV) indicates that the unit is most likely Early to Middle Ordovician in age. Thus, these rocks may have been deformed and metamorphosed synchronously with those of the Fleur de Lys Belt. Alternatively, deformation of the Granby Island strata may be attributed to movements along the Cabot Fault Zone which strikes through White Bay; however, the incipient development of regional metamorphism on the island suggests a closer link with the nearby regionally metamorphosed Fleur de Lys rocks.

INTERBLOCK FAULTS

Most contacts between the structural blocks on the peninsula are marked by major faults, herein termed interblock faults. These include the Scrape Thrust and associated faults, which separate the western part of the Paratectonic Block from the Eastern Orthotectonic Block, and the Baie Verte Line and associated fault systems, which separate the Western Orthotectonic, Transition, and Paratectonic Blocks. Part of the Baie Verte Line passes through the Eastern Orthotectonic Block and is discussed here because of its similarity to other parts of the line. The structural histories of the interblock faults are critical to any structural correlation between structural blocks.

SCRAPE THRUST AND ASSOCIATED FAULTS

The boundary between the western half of the Paratectonic Block and the Eastern Orthotectonic Block is marked by the Scrape Thrust, which juxtaposes the Point Rousse Complex to the north against the Pacquet Harbour Group to the south; subsidiary faults mark this contact along the southeast side of the complex, and both a high angle fault and a thrust fault separate the complex from the Ming's Bight Group (Figure 1-1).

Description

The Scrape Thrust has been traced from South Brook in the northwest to the area east of Three Corner Pond (Figure 1-1). It is named after the cliff face, locally known as the Scrape, at the northeast end of Scrape Pond (Figure 1-1). The fault is well defined by a sharply delineated depression along most of its course. The rocks in the area of the fault are highly strained, and locally protomylonitic rocks are found in the Pacquet Harbour Group immediately bordering the fault. Characteristically, the thrust is marked by small pods of serpentinite, talc-carbonate and quartz-carbonate alterations of ultramafic rock; these are best exposed at the pond just northwest of Scrape Pond and along the Ming's Bight road (Figure 1-1).

The thrust zone is best exposed in the Ming's Bight road area, where it is approximately 70 m wide. Here, highly strained greenschists of the Point Rousse Complex structurally overlie a zone of intensely cleaved, buff weathering, talc-carbonate schist (Plate 7-24) to the south. The northwest trending fault contact here dips approximately 40° to the northeast. The talc-carbonate schist grades structurally downward into more massive, though still highly deformed serpentinite. This serpentinite is juxtaposed against very intensely foliated amphibolites of the Pacquet Harbour Group along a moderately to steeply dipping fault plane. The uppermost 2 m of the Pacquet Harbour Group is very thinly layered (millimetre scale) and is affected by tight late folds that plunge moderately to the northeast (Plate 7-25).

Structural Relationships

The orientation of fabrics and layering in the region of the fault suggests that it is a composite zone, related to deformational events in both the Paratectonic and the Eastern Orthotectonic Blocks. Both layering and the S_M fabric in the Pacquet Harbour Group appear to be molded around ultramafic pods in the zone, parallel to contacts with the pods. However, on a regional scale, it appears as though both structures are truncated at a low angle by the thrust in sections where ultramafic pods are absent. Structures in the Point Rousse Complex appear to be parallel to the fault zone along much of its length. These relationships suggest that the Pacquet Harbour - ultramafic fault contact is older and is locally truncated by the younger, structurally higher Point Rousse - ultramafic fault contact. The structurally lower, older fault may have formed during D_M of the Eastern Orthotectonic Block as indicated by the parallelism of the fault, layering and the S_M fabric; similarly, the younger, structurally higher fault involving Point Rousse Complex rocks is likely related to D_M of the western Paratectonic Block. These relationships infer that D_M of the Eastern Orthotectonic Block is older than that of the western Paratectonic Block.

East and northeast of Three Corner Pond, the thrust and the high angle faults between the Paratectonic and Eastern Orthotectonic Blocks are poorly exposed, with the exception of the thrust fault that separates the Ming's Bight Group from a tongue of serpentinite on the east side of Ming's Bight. The thrust is marked by a breccia zone containing deformed frag-

Plate 7-24: *More resistant, highly deformed greenschist of the Point Rousse Complex structurally overlying intensely deformed talc-carbonate schist pod along the Scrape Thrust on the Ming's Bight road.*

ments of the Ming's Bight Group. Kennedy and Phillips (1971) correlated fabrics in the group through this thrust zone and into the margin of the serpentinite; Kidd et al. (1978) refuted the correlation. It appears as though this thrust postdates the major fabrics in the Ming's Bight Group based on the predeformed fragments in the zone.

Plate 7-25: *Amphibolitic tectonite of the Pacquet Harbour Group that directly underlies the talc-carbonate schist in Plate 7-24.*

BAIE VERTE LINE AND ASSOCIATED FAULT SYSTEMS

The term Baie Verte Line is herein used to designate the northernmost segment of the Baie Verte - Brompton Line, which has been defined as a "narrow structural zone marked by discontinuous ophiolitic complexes" along the west flank of the Canadian Appalachians (St-Julien et al., 1976; Williams and St-Julien, 1978, 1982). This feature has also been termed the Baie Verte Lineament (Bird and Dewey, 1970; Kidd, 1977), though Kidd's (1977) definition of the lineament is unwieldly since it includes rocks of the Flat Water Pond and Micmac Lake Groups. For the purposes of the present, peninsula-scale study, I herein define the Baie Verte Line as the tectonic zone which separates the Fleur de Lys Belt from the Baie Verte Belt (Figure 1-1); thus, the line separates the Western Orthotectonic, Transition and Paratectonic Blocks; in addition, it passes through the Eastern Orthotectonic Block.

To the northeast, the line appears to extend eastward from Pacquet Harbour, beneath the waters of the Atlantic Ocean; to the south, the line is truncated by the Green Bay Fault. In the Point Rousse area, the line is unexposed, as it has been overthrust by the Point Rousse Complex (see Structural relationships, below).

The line is a fundamental structure that was subsequently structurally modified in different ways. Tectonic zones that record the early history of the line are preserved along its length but define the line only on the east half of the peninsula. These zones have been polydeformed and polymetamorphosed. South of Marble Cove, the line was accentuated and is defined by late faults which postdate most of the tectonism of the early structural zones. In this area, the line involves at least two generations of late faults that bound the Transition Blocks.

The Baie Verte Line is the most significant structure on the peninsula. Based on relationships between faults that define the line and regional structural patterns, it appears that regional tectonism on the peninsula has centered upon the line.

Early Tectonic Zones

Structural zones attributed to the earliest phases of the local tectonic history are preserved along the entire length of the line, but define it only in the area east of Ming's Bight. Here, the line is defined by mélange zones within the southern part of the Ming's Bight Group and the rocks along the tentatively D_E phase Big Brook slide zone (for descriptions see Eastern Orthotectonic Block). Following D_E, these zones were polydeformed and polymetamorphosed.

Where the line is defined by late faults (see below), early structural zones are generally preserved to the northwest in the Fleur de Lys terrane. The D_E Slaughterhouse Slide (Bursnall, 1975) and Coachman's Mélange (Williams, 1977a) occur just north of the Marble Cove Slide (see below), which defines the line north of Baie Verte. The Roadside Slide (Kidd, 1974) appears to be in close association with the Baie Verte Road Fault system (see below), which represents the line south of Baie Verte. Thus, the late faults that define most sections of the Baie Verte Line appear to be keyed to earlier zones of disruption.

All of these early tectonic zones that either define or are closely associated with the Baie Verte Line have been related to the same Early Ordovician emplacement of ophiolitic rocks over the Fleur de Lys terrane (see the orthotectonic blocks). Since the terrane to the east of the line is underlain by Paleozoic ophiolitic crust, the Baie Verte Line must represent the early root for ophiolites that were obducted over the Fleur de Lys terrane (Bursnall, 1975; Williams et al., 1977; Hibbard, 1982) during the widespread Early Ordovician D_E event.

Late Faults

The segment of the line between Birchy Lake and Baie Verte involves at least two generations of late faults that border the Transition Blocks. This system is herein termed the Baie Verte Road Fault system. Just north of Baie Verte, the line is offset to the west by the Advocate Fault (Figure 1-1) and further affected by a steep, late, northerly trending fault [Figure 1-1]. North of this structurally disturbed area, the line is defined by a late fault, the Marble Cove Slide, on the north side of the northern Transition Block; the tectonic zone bounding the south side of the block here may be equivalent to part of the Baie Verte Road Fault system. The major displacement on all of the D_L steep faults appears to involve downthrow movement to the east.

BAIE VERTE ROAD FAULT SYSTEM

This system involves two major faults that bound the ultramafic rocks of the Transition Blocks south of the Advocate Fault (Figure 1-1). The westerly fault was first recognized by Watson (1947) and later was termed the Baie Verte Road Fault (Neale and Kennedy, 1967); the easterly one is herein referred to informally as the Virginite fault. The Virginite fault appears to be the younger of the two, since faults that offset the Baie Verte Road Fault are truncated by the Virginite fault (e.g. east of Upper Red Cliff Pond) (Kidd, 1974); hence, where ultramafic bodies are present, the Baie Verte Line is defined by the Baie Verte Road Fault, whereas it is defined by the Virginite fault where the bodies are absent.

The Baie Verte Road Fault is poorly exposed; a completely exposed section through the zone is lacking. The best sections of the fault are on Middle Arm Brook and in an area about 1 km north of the Burlington road - Baie Verte highway junction. These locations were described by Kidd (1974):

> ...[On Middle Arm Brook]... there is a 40 metre gap between normal chlorite/biotite grade semipelitic Fleur de Lys schists and schistose serpentinite. About 40 metres of schistose and fish-scale serpentinite adjoin a rapid transition to undeformed serpentinised ultramafic rock to the east. The foliated serpentinite in this locality is abnormal, in that it is mostly statically recrystallised and contains a relative abundance of carbonate (approx. 20%). The foliation seen in outcrop is mostly due to the weathering out of lenticular carbonate veins, which formed subparallel to, and later than the one original subvertical foliation. In places poorly-developed coarsely-spaced slip planes cut the foliation at a low angle, occasionally accompanied by small impersistent crenulations. A few larger asymmetric crenulation folds of the same type, with a wavelength up to a metre and verging up toward a hypothetical antiform to the west, are seen in the outcrop to the south of the Brook. These resemble F_3 folds in the Fleur de Lys, but no such F_3 folds are seen in the Fleur de Lys outcrop to the west.
>
> The other well-exposed locality (north of the Burlington Road) shows platy fish-scale to schistose serpentinite with one vertical foliation 8 metres from normal chlorite/biotite grade interbanded mafic and semipelitic schist. A few metres of schistose/fish-scale serpentinite grades rapidly eastward to a few metres of shear polyhedra serpentinite, and in turn to massive serpentinised ultramafic rock. This exposure is typical, although better-exposed, of all other localities near the western contact of the ultramafic bodies.

The Virginite fault occurs on the east side of the ultramafic bodies on the line and defines the line where these are absent. Virginite (magnesite-quartz-fuchsite rock described in Chapter V) is common along the fault, and Kidd (1974) estimated that it is present along 40% of the fault between Micmac Lake and Flat Water Pond. He described the occurrence of virginite along the fault:

> Long lenticular bodies of virginite up to 80 metres wide, but not usually more than 40 metres wide, are found along parts of the [fault] contact, where the large ultramafic bodies are absent... It is also found, in the same long lenses, on the eastern margin of the Flatwater ultramafic body from south of the Bear Cove Road to just north of Middle Arm Brook. In the area just north and south of Middle Arm Brook there is a narrow strip of the ultramafic body to the east of the virginite...
>
> Roadcuts through the virginite within the ultramafic body just south of Middle Arm Brook show that it is bordered by a zone of foliated carbonate serpentine-talc rock about a metre wide, which is in turn adjoined by a narrow zone of schistose/fish-scale serpentinite, which passes outward through shear polyhedra serpentinite to massive ultramafic rock... The virginite terminates just to the north of Middle Arm Brook, and the tectonic discontinuity continues northward within the ultramafic body as a poorly-exposed zone of foliated carbonate-serpentine-talc rock, bordered by shear polyhedra serpentinite...
>
> ... The particular elements, and their proportions in the virginite indicate that it is derived mostly or entirely from ultramafic rock... The way it was formed is not known, but the large CO_2 content may be explained by channeling of the CO_2, derived from microinclusions in a large volume of serpentinizing ultramafic rock, along a major tectonic discontinuity.

Where these ultramafic and ultramafic-derived rocks are absent, the Fleur de Lys Supergroup is juxtaposed directly against the Flat Water Pond Group. The best exposure of this contact is along the powerline immediately west of the Baie Verte highway, approximately 3 km north of the Burlington road junction. Here, very intensely cleaved slates, mafic volcanic rocks and marble are approximately 8 m from the greenschist of the Birchy Complex. The slates here resemble "paper schists" and the marble is mylonitized; the greenschists in this zone display incipient brecciation at their eastern margin.

Structural relationships: The relationship between faults of the Baie Verte Line and the surrounding regional deformation in the Western Orthotectonic, Transition and Paratectonic Blocks is complex and uncertain; it has been a major topic of controversy in the area (Neale and Kennedy, 1967; de Wit, 1972; Kidd, 1974; Burnsall, 1975; Bursnall and de Wit, 1975). This point is significant in terms of structural correlation between the major structural blocks. Essentially, Kidd (1974) maintained that the Baie Verte Road Fault truncates structures of the three major deformation phases in the orthotectonic block and that structures in the ultramafic rocks are separate from any in the westerly terrane, whereas most other workers indicate that ultramafic bodies of the Transition Blocks were involved in the regional deformation of the Western Orthotectonic Block (Neale and Kennedy, 1967; de Wit, 1972; Bursnall and de Wit, 1975).

Kidd (1974) indicated that the Baie Verte Road Fault and the Virginite fault both truncate D_L structures on a regional scale and that these faults truncate the Wild Cove Pond Igneous Suite. Thus, he interpreted minor structures within the Transition Block ultramafics to be related to either D_M of the western part of the Paratectonic Block or movement along the fault, or both. Based on the parallelism of the Virginite fault with the S_M fabric in the Paratectonic Block, he related the fault to D_M of this part of the block.

Other workers have made structural correlations across the Baie Verte Road Fault; Neale and Kennedy (1967) believed that the Micmac Lake ultramafic body was involved in F_L folding of the Western Orthotectonic Block; Kidd (1974) demonstrated that the unusual pattern of this body, just south of Wild Cove Pond, was probably due to post-D_L folding and faulting. De Wit (1972) indicated that fabrics along the western margin of the Flat Water ultramafic body were equivalent to S_M and S_L of the Western Orthotectonic Block, based on the parallelism of the fabrics. In addition, he indicated that S_L of the orthotectonic block is most likely equivalent to S_M of the Baie Verte Group, because of their mutual parallelism. Bursnall and de Wit (1975) reiterated this interpretation, with further support based on relationships to the north. Here, Bursnall (1975) demonstrated that fabrics in the Western Orthotectonic Block are equivalent to fabrics in the Advocate Complex in the northern Transition Block (see Transition Blocks); this supports de Wit's (1972) view.

On the basis of these observations and new data in the present study, I suggest that ultramafic rocks of the Transition Blocks were emplaced against the Western Orthotectonic Block and deformed during the tectonism of the latter block, but that this junction was a zone of continued movement; the Virginite fault later truncated the Baie Verte Road Fault during D_M of the western Paratectonic Block. More precise structural correlation between the blocks has not been established at this time.

MARBLE COVE SLIDE

This tectonic zone, north of the Advocate Fault, was first recognized by Watson (1947) in Marble Cove, and was named and described by Bursnall (1975). It intervenes between the Western Orthotectonic and northern Transition Blocks in the area north of Baie Verte. The slide separates Rattling Brook psammitic schists from serpentinized ultramafics of the Marble Cove sequence of the Advocate Group between Marble Cove and Seal Cove (Figure 1-1); inland, the slide separates the Birchy Complex from Marble Cove sequence metagabbro (Figure 1-1). The slide not only marks the southeasterly limit of the Fleur de Lys Supergroup, but also marks a major change in structural trends of the area (Figure 1-1). Northeast trending lithological units and minor structures of the Fleur de Lys Supergroup are truncated by the slide; south of it, the Marble Cove sequence rocks and structures trend east-northeasterly.

Bursnall (1975) indicated that movement along the slide occurred many times during D_L of the Western Orthotectonic and northern Transition Blocks. He also indicated that earliest movements may have been during D_M. Though the fault is relatively late in the local tectonic history, Bursnall (1975) traced D_M and D_L structures of the Western Orthotectonic Block into the northern Transition Block. The slide may be equivalent to the Baie Verte Road Fault system. Thus, the orthotectonic block and northern Transition Block appear to have a common deformation history.

The steep, late fault at the southern margin of the northern Transition Block appears to be localized along an unconformity (see Advocate Complex) across which structural correlation is difficult (Bursnall, 1975). This fault may be a northerly equivalent of the Virginite fault.

Seaward Extension of the Line

The east-west trending part of the Baie Verte Line is directly on strike with a major subsea structure, interpreted from seismic and magnetic data, off the north coast of the Baie Verte Peninsula and emanating from Pacquet Harbour (Haworth et al., 1976) (Figure 7-5). The northeastward extent of the subsea structure is uncertain. Remarkably, the surface trace of this structure mimics the southern limit of polydeformation and upper greenschist to amphibolite grade metamorphism on the eastern portion of the peninsula (Figure 1-1). Acadian polydeformation of the volcanic terrane thus appears to be keyed to the easterly trending part of the Baie Verte Line and its possible seaward extension.

Summary of Relationships along the Baie Verte Line

The Baie Verte Line is the most significant structure on the Baie Verte Peninsula. It marks the early tectonic zone along which the Fleur de Lys Belt was juxtaposed against the Baie Verte Belt. The Baie Verte Belt ophiolites were emplaced westward over the former terrane. These events are recorded by early faults and mélange zones marked by relict fragments of ophiolitic rocks. Based on these relationships and the nature of the Fleur de Lys and Baie Verte Belts, the line has been interpreted as marking the early Paleozoic continent-ocean interface (St-Julien et al., 1976; Williams and St-Julien, 1978, 1982).

Subsequent deformation on the peninsula is centered on this major structure, but inhomogeneously distributed. In addition, these later structures have a reverse, west-over-east, polarity from the earlier event, as late faults that mark the line are consistently downthrown to the east. The segment of the line from Birchy Lake to Marble Cove may be as old as D_M of the Western Orthotectonic Block; thus, ultramafic pods of the Advocate Complex along the line were probably deformed, in part, with the Fleur de Lys rocks. Subsequent movement along the Virginite fault reactivated parts of the line and juxtaposed the Transition Block ultramafics against the Paratectonic Block; this event was most likely Acadian, the age of D_M in the latter block. The east trending segment of the line between Ming's Bight and Pacquet Harbour appears to have been unaffected by late faults, but it was subjected to the Acadian deformation and metamorphism of the Eastern Orthotectonic Block. Considering the seaward extension of this part of the line, Acadian deformation appears to have been centered upon the line.

BAIE VERTE FLEXURE

Church (1965b) first noted the change in geological trends of rocks in the Baie Verte Belt from north-northeasterly to east-west; in this study, it has been noted that the Fleur de Lys Supergroup and the Baie Verte Line are also deflected eastward around the northern part of the peninsula (Figures 1-1, 7-5). This major flexure is termed the Baie Verte Flexure (Hibbard, 1982).

The continuation of the Fleur de Lys Supergroup along the eastern limb of the Baie Verte Flexure has previously been unrecognized largely because it is almost totally submerged beneath the Atlantic Ocean. The extension of the supergroup around the northern portion of the peninsula is strongly inferred from the generally east-west structural trends of the Ming's Bight Group and of metaclastic rocks on the Horse Islands (Hibbard and Bursnall, 1979), both of which are lithically identical to and here included in the supergroup. In addition, Haworth et al. (1976) identified a seismically opaque unit off the north coast of the peninsula that is continuous between the main outcrop belt of the supergroup and the Horse Islands and Ming's Bight Groups; they interpreted this unit as an offshore extension of the Fleur de Lys Supergroup. On the Baie Verte Peninsula, the eastward continuity of the Fleur de Lys Supergroup, from the main outcrop belt toward

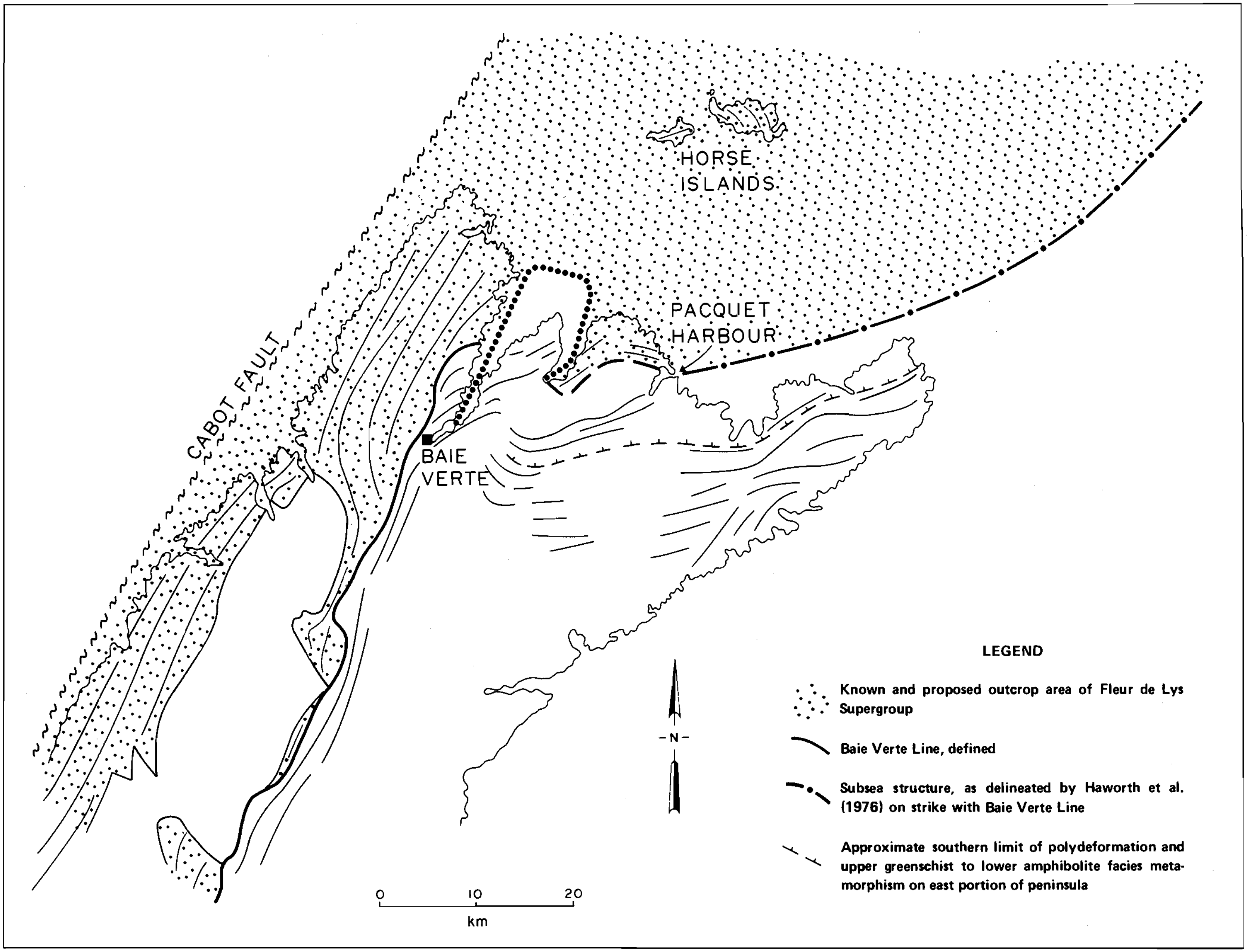


Figure 7-5: *Major elements around the Baie Verte Flexure.*

Ming's Bight, is apparently disrupted (Figure 7-5) by ophiolitic and volcanic rocks of the Point Rousse Complex (Williams et al., 1977). These rocks exhibit a single penetrative fabric and lower greenschist facies metamorphism, and thus must represent a higher structural level than surrounding polydeformed, upper greenschist to lower amphibolite facies rocks of the Pacquet Harbour and Ming's Bight Groups. I suggest that the Point Rousse Complex structurally overlies, and thereby masks, underlying rocks of the Fleur de Lys Supergroup.

The Baie Verte Line is also brought around to an east-west trend along the easterly limb of the flexure, and it passes through Ming's Bight and Pacquet Harbour (Figure 7-5). Previously, the easterly extension of the Baie Verte Line was unrecognized mainly because the structural manifestation of this segment of the line is distinctly different from that of the previously known and established portion of the line along the western limb of the flexure (see Baie Verte Line).

The Baie Verte Flexure is thus defined by the arcuate trend of similar rock types that have the same early history involving the Baie Verte Line, the early Paleozoic continent-ocean interface (Williams et al., 1977; Williams and St-Julien, 1978). Subsequent polydeformation is inhomogeneously distributed along the flexure, but also appears to be keyed to this ancient interface. $^{40}Ar/^{39}Ar$ radiometric mineral ages, regional relationships, and structural trends indicate that Taconic polydeformation involved generally east-west compression of rocks along the north-northeast trending parts of the flexure to the west of the Baie Verte Line; Acadian tectonism was accomplished by north-south compression that influenced rocks on the east trending segment of the flexure near the line. It is apparent from this unusual tectonic pattern that the framework of the flexure has controlled tectonic development along the ancient continent-ocean interface since the Taconic Orogeny or earlier. Thus, I interpret the Baie Verte Flexure to reflect the original geometry of the Paleozoic continent-ocean interface (the Baie Verte Line); specifically, the flexure is interpreted to represent a morphological feature of the ancient continental margin. Taconic and Acadian polydeformational events were noncoaxial events that affected different parts of this local irregularity in the margin according to the orientation of maximum compression during each orogenic episode.

MAJOR LATE FAULTS

Two major late faults truncate the geology of the Baie Verte Peninsula to the west and to the south; the Cabot Fault (Wilson, 1962) truncates the Fleur de Lys Belt to the west, whereas the Green Bay Fault (Upadhyay et al., 1971) marks the southerly limit of the Fleur de Lys Belt on the peninsula and offsets the Baie Verte Belt rocks. Previous workers who reviewed the regional fault system in western Newfoundland considered the Green Bay Fault as a splay of the Cabot Fault (Wilson, 1962; Webb, 1969; Belt, 1969; Arthaud and Matte, 1977) because these faults appear to join in the area southwest of Sandy Lake. Both faults are characterized by pronounced physiographic depressions and by the ponding of Carboniferous strata along these depressions.

CABOT FAULT

This fault was first recognized by Murray (Murray and Howley, 1881) as a major fault that traverses western Newfoundland from the Codroy Valley to White Bay. This major tectonic zone was later recognized to extend southward at least as far as Boston, Massachusetts by Wilson (1962), who termed it the Cabot Fault. In the local area, the fault has also been termed the White Bay Fault (Heyl, 1937; Betz, 1943, 1948; Cameron, 1966; Arthaud and Matte, 1977) and the Hampden Fault (Webb, 1969; Belt, 1969). The name Cabot Fault is used in this report.

In the area between Hampden and Sandy Lake, the fault separates the White Bay Group and Wild Cove Pond Igneous Suite from Carboniferous rocks to the west (Figure 1-1); a northern submarine extension of the fault may separate the Granby Island Formation from the White Bay Group (Figure 1-1).

South of Hampden, the fault is marked by local shearing and brecciation of nearby Fleur de Lys rocks and folding and shearing of westerly Carboniferous strata. Its effect on the Wild Cove Pond Igneous Suite was not observed. The actual fault plane is best exposed on Rocky Brook (Figure 1-1), where it is a remarkably abrupt and nearly undisturbed "innocent" appearing zone. Here, moderately to steeply dipping, blackish gray Carboniferous clastic rocks are juxtaposed against amphibolite and nearby granitic gneiss of the Oody Mountain Formation. The fault is marked by less than 1 m of black, shaly gouge. Rocks of the Oody Mountain Formation are slightly sheared and show incipient brecciation near the fault.

The sense of displacement on the Cabot Fault has long been a controversial topic; some workers have considered it to be a major thrust fault (Heyl, 1937; Betz, 1943, 1948; Cameron, 1966), whereas others have considered it as a strike-slip fault (Wilson, 1962; Webb, 1969; Belt, 1969; Arthaud and Matte, 1977). Heyl (1937) and Betz (1943, 1948) interpreted it as a high angle thrust mainly because of the presence of minor high angle thrust faults within Paleozoic strata on the west side of White Bay; Cameron's (1966) evidence for a thrust resolution for the fault was based on the overall geometry of the fault zone from Boston to White Bay. Wilson (1962) used the same line of reasoning as Cameron (1966) for interpreting the fault as a strike-slip structure, and he compared it with the San Andreas Fault. The other workers cited circumstantial evidence for strike-slip movement on the fault, including the vergence of minor structures and the apparent offset of units (that have not been proven equivalent) on either side of the fault.

In the present study, there is no direct evidence for lateral movement on this section of the Cabot Fault, though the juxtaposition of the crystalline Fleur de Lys rocks against only locally deformed Carboniferous sedimentary rocks indicates that the east side of the fault moved upward relative to the west side. The latest movements along the fault postdate Carboniferous deposition.

GREEN BAY FAULT

A major fault in the area of Southwest Arm, Green Bay, was first recognized by Neale et al. (1960) and it was later extrapolated inland to the Birchy Lake area (Neale and Nash, 1963). The fault was subsequently named the Green Bay Fault by Upadhyay et al. (1971) and considered as a part of the Long Range Fault (Arthaud and Matte, 1977). The name Green Bay Fault is used in this report. Carboniferous sediments occur along the fault zone at Indian Pond (Figure 1-1).

The fault is unexposed in the Birchy Lake - Indian Pond area, though locally, as in Tea Bay, Birchy Lake, rocks are sheared along a probable splay of the main fault (Figure 1-1). In this area, it apparently truncates the Wild Cove Pond Igneous Suite, the Baie Verte Line, the Advocate Complex, and the Burlington Granodiorite. It also offsets the Micmac Lake Group. Neale (1962) described the fault further to the northeast where rocks of the Baie Verte Belt are juxtaposed against the Ordovician Lushs Bight Group and Silurian(?) Springdale Group:

> A linear scarp forms the northwest shore of Southwest Arm and extends about 8 miles southwest of Rattling Brook. It is interpreted as a fault line scarp over at least part of its length. Evidence of dislocation was noted along the coast about 2 miles northeast of the mouth of Rattling Brook where a pluton of red granite (to the north) is separated by a zone of cataclasis and shearing from basic volcanic rocks of the Lush's Bight Group (to the south) - which are, incidentally, cut by dykes of the same red granite... At and near the mouth of Rattling Brook westward dipping red siltstone, sandstone and conglomerate of the Springdale Group terminate westward against sheared syenite which appears to grade westward into quartz-feldspar porphyry and flow layered rhyolite. Near the mouth of Corner Brook, basic lavas of the Lush's Bight Group are intensely sheared along lines parallel to the scarp front and quartz-feldspar porphyry along the scarp is cut by shear zones... Beyond this the scarp is ill-defined, outcrops are few, and the fault trace is projected to Shoal Pond only on the basis of rather weakly defined topographic linears. South of the map-area it coincides with Indian Brook valley. Most shear zones associated with this fault dip moderately to steeply southeast. Drag effects and feather faults along subsidiary shears suggest that the west side moved north relative to the east side.

Further to the northeast, the fault appears to extend into Green Bay. The southwestward extension of the fault below Birchy Lake is uncertain. It appears to continue into the Deer Lake Basin (Belt, 1969; Hyde, 1979a,b) and may either merge with the Cabot Fault, as suggsted by some workers (Wilson, 1962; Webb, 1969; Belt, 1969; Arthaud and Matte, 1977), or join the Grand Lake Fault (Hyde and Ware, 1981) which marks the east side of the Carboniferous Deer Lake Basin.

Most workers who have reviewed the regional fault system have agreed with Neale's (1962) interpretation that there has been right-lateral strike-slip movement along this fault (Webb, 1969; Belt, 1969; Upahyay et al., 1971; Marten, 1971; Upadhyay, 1973; Arthaud and Matte, 1977). There is compelling evidence supporting this conclusion. Upadhyay et al. (1971) and Marten (1971) noted that the submarine extension of the Green Bay Fault truncates the Lower Ordovician Snooks Arm Group on the northern limb of a major syncline (Neale, 1957); they suggested that the southern limb of the syncline is represented by the Lower Ordovician Western Arm Group on the Green Bay Peninsula immediately east of Southwest Arm. This relationship indicates a right-lateral displacement of approximately 25 km along the Green Bay Fault. In addition, the Baie Verte Line is truncated by the fault; the nearest established trace of the line is approximately 100 km to the southwest, on Glover Island (Knapp et al., 1979; Knapp, 1980), although ultramafic rocks similar to those along the line on the peninsula occur just southeast of Sandy Lake and at Deer Lake, to the west. These relationships all indicate significant right-lateral movement along the fault; the maximum displacement is 100 km. The latest displacement must also have postdated the depositional age of the Carbonifeous strata on the southeast side of Sandy Lake.

SUMMARY OF TECTONISM ON THE BAIE VERTE PENINSULA

Major facets of the tectonic history of the peninsula are outlined in Table 7-6 and summarized below. The earliest regional tectonism on the peninsula involved deformation, metamorphism and selective migmatization of rocks in the core of the East Pond Metamorphic Suite; structural features in these rocks indicate that they suffered intense tectonism prior to deposition of other rocks in the Fleur de Lys Belt. Based on regional correlation, this event is presumed to be Grenvillian in age (circa 1000 Ma).

Subsequent tectonism on the peninsula was centered upon the Baie Verte Line, and the pattern of tectonism was governed by the original geometry of the line, now reflected in the Baie Verte Flexure.

Initial deformation and metamorphism of supracrustal rocks above the crystalline basement involved the tectonic juxtaposition of the continental Fleur de Lys and oceanic Baie Verte Belts along the early Baie Verte Line. This event involved the westward transport of the Baie Verte ophiolitic rocks over the Fleur de Lys Belt. Based on the distribution of D_E slides and mélange zones and the distribution of later tectonic events on the peninsula, the early form of the line had an arcuate trend defined by the Baie Verte Flexure; this most likely reflects a morphological irregularity in the ancient continental margin. The obduction event is manifested as D_E and D_M structures in the Western Orthotectonic and Transition Blocks and at least D_E structures in the Eastern Orthotectonic Block; D_M structures of the Ordovician units in the eastern part of the Paratectonic Block also probably formed either during or soon after this event. Regional correlation indicates that these events ranged from possibly Tremadocian to Caradocian in age.

The later, D_L, stages of deformation in the Western Orthotectonic Block appear to be related to a release of the D_M stress regime (de Wit, 1972, 1980) and the initiation of easterly directed structures near the Baie Verte Line, a direct reversal in structural polarity from earlier events (Bursnall, 1975). Significantly, MP_M peak metamorphism in this block indicates that maximum physical conditions were attained just prior to this reversal and probable uplift event. Bursnall and de Wit (1975) indicated that D_L structures affected the Transition Blocks. Based on $^{40}Ar/^{39}Ar$ mineral ages, it appears that this event subsided by Middle Silurian (Dallmeyer, 1977) in most of the Western Orthotectonic Block, although younger age dates (Dallmeyer, 1977; Lowdon, 1961) from north of the Little Lobster Harbour Fault indicate that elevated temperatures were maintained somewhat longer in northern parts of the block. These younger dates may reflect an Acadian tectonic influence on this part of the block, which may have originally lain further to the east and was later faulted into its present position along the Little Lobster Harbour Fault.

Acadian tectonic features on the peninsula are distinguished from earlier events by regional stratigraphic relationships and $^{40}Ar/^{39}Ar$ age relationships. They involve

Table 7-6: *Summary of tectonism on the Baie Verte Peninsula.*

OROGENY	Western Orthotectonic Block	Transition Blocks	Eastern Orthotectonic Block	Paratectonic Block		Horse Islands Block	Granby Island Block
				West	East		
Acadian	D_L (?) - Reverse structural polarity and uplift for part of area north of Little Lobster Harbour Fault D_L - Reverse structural polarity and uplift for much of block	D_L (?) - Reverse structural polarity and uplift	D_L - Emplacement of Point Rousse ophiolite over the block D_M - For Siluro-Devonian rocks, emplacement of ultramafic rocks on Scrape Thrust. For older rocks, coaxial with Taconic D_M and tightening of these structures	D_L - Relaxation of D_M D_M - Westerly directed structural polarity; emplacement of the Point Rousse Complex and deformation to the east of the Baie Verte Line	D_L (?) - Possible weak late fabric D_M - For Siluro-Devonian rocks: tightening of Taconic D_M	D_L (?) - Reverse structural polarity and uplift	
Taconic	D_M–D_E: Westward obduction of ophiolites over the Fleur de Lys terrane	D_M–D_E: Emplacement of ophiolites in the block; formation of dynamothermal aureole	D_M (?) - For Ordovician rocks only; related to obduction D_E - Westward obduction of ophiolites over Fleur de Lys terrane		D_M - In Ordovician rocks only; probably related to obduction	D_M (?)–D_E: Westward obduction of ophiolites over the Fleur de Lys terrane ?	D_M (?)–D_E (?): Westward obduction of ophiolites over the terrane ?
Grenvillian		D_B - Gneissic banding and migmatization in parts of East Pond Metamorphic Suite					

overthrusting of the Eastern Orthotectonic Block by the Paratectonic Block along the Scrape Thrust system and accentuating of the Baie Verte Line by the structural juxtaposition of the Western Orthotectonic and Transition Blocks against the Paratectonic Block along the Virginite fault. These events also involved an eastward structural polarity. During the earliest phases of this event, D_M structures of the Eastern Orthotectonic Block and the Siluro-Devonian parts of the eastern Paratectonic Block were formed. This event was probably coaxial with D_M Taconic structures developed in the Ordovician rocks of these blocks and likely involved amplification of these earlier structures. The emplacement of ultramafic rocks against the Pacquet Harbour Group along the immature form of the Scrape Thrust appears to have also occurred at this time. Subsequent Acadian tectonism in easterly areas involved the emplacement of the Point Rousse Complex over the Pacquet Harbour Group along the Scrape Thrust, and the formation of D_L structures and associated metamorphism in the Eastern Orthotectonic Block and easterly Paratectonic Block, as well as the formation of D_M structures in the Point Rousse Complex. Tentative correlation of D_M structures throughout the western part of the Paratectonic Block indicates that D_M structures and the Virginite fault formed essentially synchronously with the easterly events. Later structures in the western Paratectonic Block have been related to relaxation of D_M strain (Kidd, 1974).

The Acadian event on the peninsula may be related to the combined effects of uplift of the Fleur de Lys terrane and the initiation of northward movement of the Baie Verte Belt against the east limb of the Baie Verte Flexure. The latter event is speculatively related to precocious movements along the global scale Variscan shear zone and related minor faults, as outlined by Arthaud and Matte (1977). Formation of the Cabot and Green Bay Faults may be related to this event and further displacement along these faults may have occurred during later development of this megashear system (Arthaud and Matte, 1977). These faults helped to form major Carboniferous basins found along their traces.

The timing of deformation and metamorphism on the Granby Island and Horse Islands Blocks is uncertain due to the insular nature of these areas. The development of regional metamorphism on Granby Island suggests that it is related to tectonics in the nearby Western Orthotectonic Block, though specific correlations are not possible. Tectonism on the Horse Islands is very similar to that in the rest of the Fleur de Lys terrane, although it is uncertain whether polytectonic events there were dominantly Taconic as in the Western Orthotectonic Block, or mainly Acadian as in the Eastern Orthotectonic Block.

Workers at the old Tilt Cove Mine on the Baie Verte Peninsula were featured on this stamp, the first ever issued to commemorate mining.

CHAPTER VIII

MINERAL DEPOSITS

INTRODUCTION

Mineral deposits have long been a major attraction on the Baie Verte Peninsula. Over 2000 years ago, the Dorset people were most likely lured to the peninsula by a soapstone deposit at Fleur de Lys (Chris Nagle, personal communication, 1981) (see Frontispiece); presumably, they carved the soapstone into oil lamps. In more recent history, the peninsula has been a center for exploration and exploitation of both metallic and industrial mineral deposits. Since 1864, the area has been host to nine producing mines, eight of which have been worked for base and precious metals and the ninth for asbestos. As of mid-1983, only the asbestos mine is active (operated by Baie Verte Mines Incorporated, formerly Advocate Mines Limited). However, mining remains a major factor in the economy of the Baie Verte Peninsula. Considering the number of ore bodies already discovered on the peninsula and the numerous prospecting environments present, the potential for the discovery of future ore bodies is promising.

Almost all of the deposits on the peninsula have been previously described in various company, government and university reports; very little new work was done on them, directly, in this study. However, this report serves as a compendium of the previous work. In addition, the geology and geochemistry outlined in this report help to clarify the environments of formation of the deposits.

The tectonic juxtaposition of the Fleur de Lys and Baie Verte Belts concentrated many environments favorable to mineralization in the area of the peninsula. These environments correspond roughly to the major stratigraphic divisions outlined in this report, including the ophiolitic rocks, the volcanic cover, the Fleur de Lys Supergroup, and the intrusive rocks. They form the framework for the following discussion.

In this chapter, the following terms are used to describe the status of a deposit:

indication - deposit upon which no known development work has been done.

showing - deposit upon which some development work has been undertaken, though insufficient to provide an estimate of its spatial dimensions.

prospect - deposit upon which enough development work has been done to make a reasonable estimate of its extent.

All mineral abbreviations used in the tables in this chapter are listed in Appendix I.

OPHIOLITES AND CLOSELY ASSOCIATED ROCKS

All of the mines on the Baie Verte Peninsula and many of the major showings are within the ophiolitic rocks and in closely associated rocks of the Point Rousse and Advocate Complexes and the Pacquet Harbour Group. Both metallic and industrial mineral deposits occur in these rocks. The style of mineralization varies between the different ophiolites, so they are described in stratigraphic order for both metallic and industrial deposits; following the descriptions, they are compared and summarized.

METALLIC MINERAL DEPOSITS

Betts Cove Complex

The Betts Cove Complex is the best preserved ophiolite on the peninsula and has been the most prolific producer of metallic minerals. It is host to two major sulfide deposits that were formerly mined at Tilt Cove and at Betts Cove, as well as numerous other metallic prospects and showings. Both of the mines are located near the pillow lava - sheeted dike interface of the complex.

TILT COVE MINES

LOCATION AND PREVIOUS STUDIES: The abandoned mines are located in the community of Tilt Cove on the southeast corner of the Cape St. John Peninsula. The community is accessible from the La Scie highway by approximately 6 km of gravel road and by boat. The harbor at Tilt Cove is only a small indentation in the rugged Notre Dame Bay coastline, but it is oriented such that only strong southerly winds affect it. The depth of water at the pier in Tilt Cove is 7 to 12 m.

Numerous reports, published and unpublished, describe many different facets of the Tilt Cove Mines; the following description is distilled primarily from the work of Snelgrove (1931), Douglas et al. (1940), Donaghue et al. (1959), Papezik (1964), Craig (1967), Upadhyay (1973), Upadhyay and Strong (1973), and Squires (1981). In addition, part of the historical section is from Martin (1983).

HISTORY OF DEVELOPMENT: Massive sulfide ore was first recognized at Tilt Cove in 1857 by Smith McKay, a surveyor. It is generally believed that McKay met a Tilt Cove fisherman, Isaac Winser, who had either used ore from the cove for boat ballast or had a sample in his home; by the way of this encounter, McKay discovered the copper ore. McKay went into partnership with C.F. Bennett, a St. John's financier, and they formed the Union Mining Company. They opened the West or Union Mine in Tilt Cove on July 27, 1864; Bennett later gained sole ownership of the workings. During the period 1869-1876, a small nickel ore body was discovered in the West Mine; this ore, like the copper ore, was shipped to Swansea, Wales, for smelting.

The East Mine deposit was discovered in 1886; Bennett died before it was worked. In 1888, the property was leased to the Tilt Cove Copper Company Limited. The company installed smelters at Tilt Cove but soon encountered insurmountable difficulties: the East Mine surface facilities burned down in 1890 and copper prices dropped dramatically. In addition, the smelters left the local countryside virtually devoid of vegetation and were the ultimate cause of local fires in the townsite. Unable to continue the operation, the company, through its principal director, J. Taylor, subleased the mine in 1890 to the Cape Copper Company, which also controlled the Swansea smelters.

The Tilt Cove Mines were revived by the Cape Copper Company; they resumed shipments to Swansea for smelting and started new ore shipments to New York. The prosperity of the company at this time was due mainly to (i) new advances in pyrite smelting, allowing sulfur to be retrieved from the ore, (ii) the discovery of gold and silver in the ore, and (iii) the abolition of tariffs on ore by the United States. These favorable conditions also allowed the company to rework the West Mine dumps, while operations continued in both the East and West Mines. Underground operations ceased at the West Mine in 1902, though they resumed during the period of 1907-1911 as opencut workings; during this time, the reserves at the East Mine were severely depleted. Mining dwindled under the Cape Copper Company until September, 1913, when the company cancelled its option with the Tilt Cove Copper Company. The latter company was liquidated in 1914; its holdings were subsequently taken over by local Notre Dame Bay merchants, of whom Messrs. R.G. Rendell and J.M. Jackman were the most prominent. They formed the Tilt Cove Mining Company, which flourished briefly due to high copper prices during World War I, but later collapsed. During the war, freight rates rose to such a level that the company had difficulty shipping ore to the smelters. In addition, the owners installed an expensive concentrating plant, which operated for only a few days before shutting down due to the miserably poor results. The property was returned to its trustees in 1916 or 1917.

During the period from 1917 to 1954, several concerns showed interest in the mines, though no new reserves were discovered. In 1954, the Maritime Mining Corporation Limited (renamed First Maritime Mining Corporation Limited in 1964), of which the most prominent backer was Dr. M.J. Boylen, dewatered the old workings and renewed development of the area. By 1955, it had located several new, large ore bodies in both the East and West Mines by diamond drilling, and started production in August, 1957. Ironically, these new bodies were very close to the old workings, as noted by Kall (1968):

> The old-timers' last attempt to find new ore was made in the East Mine in 1911, where a 1,350-foot tunnel, driven on the sea-level, failed to encounter mineralization.
>
> Exploration work done in 1955 and 1956 showed that if the last 900 feet of the drift had continued on its original course, Tilt Cove would never have become a ghost town. However, for some unknown reason, the original course of the tunnel was changed by 30 degrees after the first 400 feet were completed, and thus the drift missed a huge orebody, which, in 1956, became the lifespark of Tilt Cove. A little later, subsequent discoveries in the West Mine showed that here as well, the oldtimers had missed large orebodies by the hair of their chins.

These deposits were exhausted in June, 1967, and operations ceased. At the time of this writing, the property is being explored again by Newmont Exploration of Canada Limited.

LOCAL GEOLOGY: In the Tilt Cove area, the Betts Cove Complex is flanked to the south by the Bobby Cove Formation of the Snooks Arm Group and to the north by the Cape St. John Group; the Ordovician units are intruded by sills and dikes of the Cape Brulé porphyry (Figure 8-1) (see Chapter V). These units are all disposed on the steep to slightly overturned northern limb of the gently east plunging regional syncline in the area (Neale, 1957). The detailed structure of the area is complex and characterized by numerous faults and drag folds of many generations; the youngest faults offset the porphyry intrusions.

The Betts Cove Complex has been greatly modified by the faulting. The ultramafic member ranges from 600 m wide in the area just west of Windsor Lake to less than 100 m wide in the area east of Beaver Cove Pond. It is generally highly serpentinized, steatitized, and carbonatized in the mine area (Upadhyay, 1973). The ultramafic member appears to be faulted against all other members of the complex in the Tilt Cove area. The gabbroic and sheeted dike members are poorly represented at Tilt Cove; the gabbro attains a maximum outcrop width of approximately 300 m southeast of Windsor Lake, and appears to be in gradational contact with the sheeted dikes (Figure 8-1). The dikes are mostly brecciated (Upadhyay, 1973). Squires (1981) suggested that the contact between the dikes and the overlying pillow lava member is gradational in the area south of Windsor Lake (Figure 8-1). The pillow lava member attains a maximum width of approximately 400 m south of Windsor Lake, and its outcrop pattern to the east and west is highly modified by faulting; a thin sliver of the member outcrops to the north of the ultramafic member between Windsor Lake and Beaver Cove Pond. The most significant mineralization in the Tilt Cove area is confined to the pillow lava member. Based on geochemical analyses of the pillow lavas in the fault block between Mud Pond and Beaver Cove Pond (Figure 8-1) (Craig, 1967), it appears that both high magnesian and tholeiitic lavas are present.

The Betts Cove Complex is conformably overlain to the south by sediments and volcaniclastics of the Bobby Cove Formation and both of these units are unconformably overlain by the Cape St. John Group (Strong, 1980; Squires, 1981). The Cape St. John Group in the area contains dominantly subaerial pyroclastics, sandstone, and subaerial mafic lavas. Locally, such as west of Windsor Lake, it appears to be gradational with the Cape Brulé porphyry, which intrudes both the ophiolite and the Bobby Cove Formation nearby. The porphyry has been affected by a set of north trending steep faults (Figure 8-1), the most prominent and important of which is the Valley Fault, between Windsor Lake and Tilt Cove; this fault is an important factor in the present distribution of major ore bodies in the area (Squires, 1981).

MINERALIZATION: The Tilt Cove deposits are massive as well as stockwork sulfide bodies. Minor nickel mineralization is present, though unrelated to the sulfides. The sulfide deposits are largely confined to the sheeted dike and pillow lava members of the complex, but minor mineralization is found at one locality in the ultramafic member. The nickel deposits are close to, but not directly related to, the main sulfide mineralization. All of the ore bodies have undergone tectonism and local remobilization (Squires, 1981).

The main constituents of the Tilt Cove sulfide ores are pyrite, chalcopyrite and magnetite with minor sphalerite and pyrrhotite and local concentrations of silver and gold (Donaghue et al., 1959). Pyrite is found in massive, stockwork and disseminated modes. Chalcopyrite is typically associated with the pyrite and is locally altered to covellite. Magnetite is commonly disseminated throughout the ore, and in places forms patchy masses. Locally, hematite is found in the ore along fault zones (Squires, 1981). The sphalerite and pyrrhotite are found only in small amounts and the native silver occurs in the form of wire (Donaghue et al., 1959); the gold locally occurs in gold-silver alloys (Donaghue et al., 1959). In addition, Donaghue et al. (1959) noted dendritic encrustations of native copper on clasts in glacial till near the mine workings.

Squires (1981) summarized the distribution of ore zones in the sheeted dike and pillow lava members at Tilt Cove. He distinguished two mining areas, the West and East Mines, on opposite sides of the Valley Fault. The West Mine is composed of the West Zone and the Cove Zone, whereas the Cliff Zone, "A" Zone, and Main Zone constitute the East Mine. The distribution and inter-relationships of these bodies depicted in Figures 8-1 and 8-2 and the following brief descriptions are from Squires (1981); more comprehensive descriptions of the ore bodies were given by Bichan (1958).

The West Mine Zone is a collective term here used to group the similar stockwork sulphide zones that were individually named by mine geologists on the basis of varying ore grade and stoping levels. It is stratigraphically situated at the base of the ophiolitic Pillow Lava Member as is indicated by close proximity to a small sheeted dike breccia unit (Upadhyay, 1973). The zone is a north-striking moderately to steeply east-dipping disseminated stockwork type deposit....

The West Mine Zone is succeeded to the east and at depth by the Cove Zone. Very little information is available on this deposit except that it is a massive, north-striking body which is terminated to the east by the (inferred moderately east-dipping) Valley Fault.

East of the Valley Fault, the "A" Zone is first encountered. This zone is briefly described as a low grade stockwork zone which underlies both the Cliff Zone and the Main Zone, and is separated from the Main Zone by a fault....

The Cliff Zone is here interpreted as a dislocated basal portion of the Main Zone massive sulphides. Evidence for this is given by the report that at least part of the Cliff Zone is definitely separated from the Main Zone by low angle faults (Bichan, 1958). This zone is also north-striking and is at a relatively shallow depth.

The "Main Zone" is another collective term which groups a number of related massive ore bodies. Some of these bodies are the North, Main and South Lodes, and Zone 26, and they form a north-striking, steeply east-dipping massive unit. These bodies are significantly faulted and a quartz porphyry dike separates the Main and South Lodes.... Unfortunately, no detailed information is available to accurately delineate these structures. Conceptually important is the observation that the Main Zone overlies the "A" Zone sulphides.

It is interpreted here that the stockwork and massive bodies of the East Mine area are a disjointed continuation of the respective West Mine area bodies which are separated by the Valley Fault.

In addition to these features, Squires (1981) noted that the massive sulfides of the Cliff Zone contain significant quantities of quartz and that ore fragments from the Main Zone open pits display sedimentary layering. He also outlined the mineral paragenesis for ore samples from the Main and Cliff Zones. The reader is referred to his work for details. The geochemical affinity of the lavas associated with the mineralization is uncertain, though Craig (1967) documented the presence of both tholeiitic and high magnesian lavas in the area of Mud Pond, less than 1 km east of the Main Zone; this suggests that both lava types are present in the mineralized areas.

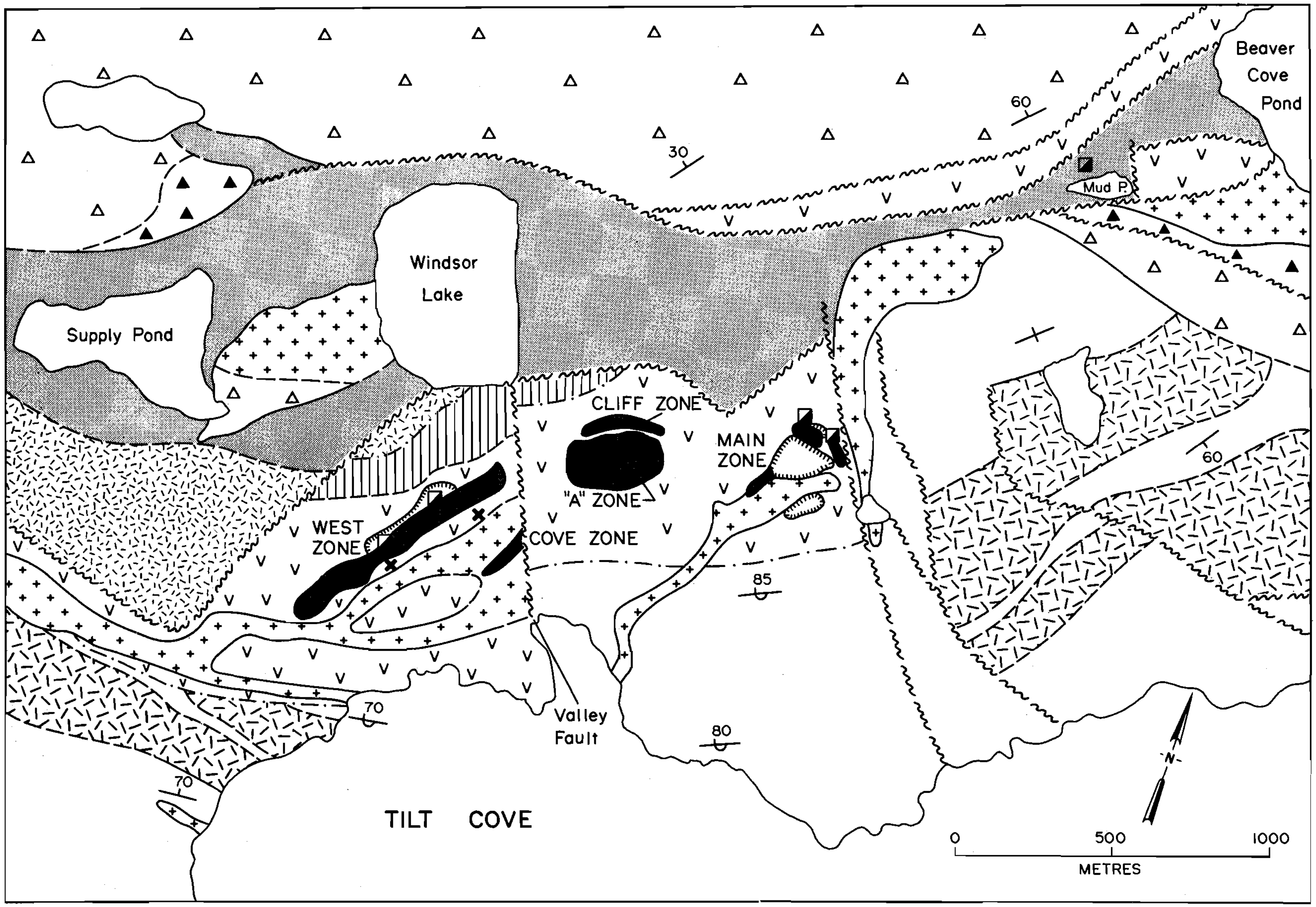
One sulfide occurrence is in the ultramafic member of the complex at Mud Pond. The sulfides form disjointed veins and lenses that were tectonized with the host ultramafic rock (Squires, 1981). Squires (1981) indicated the presence of two vein types, one mainly pyrite, and the other dominantly chalcopyrite; the two types display highly contrasting mineral parageneses. The reader is referred to Squires' work for details.

Nickel mineralization occurs mainly near the contact of the pillow lava and a subsurface fault sliver of talc-carbonate rock near the West Zone at about the 1000 level, but also is found in calcite veins within the quartz-feldspar porphyry near the West Zone (Figure 8-1) (Papezik, 1964). The main deposit is inaccessible at present, though on the basis of hand samples Papezik (1964) speculated that the nickel minerals are in massive pockets and disseminated grains in the host rocks. E. Sampson of Princeton University first identified the following minerals in the main deposit: niccolite, maucherite, chloanthite, gersdorffite, arsenopyrite and millerite (reported in Snelgrove, 1931). Subsequently, Snelgrove (1931) reported annabergite and erythrite as secondary minerals. Papezik (1964) identified violarite in the main body and either rammelsbergite or pararammelsbergite from the calcite veins.

ORIGIN OF THE ORES: In 1865, Murray (reported in Murray and Howley, 1881) visited the Tilt Cove Mines and concluded that the ore deposits were strata-bound and were in the same position relative to the ultramafic rocks throughout the present Betts Cove Complex. Snelgrove (1931) confirmed that the ore at Tilt Cove occupied a particular stratigraphic level, but considered the ore deposits as replacements and veins related to nearby quartz diorites (gabbro member of present complex). He attributed their consistent stratigraphic level to their distance from the diorites, which trend parallel to the local stratigraphy. A.N. Rove (Douglas et al., 1940) speculated that the ore bodies were deposited by mineralizing fluids at essentially the time that quartz porphyry dikes intruded the local sequence; however, he considered the mineralization to be derived from nearby mafic-ultramafic rocks rather than the porphyry.

More recently, Upadhyay and Strong (1973) suggested that the Tilt Cove ore bodies are stratigraphically controlled and formed at an oceanic spreading center. They likened the Tilt Cove and Betts Cove deposits to similar deposits in oceanic crust at Cyprus (see Betts Cove Mine below). Squires' (1981) work supported this concept; by correlating ore bodies across the Valley Fault, he showed that the deposits occur near or at the base of the pillow lava member, and that the massive ore overlies the stockwork ore, similar to the relationship at Cyprus. In addition, he reported that the ore is locally bedded. His discovery of a distinct quartz-bearing ore at Tilt Cove provides a further link to the Cyprus deposits, as a distinct quartz-bearing "B Zone" ore is also found in the latter rocks.

On the basis of the local geology, ore relationships, and ore mineral paragenesis, Squires (1981) erected a geological history for the sulfide mineralization in the complex which is shown in Table 8-1.


Beaver
Cove
Pond
Mud P.
Windsor
Lake
Supply Pond
CLIFF ZONE
MAIN
ZONE
"A" ZONE
WEST
ZONE
COVE ZONE
Valley
Fault
TILT COVE
60
30
60
85
70
80
70
0
500
1000
METRES
N

Cape St. John Group

Mafic volcanic flows

Unseparated felsic volcanic and volcaniclastic rocks; minor mafic flows

Cape Brulé Porphyry

Snooks Arm Group

Diabase dikes

Bobby Cove Formation: *mainly volcaniclastic rocks*

Betts Cove Complex

Pillow lava member

Sheeted dike members

Gabbro member

Ultramafic member

Approximate surface projection of ore bodies

Approximate surface projection of nickel deposits

Shaft

Open pit

Geological contact (defined, assumed, gradational)

Bedding (inclined, vertical, overturned)

Figure 8-1: *Geology of the Tilt Cove area (after Squires, 1981).*

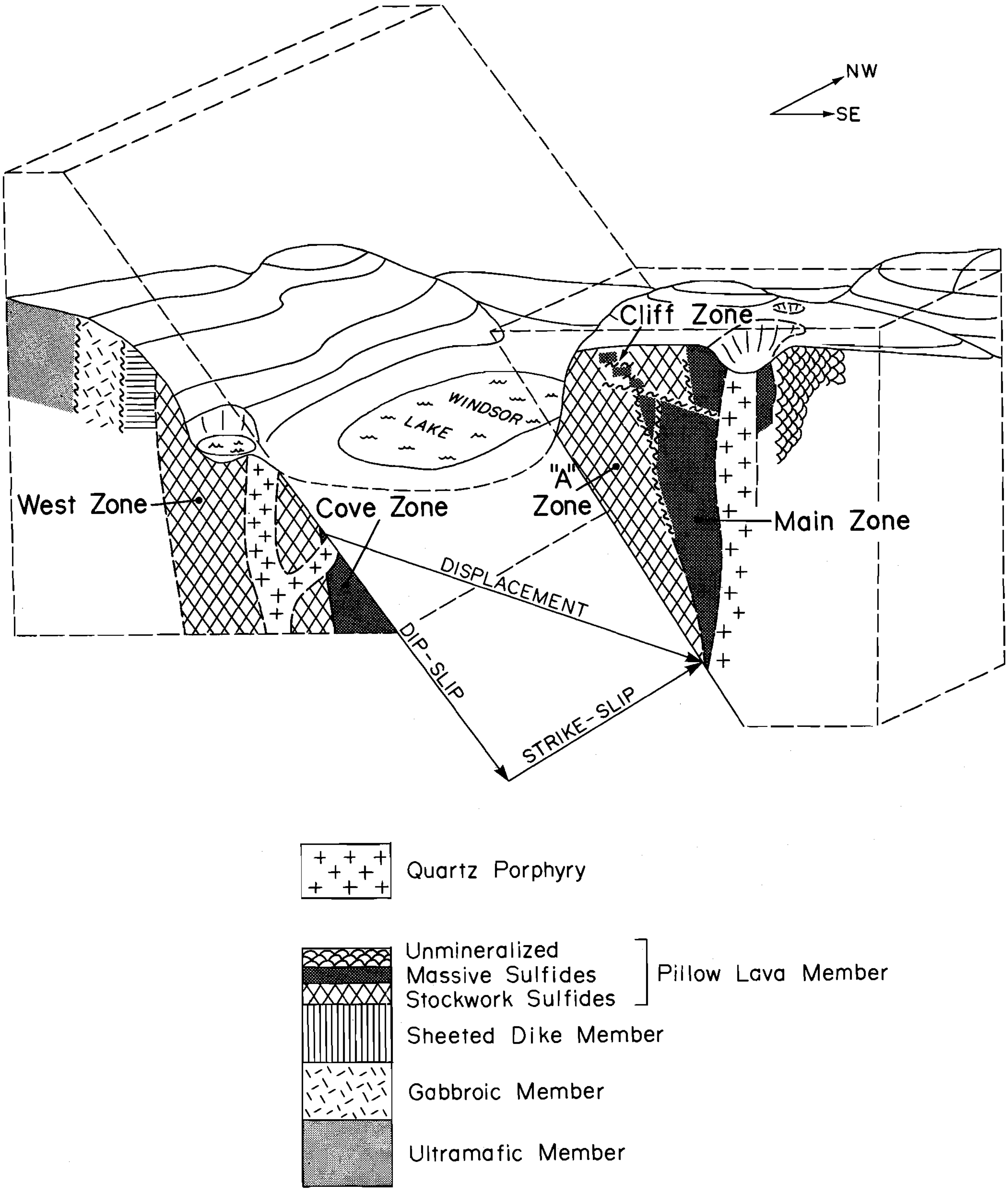


Figure 8-2: *Relationships of ore bodies to the Valley Fault (after Squires, 1981).*

Table 8-1: *History of sulfide mineralization at Tilt Cove (Squires, 1981).*

Period	ORDOVICIAN Early	ORDOVICIAN Middle	SILURIAN	SILURIAN	DEVONIAN
Event	Constructive plate volcanism	Taconic Orogeny, ophiolite vertical; pillow lavas and ultramafics juxtaposed	Felsic volcanism; ophiolite exposed at surface	Volcanism stops; thick cover of Silurian strata	Acadian Orogeny; faulting
Pillow Lava Member	Cyprus-type sulfide deposition	Remobilization of sulfides into faults	Oxidation of sulfides; hematite formed	Fluids less oxidized; chalcopyrite stable	Sulfide remobilization along faults
Ultramafic Member		Introduction of sulfides via faults and shears	Oxidation of sulfides; hematite formed	Fluids less oxidized; chalcopyrite stable; hematite pseudomorphed to magnetite	Sulfide remobilization along faults

Papezik (1964) discussed the possible origin of nickel mineralization at Tilt Cove:

> The source of the nickel in the Tilt Cove deposit was undoubtedly the peridotite; the copper and iron sulphides are of a different origin. The nickel was probably re-mobilized, transported for a short distance and deposited during subsequent folding and fracturing. This movement may have been facilitated by the intrusion of later granitic rocks, the "quartz porphyry". The nickel-bearing veins in the quartz porphyry indicate that some mobility persisted even after the crystallization of at least part of the intrusive rock.

PRODUCTION AND FUTURE POTENTIAL: DeGrace et al. (1975) reported the following production figures for the Tilt Cove Mines:

> The Tilt Cove Mine opened in 1864 and operated continuously until 1917. It is reported to have produced 1,491,136 tons of copper ore, 78,015 tons of regulus (matte) and 5,416 tons of copper ingots. The total amount of copper recovered from these products was estimated at over 61,000 short tons. The West mine contained 8 to 12 percent copper and the East mine about 4 percent. Some gold, silver and much sulphur were recovered at Swansea, Wales, where most of the ore was shipped for smelting.
>
> In 1954 First Maritime Mining Corporation Limited commenced development and rehabilitation of the old workings at the Tilt Cove mine. Regular production, at 2,000 tons of copper ore a day, commenced in 1957. Operations ceased in 1967 when all ore was exhausted. Total production over the ten-year period amounted to 183,597,125 pounds of copper and 42,425 ounces of gold from approximately 7,400,000 tons of ore.

In addition, between 1869 and 1876, the West Mine produced 411 tons of nickel ore, which contained up to 24% nickel (Snelgrove and Baird, 1953).

There are no known reserves at Tilt Cove; Newmont Exploration of Canada is actively searching for new ore bodies.

BETTS COVE MINE

LOCATION AND PREVIOUS STUDIES: The Betts Cove Mine is situated on the west summit of Mine Hill approximately one kilometre west of the harbor of Betts Cove; it is approximately 16 km southwest of Tilt Cove. The nearest community is Nippers Harbour, just over 3 km to the southwest.

The mine has been mentioned in many previous works. The following description is taken largely from Douglas et al. (1940), Upadhyay and Strong (1973) and Martin (1983).

HISTORY OF DEVELOPMENT: The early work of Robert Knight, deputy surveyor for Newfoundland, first brought attention to the mineral potential of the Betts Cove area in the 1860's. Baron Francis von Ellershausen organized the Betts Cove Mining Company in 1864, and mining commenced on the sulfide deposit at Betts Cove in 1875. Six shafts were dug to reach the ore and, within one year of opening, smelters were installed at the site. In 1880, the Betts Cove Mining Company sold the mine to Newfoundland Consolidated Copper Company Limited, a New York - based concern that also had controlling shares in the Little Bay Mine across Notre Dame Bay to the southeast of Betts Cove. In 1883, a landslide devastated the buildings and machinery at the mine site and collapsed the mine roof, thus forcing the closure of the mine. The final ore shipment was made in 1886.

Since the turn of the century, work has been carried out intermittently at the mine site, without any new developments, by the following companies: Newfoundland Copper Concentrating Company (1900), Newfoundland Exploration Syndicate (1905), Pilleys Island Pyrites Company (1906), Falconbridge Nickel Mines (1953-54), Advocate Mines Limited (1955, 1966), Consolidated Morrison (1974), and Consolidated Rambler Mines (1975).

LOCAL GEOLOGY: The most complete section of the Betts Cove Complex is exposed in the Betts Cove area (Upadhyay et al., 1971; Upadhyay, 1973; Coish, 1977a). The local ophiolitic succession, depicted in Figure 8-3, was briefly summarized by Upadhyay and Strong (1973):

> The ultramafic rocks are pyroxenite, harzburgite and dunite. These lithologies are regularly interlayered on all scales from several millimeters to 1.5 meters, and show pronounced cumulus textures.

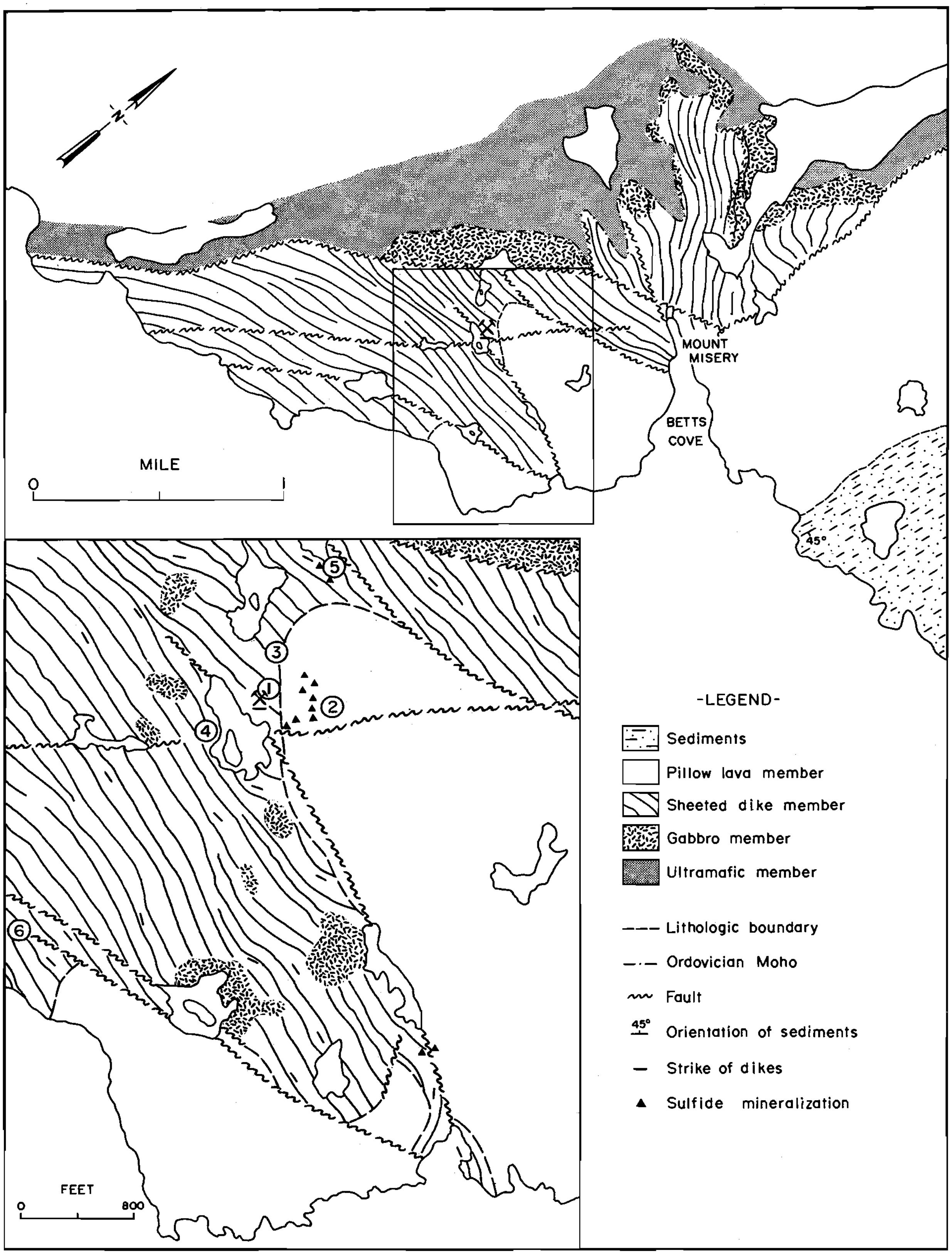


Figure 8-3: *Geology of the Betts Cove area (after Upadhyay and Strong, 1973). Circled numbers refer to mineral occurrences described by Snelgrove (1929) as quoted in text.*

The upper contact of the ultramafic member is extremely variable in character but normally consists of a complex transition zone of interlayered pyroxenite and gabbro which grades upward into gabbro. The gabbro member is regionally discontinuous and cut by abundant basic dikes which increase in number toward the top, grading into the sheeted dike unit which consists of 100 percent dikes. Where gabbro is absent the sheeted dikes die out rapidly downward into the ultramafic member.

The overlying pillow lavas are generally faulted against the sheeted complex, but in some cases there is a rapid but continuous upward decrease in number of dikes to pillow lavas. The pillow lavas also contain pillow breccias, aquagene tuffs and abundant interstitial red chert. The sedimentary units include immature quartz-free volcaniclastic sediments, pyroxene-andesite tuff and agglomerate, graywacke, minor argillite, and red and green chert.

MINERALIZATION: Upadhyay and Strong (1973) summarized the sulfide mineralization at the mine:

In the Betts Cove area sulfide mineralization occurs at a number of places within the sheeted dikes and pillow lavas; sporadic pyrite and minor chalcopyrite are particularly common in sheeted dikes. Pyrite is sparsely disseminated throughout the gabbro, sheeted dikes, and pillow lavas. Detailed mapping of the Betts Cove mine and Mount Misery areas shows that the dominant concentration of sulfides occurs along or near the contact between sheeted dikes and overlying pillow lavas...

The following evidence is taken as indicating that the chloritized fault zones which carry many of the sulfide concentrations postdate the mineralization, i.e., they did not act as 'controls'' for sulfide deposition. The sulfides occur in all forms ranging from banded massive lenses to disseminated zones. The largest sulfide bodies occur within chloritized fault zones, most of which lie close to the sheeted dikes/pillow lava contact. Ore samples from such parts are strongly foliated.

Mineralization in pillow lavas shows maximum concentration in the spaces between pillows although it is disseminated through them. The sulfide-rich rims are now foliated, producing an augen structure around the massive pillows. The concentration of sulfides between pillows is taken as a primary feature which was further accentuated by deformation, with the interstices of the pillows being more susceptible to shearing and remobilization than the interiors of the massive pillows.

Massive and banded sulfides are best seen in material from the abandoned mine dump, where numerous blocks show sedimentary lamination and slump folds which are indicative of initial sedimentary deposition of sulfides. These were later modified by deformation and remobilization as suggested by veins and stringers of sulfides within the associated silicate minerals.

They have also reported a simple mineralogy for the Betts Cove ores, consisting primarily of pyrite, with subordinate chalcopyrite and local concentrations of sphalerite.

The surface expression of the main Betts Cove body is reportedly meager and the shape of the main deposit and the form of the mine workings are only poorly known (Douglas et al., 1940). Snelgrove (1929) noted that the massive deposits were associated with zones of chlorite schist, and briefly described six of these zones (Figure 8-3).

Zone 1: This is the most southerly of the three zones on and near Mine Hill and is evidently the main zone. Its strike is South 63 degrees West (true), and its dip 65 degrees North. Its most westerly limit is covered with talus from the cave-in, but it is presumably intersected by the main shaft, No. 1. At the south side of Mine Hill it is exposed with widths up to about 40 feet, and contains ore shoots, largely pyritic, up to 24 feet wide. Here it is opened upon by the inclined shafts, numbers 3 and 4, down the plane of schistosity, and also by two test pits, numbers 3 and 4, further east, where the zone swings abruptly to the southeast and peters out. Two tunnels also top this zone.

Zone 2: This is also on the south side of Mine Hill and is developed by a 20 ft. tunnel at the east along a 6 ft. pyritic zone at the intersection of conjugate fractures. It is also exposed along the cave-in fault (dipping 75 degrees to 80 degrees south) with a dip of about 45 degrees north. Here the outcrop of the zone follows closely the 500 ft. contour and swings north around the west side of Mine Hill, with a dip of 55 degrees East. At pit 5, it again swings about east-west (true), with a dip of 75 degrees North. Pyritic shoots up to 20 ft. wide are exposed along the outcrop of this zone.

Zone 3: to the northwest of Mine Hill, is developed by a pit, No. 6, in about 8 ft. of pyritic ore, which is cut by a basaltic dike, apparently of post-mineralization age, since it is unmineralized and does not share the schistosity of the zone.

Zone 4: which is poorly defined, exists at the south shore of the pond south of Mine Hill, where pits numbers 1 and 2 are located, along an east-west (true) strike.

Zone 5: one-quarter of a mile north of Mine Hill, is in lava in which remnants of pillow structure could not be detected - unlike all other zones. This zone also strikes about east-west and evidently dips steeply. The mineralization exposed is highly disseminated pyrite and chalcopyrite.

Zone 6: a shear slightly over 5 ft. wide, in chlorite schist altered from pillow lava, occurs about 3000 ft. south of Mine Hill and strikes about east-west (true) with a dip of 70 degrees North.

ORIGIN OF THE ORE: Upadhyay and Strong (1973) noted that the stratigraphic evidence for the Betts Cove ore being concentrated at the sheeted dike - pillow lava interface

... contradicts the suggestion of earlier workers that the early dioritic-gabbroic (Snelgrove, 1931; Douglas et al., 1940) and granitic (Baird, 1951) intrusions caused an epigenetic hydrothermal mineralization localized along pre-existing faults. Furthermore, since the gabbro rocks occur as screens within the sheeted dikes (i.e., they do not intrude the dikes), they could not have been responsible for any epigenetic mineralization.

They suggested the following model for the genesis of the Betts Cove deposits (Figure 8-4):

(1) Lower Ordovician sea-floor spreading resulting in production of the ophiolite (oceanic crust) complex [Figure 8-4a], accompanied in the very earliest stages by volcanic exhalative mineralization to produce typical massive sulfides underlain by disseminated "stockwork" mineralization [Figure 8-4b].

(2) Faulting and shearing at a later stage resulted in the remobilization of the sulfides and their location in chloritic fault zones [Figure 8-4c].

The mineralization resulted primarily from syngenetic precipitation of sulfides from upwelling geothermal brines during hiatuses in volcanism in an oceanic ridge-type environment with the less important disseminated or "stockwork" mineralization produced by replacement and cavity-filling in the underlying rocks during ascent of the volcanic exhalations.

PRODUCTION AND RESERVES: The production and reserves of the mine were summarized by Snelgrove and Baird (1953):

Between 1875 and 1878 Betts Cove mine shipped 95,999 tons of ore. The total production was 130,682 tons of ore and regulus in addition to 2,450 tons of iron pyrite. During the working of the mine the ore was handpicked at the surface and some was jigged before being shipped. The average grade of the shipped ore is not definitely known, but is said to have averaged nearly 10 percent copper.

No estimates of reserves are available. Three ore dumps remain on the property, the tonnage of which is unknown but is much smaller than those of Tilt Cove. Two of these dumps appear to contain material rejected because of high zinc content. Grab samples from these yielded 4.14 percent zinc, and 0.39 to 1.83 percent copper. Precious metals are present to the amount of 6 dwts of gold and 18 dwts of silver, both per long ton. The third dump was similarly sampled and yielded 9.42 percent copper with only traces of zinc and gold.

It is interesting to note that the reserves are uncertain because the inner reaches of the mine were inaccessible following the mine collapse of 1883. Snelgrove (1929) quoted an old Newfoundland Survey report in which it was reported, by the Newfoundland Exploration Syndicate, that "considerable ore still existed in the first and second levels and

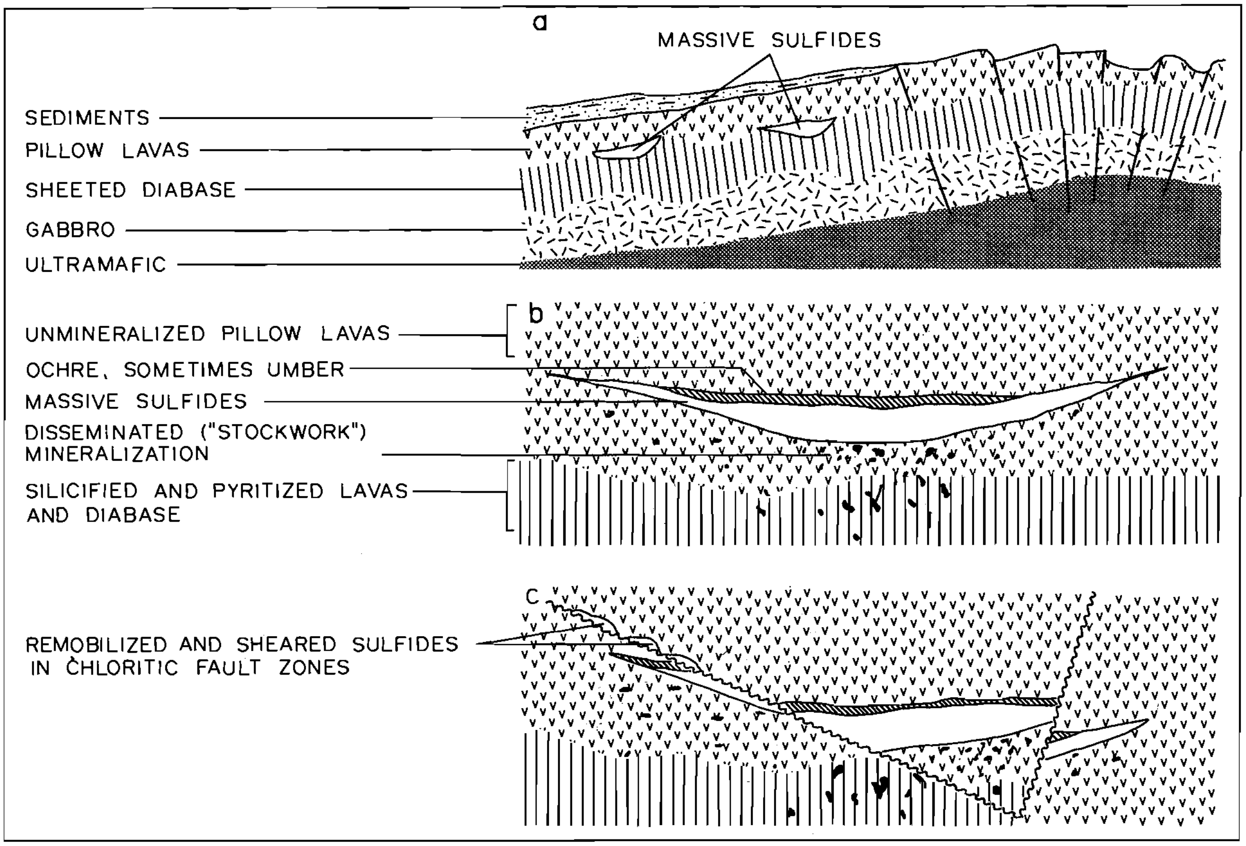


Figure 8-4: *Stages of formation and concentration of the Betts Cove ore (from Upadhyay and Strong, 1973).*

it is generally believed that... bodies of ore occur below the lowest workings." Subsequently, drilling was carried out in the old workings by Falconbridge in 1952 and 1953 and by Advocate in 1955; reportedly, mineralization was encountered, but it was not of economic grade.

OTHER METALLIC MINERAL OCCURRENCES

The Betts Cove Complex hosts a multitude of smaller metallic mineral deposits; these are presented in Table 8-2 and located in Figure 8-5. (Note that only a few of the major occurrences are shown on Figure 1-1). The table is based solely on data from the Mineral Occurrence Data System of the Department of Mines and Energy. Selected references are given for each occurrence.

Most of the occurrences listed in Table 8-2 have characteristics that are typical of the Betts Cove deposit and, hence, are termed Betts Cove type occurrences (Dean, 1977). In particular, these deposits are within the sheeted dike - pillow lava members of the complex and generally display a simple py-cp-po mineralogy. Locally, such as at the Long Pond Prospect, galena is included in these deposits.

Two other types of deposits are found in the complex, including vein type and lead mineralization in the ultramafic member. The vein type deposits (numbers 23 and 24 on Table 8-2) are typified by cp-gn mineralization in quartz veins of uncertain age. The source of the mineralization is likewise uncertain. Similarly, the local lead mineralization in the ultramafic member (numbers 25 and 26 on Table 8-2) is of uncertain origin.

SUMMARY OF METALLIC MINERALIZATION

Almost all of the metallic mineral deposits of the Betts Cove Complex are characterized by the following features and have been termed Betts Cove type deposits (Dean, 1977):

(i) they contain simple py - cp - po mineralogy;

(ii) host rocks are either sheeted dike or pillow lava, in many cases at the contacts of these two members (locally, mineralization in gabbro member);

(iii) mineralization occurs as disseminated stockwork and as massive strata-bound lenses, and is interpreted to be volcanogenic in origin;

(iv) in many cases, zones of mineralization are apparently weak zones, susceptible to later, intense deformation.

These deposits are comparable to Cyprus-type and other ocean ridge generated deposits (Upadhyay and Strong, 1973; Squires, 1981).

Locally within the complex, chalcopyrite-galena mineralization occurs in quartz veins and galena mineralization occurs in the ultramafic member; the origins of these

Table 8-2: *Betts Cove Complex metallic mineral occurrences. (Number refers to Figure 8-5. For mineral abbreviations see Appendix I.)*

OCCURRENCE	STATUS	HOST ROCK	MINERALOGY	TYPE	COMMENTS	SELECTED REFERENCES
1. Nippers Harbour Prospect	prospect	chloritized, sheared sheeted dike member	py-cp-po	remobilized Betts Cove	ore associated with fault breccia; one shaft present	Baird (1951), Riccio (1975)
2. Mount Misery	showing	at or near dike - pillow member contact	py-cp-po	Betts Cove type	grab sample from dump - 3.46% Cu, 0.02 oz./ton Au, 0.10 oz./ton Ag; shaft, pit and adit present	Snelgrove (1929), Upadhyay & Strong (1973), Douglas et al. (1940)
3. Burtons Pond	showing	sheeted dike member	cp-py-asp-sp ± Ag ± Au; 0.31 oz./ton Ag	stockwork Betts Cove	grab sample - 1.21% Cu, 2.68% As, 0.42 oz./ton Au	Douglas et al. (1940), Snelgrove (1929)
4. Long Pond	prospect	pillow lava member	cp-py-mag-gn ± Au	Betts Cove type	no massive ore, but lenses locally layered; maximum of 0.72 oz./ton Au detected in core	Nudulama Mines Ltd. (1956), Upadhyay (1974)
5. Muirs Pond	prospect	chloritized and schistose gabbro member	po-py-cp-asp(?) ± Au ± Ag	remobilized Betts Cove type	2-foot channel sample - 0.90% Cu, 0.28 oz./ton Au, 1.26 oz./ton Ag; massive grab sample - 1.78% Cu, 0.12 oz./ton Au, 0.88 oz./ton Ag	Advocate Mines Ltd. (1967)
6. Nippers Harbour No. 1	prospect	sheeted dike member	py-cp-po	stockwork Betts Cove	silicified and lens type setting, ore contains glassy quartz grains; best assay - 2.75% Cu, 1.05 oz./ton Au, 1.30 oz./ton Ag	Riccio (1975)
7. Betts Big Pond	showing	sheeted dike member within 200 m of pillow lava member	cp-po-py, ptl	Betts Cove type	2 pits present	Baragar (1954)
8. Nippers Harbour No. 4	showing	gabbro member cut by diabase dikes	py-cp ± Au	remobilized Betts Cove type	appears to be localized on a small shear zone	Advocate Mines Ltd. (1967)
9. The Lowlands	indication	diabase dike member	py-cp	remobilized Betts Cove type	close to Betts Cove Mine	Snelgrove (1929)
10. Fault Cove	indication	chlorite schist zone in dike and pillow members	py-cp	remobilized Betts Cove type	close to Betts Cove Mine	Upadhyay (1974)
11. Betts Cove Cliffs	indication	upper part of dike member	py-cp	remobilized Betts Cove type	close to Betts Cove Mine	Upadhyay (1974)
12. Betts Cove Brook	indication	sheeted dike member near pillow member	py-cp	remobilized Betts Cove type	close to Betts Cove Mine	Upadhyay (1974)
13. Red Cliff Pond Runout	indication	pillow lava member	py-cp	Betts Cove type	minor indication	Riccio (1975)
14. Nippers Harbour No. 7	indication	gabbro member	sp-gn-cp	insufficient data		Advocate Mines (1967)
15. Walsh (Noble) Cove	indication	gabbro member on Stocking Harbour	py-cp-gn	shear zone mineralization	shaft present	Baird (1951)
16. East Pond	indication	sheeted dike member	py	Betts Cove type	access best by canoe	Upadhyay (1974)
17. East Pond Point	indication	argillite in pillow	py	sedimentary deposit		Upadhyay (1974)
18. Phillips No. 1	indication	pillow lava member	py	Betts Cove type		Riccio (1975)
19. Long Pond East	indication	pillow lava member	py	uncertain		Upadhyay (1974)
20. Shiner's Prospect	showing	sheeted dike and pillow lava member	py-cp-po + ma ± sp ± gn(?)	Betts Cove type	4 shafts and numerous pits present	Maclean (1947), Dean (1977)
21. Young Cove	indication	sheeted dike member	py-cp-po	Betts Cove type		Dean (1977)
22. Middle Arm Point	indication	pillow lava member	py	insufficient data		Neale et al. (1960)
23. Rogues Harbour	prospect	quartz vein along Stocking Harbour	cp-py-po-gn ± Au	vein type (quartz)	vein 3-9 m wide, 825 m long; 2 shafts, 2 adits present; up to 2.64% Cu reported	Douglas et al. (1940), Baird (1951)
24. Nippers Harbour No. 8	indication	gabbro member cut by diabase dikes	cp-gn	vein type (quartz)	1 assay - 3.6% Cu, 1.0% Pb; vein is 1.2 m wide	Advocate Mines Ltd. (1967)
25. Axe Pond	showing	in ultramafic member at contact with pillow lava member	gn-sp	in shear zone (quartz + calcite	test pit and dump present	Baird (1951), Upadhyay (1974)
26. Snooks Arm Phillips No. 3	showing	ultramafic member	gn-mlr	cavity in shear zone	cavity fill 4-5% gn	Riccio (1975)

occurrences are uncertain but do not appear to be related to the formation of the ophiolite. These small deposits do not seem to be economically significant.

Pacquet Harbour Group

This group, though lacking a complete ophiolite stratigraphy, has been considered herein as largely ophiolitic in origin (see Chapters V and VI). The group hosts the second most valuable metallic mineral deposits on the peninsula, the four major sulfide bodies collectively described here as Consolidated Rambler Mines; it hosts numerous smaller occurrences as well.

CONSOLIDATED RAMBLER MINES

LOCATION AND PREVIOUS STUDIES: Consolidated Rambler Mines property is located on the La Scie highway, in the area of the Ming's Bight branch road, approximately 18 km east of the Baie Verte community. The geology and history of the mines were described by Douglas et al. (1940), Livingston (1942), Quinn (1945), Watson (1947), Gale (1971, 1973), Heenan (1973), Tuach (1976), and Tuach and Kennedy (1978); the following description is summarized from these sources as well as from conversations with D. Burton (1981), the mine geologist at the mine until its closure in 1982.

HISTORY OF DEVELOPMENT: Sulfide mineralization was first found in the Rambler area by Enos England in October, 1903, along the banks of Rambler Brook. The occurrence became known as the England vein and was originally claimed by England and T.E. Wells, a magistrate at Little Bay. In 1904, England and Wells optioned the property to Naylor and Company of New York, which provided funds for a shaft approximately 20 m deep; although mineralization was intersected, the grade was too low and in 1907 they relinquished their option.

The property lay idle until 1936, when Enos England and his son, William, discovered a new vein, the Rambler vein, approximately 200 m north of the England vein. From 1938 to 1944, most of the work on the prospect was carried out by the Geological Survey of Newfoundland. In 1944, the England-Wells mining grant was optioned to a group of St. John's businessmen, who formed Rambler Mines Limited. The property was reoptioned to Siscoe Gold Mines in 1945 and later, in 1951, to Falconbridge Nickel Mines Limited, who formed Rambridge Mines Limited.

In 1960, the property was expropriated by the Newfoundland Government under the Undeveloped Mineral Areas Act and was leased to M.J. Boylen Engineering Limited, who formed Consolidated Rambler Mines Limited. This new company started mine development of the Rambler vein in 1961, and brought it into production in 1964 as the Rambler, or Main Mine.

Consolidated Rambler Mines Limited also carried out exploratory drilling on a sulfide occurrence, the Norris showing, approximately 1.5 km east of the Main Mine. In 1965, a shaft was sunk and mine development was started. It was named the East Mine and was brought into production in 1967, the same year the Main Mine was exhausted and shut down; the East Mine was active until 1974. While the East Mine was operative, a geochemical soil survey led to the discovery of a deposit on the shore of Big Rambler Pond, approximately 1.5 km southwest of the Main Mine. This small deposit was mined out in 1969.

Early in 1970, the Ming ore body was discovered through airborne geophysical surveys flown at tree-top level and follow-up geochemical surveys; the Ming Mine was brought into production in 1971. Due to depletion of reserves and low metal prices, the Ming Mine was shut down effective May 1, 1982. Some exploration work was continued on the property in 1982.

LOCAL GEOLOGY: The Rambler property is underlain almost exclusively by volcanic and volcaniclastic rocks extensively cut by mafic dikes and sills; all of these rocks are polydeformed and metamorphosed to upper greenschist facies. The Burlington Granodiorite outcrops along the western margin of the property. Recent studies by Gale (1971) and Tuach (1976) described the distribution of rock types in the mine area.

Tuach (1976) recognized a generally east striking, moderately north dipping metavolcanic sequence composed of metamorphosed mafic flows, mafic volcaniclastic rocks, felsic volcaniclastic rocks and mafic intrusive rocks. In addition, there are minor occurrences of ultramafic rocks, chert, and silicic rocks. He divided the sequence into five mappable rock types that complexly and gradationally interdigitate. The following lithofacies, from south to north, comprise the sequence (Figure 8-6):

(i) mafic flows - mainly elongate pillow lavas with subordinate massive flows; based on geochemistry by Gale (1973) and in this study, the flows are mainly boninitic (see Chapter VI);

(ii) mafic volcanic and volcaniclastic rocks - almost half are flow lavas; these are mostly tholeiitic lavas (Gale, 1973; see Chapter VI); the remainder of this facies is composed of agglomerate, tuff, pillow breccia and reworked fragmental rocks, with minor chert, silicic rocks, and quartz-sericite schist;

(iii) mixed felsic and mafic rocks - varying proportions of felsic and mafic volcaniclastic rocks and pillow lava; these occur around the periphery of the acid volcaniclastic sequence;

(iv) felsic volcaniclastic rocks - dominantly coarse grained keratophyric fragmental rocks with subordinate lapilli tuff; quartz porphyroblasts are common;

(v) mafic sedimentary rocks - bedded, laminated volcanogenic sedimentary rocks, reworked crystal tuff, minor pillow flows and gray chert.

Tuach (1976) and Tuach and Kennedy (1978) suggested that this stratigraphy represents a conformable, northerly younging sequence. However, major drawbacks of their model include the occurrence of medium to large scale tight to isoclinal folds in the area, the sparseness of younging indicators, and the poor exposure in the area. In addition, I have noted local small scale examples of extreme layer-parallel transposition in the area and major faults may be present in the hanging wall of the Ming Mine, as marked by serpentinite pods there (see Chapter VII). Thus, a satisfactory sequence for the area has yet to be determined and confirmed.

The sequence in the Rambler area has undergone four phases of deformation, with the second and third phases being of major consequence. The second, or main deformation produced an intense penetrative transposition fabric that is generally parallel or subparallel to primary layering. Development of an intense L-fabric accompanied the formation of this schistosity and produced a northeasterly plunging mineral, clast and pillow lineation throughout the northern part of the area. The schistosity is axial planar to minor, tight to isoclinal, northeast plunging folds of which the axes are parallel to the D_M lineation. The existence of large scale D_M folds is probable, though their location is equivocal.

These structures are overprinted by a late, moderate to shallow northeast dipping crenulation cleavage produced during D_L. This cleavage is axial planar to open, recumbent, shallowly plunging folds of the main schistosity and primary layering. Metamorphic events accompanied both of the major deformations.

MINERALIZATION: Four major sulfide ore bodies have been found on the Rambler property, including the Ming, the Rambler (or Main), the East and the Big Rambler Pond bodies. The Ming and Rambler deposits are massive and strata-bound and apparently very similar in character; the East deposit is disseminated and strata-bound and the Big Rambler Pond body appears to have stockwork disseminated and stringer mineralization. All four deposits have been recrystallized and involved in the regional deformation of the area (Gale, 1971; Tuach, 1976; Tuach and Kennedy, 1978). The following descriptions of these four deposits have been adapted from Gale (1971) and Tuach (1976).

The Ming ore body (Plate 8-1) is a stratiform massive sulfide deposit with a strike length of 100 to 150 m and an average thickness of 5 m; it plunges to the northeast at 30° for a known length of 1000 m. The deposit is tabular and elongate, and consists of up to three ore lenses (Figure 8-7). Chalcopyrite was the main ore mineral, though gold and silver were also produced. Copper grades along the ore body are somewhat erratic.

The Ming body is located near and at the northeast margin of the felsic volcaniclastic unit at the contact with the mafic volcaniclastic unit. The relationship between the host rock and the ore body at this contact is complex, though the ore body generally occurs along the contact between the felsic and mafic rocks (Figure 8-6).

Mafic volcanogenic sedimentary and tuff units in the immediate hanging wall and between the ore lenses average 15 m in thickness and consist of thin, distinctly banded layers of actinolite-biotite schist and calcite-epidote-quartz schist. Thin, coticule-like, quartz-garnet-magnetite stringers occur within these rocks. Minor disseminated pyrite ($<1\%$) is the only mineralization in the hanging wall. Two concordant ultramafic bodies up to 10 m thick of unknown affinity have been recorded in drill holes in the hanging wall.

The footwall is composed of felsic rocks ranging from chrome-mica(mariposite)-bearing quartz-sericite schist to quartz-feldspar porphyroblastic keratophyre. Disseminated pyrite locally composes up to 10% of these rocks. No stockwork to the ore body has been recognized in the footwall.

The ore body is in sharp contact with the country rocks. Four types of ore are recognized, including massive pyrite ore, banded ore, massive chalcopyrite-pyrrhotite ore, and breccia ore. The contacts between ore types are transitional. Massive, fine grained pyrite ore is the most common, forming roughly 80% of the ore body. It comprises 70% pyrite, with lesser chalcopyrite and minor sphalerite, galena, quartz, actinolite and chlorite. Isolated fragments of quartz-sericite schist and mafic sedimentary rock also occur within this ore. Locally, variation in the dominant mineral in the ore produces banding. Pyrite is also the most common constituent of the banded ore, in which it alternates with thin bands of chalcopyrite-quartz-actinolite-biotite ± carbonate and, locally, sphalerite. Banding is parallel to that in the surrounding rocks. Massive chalcopyrite-pyrrhotite ore occurs in lenses up to 2 m thick in the massive pyrite ore and as layers near the main ore body in the host rock. It is also locally concentrated along the margins of mafic dikes that crosscut the ore body. Chalcopyrite composes up to 80% of the ore, with pyrrhotite and sphalerite constituting the remainder. Breccia ore is composed of massive pyritic and banded ores in a matrix of chalcopyrite and pyrrhotite. Brecciation ranges from fractured massive ore with minor infilling of fractures with chalcopyrite, to isolated rounded fragments in a chalcopyrite matrix. This is probably a depositional breccia; however, metamorphism has destroyed primary textures.

Several subordinate metallic minerals occur within the ore body, including arsenopyrite in the massive pyrite ore and the chalcopyrite-pyrrhotite ore, and galena, tetrahedrite and gold in the massive pyritic ore. One large pod (3 m in diameter) of tennanite was also discovered. Locally, cubanite occurs as exsolution lamellae in the chalcopyrite.

The ore body is crosscut by mafic dikes that display all of the major structures of the country rock; thus, the ore as well as the dikes appear to predate the main deformational events in the area.

The main fabric in the host rocks, S_M, is manifest in the ore body by the alignment of phyllitic minerals in siliceous portions of the ore and, locally, by the orientation of pyrite and chalcopyrite pods. Minor F_M folds occur within the ore body. The shape of the Ming body parallels the intense fragment and mineral lineation, L_M, in the area. The later deformation, D_L, imposed a fracture cleavage upon the ore body and, locally, disharmonic D_L folds occur. Massive chalcopyrite-pyrrhotite ore commonly occurs in the hinges of these folds. Based on polished section studies (Tuach, 1976), it appears that some brecciation of the ore occurred during D_L.

On the basis of mineral zonation within the ore body (copper and zinc toward the structural base), it has been suggested that the entire ore body is overturned (Heenan, 1973). As the zonation is not reversed on the opposing limbs of F_L folds, it is probably not primary and, therefore, does not reflect primary facing directions (Tuach and Kennedy, 1978).

The Rambler (or Main) body was described by Gale (1971); because the mine was shut down before his study, he obtained most of his data from a mine report by Baragar (1954). The following is an abridgement of Gale's work and is schematically summarized in Figure 8-8 (Gale, 1971):

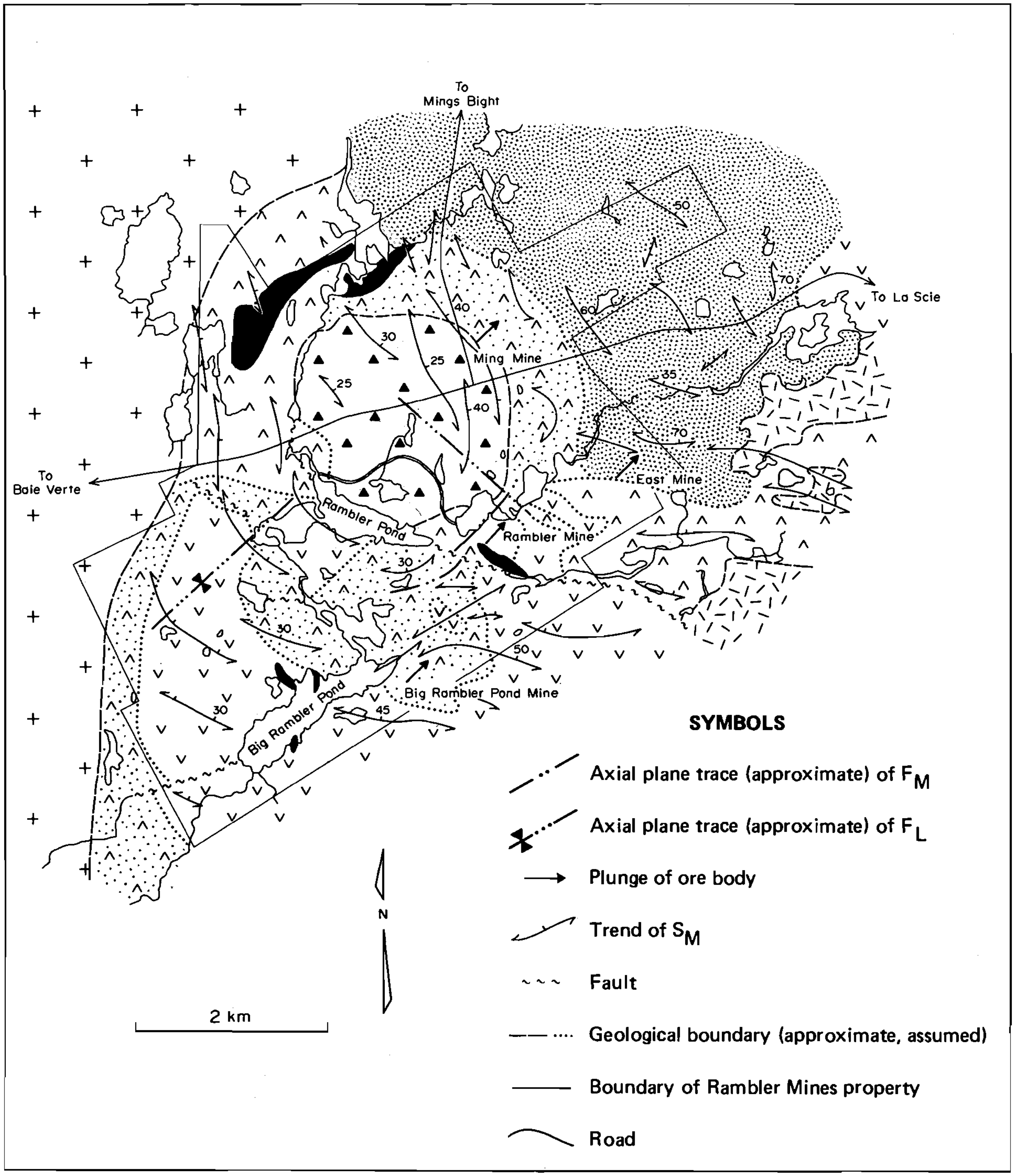


Figure 8-6: *Geology of the Consolidated Rambler Mines area (from Tuach and Kennedy, 1978).*

LEGEND

INTRUSIVE ROCKS

Cape Brule Porphyry

Burlington Granodiorite

MAFIC INTRUSIONS: Metadiabase and metagabbro

VOLCANIC SEQUENCE

MAFIC SEDIMENTS: Predominantly bedded and laminated graywackes, waterlain mafic tuffs and reworked mafic tuffs; minor conglomerate and chert horizons; minor mafic volcanic flows.

FELSIC VOLCANICLASTIC ROCKS: Predominantly dacitic agglomerate; minor felsic tuffs and/or flows; minor mafic tuffs and flows; minor chert and quartz-sericite schist.

MIXED FELSIC AND MAFIC ROCKS: Intercalated felsic and mafic volcaniclastics; mafic flows; minor quartz-sericite schist; minor chert horizons.

MAFIC FLOWS AND VOLCANICLASTIC ROCKS: Predominantly agglomerate, flow, mafic tuff and/or sediment; minor quartz-sericite schist; chert horizons and minor felsic intrusions.

MAFIC FLOWS: Predominantly pillowed and massive flows; minor mafic agglomerate and tuff; minor felsic intrusions.

N.B. Approximately 25% of the area underlain by the volcanic sequence is composed of mafic intrusive rocks.

Plate 8-1: *The Ming ore body outcropping beneath hanging-wall chlorite schist in the open pit just west of the main shaft.*

The ore body is 100 to 300 feet wide, seldom exceeds 50 feet in thickness, and has been traced by surface diamond drilling for a distance of approximately 3500 feet down dip from its surface exposures...

The main mineralization is a band of massive sulfide comprised mainly of pyrite with zones of sphalerite and chalcopyrite. In addition the deposits contain minor galena, pyrrhotite and recoverable gold and silver. Several pyritic parts of the ore body contained sufficient gold to warrant mining for that metal alone. Disseminated sulfides, mainly pyrite and traces of chalcopyrite, are found in the footwall for distances of up to 1000 feet below the massive sulfide horizon, but cannot be profitably extracted...

The sulfides vary from massive to weakly disseminated. Near the top of the ore zone sections of massive sulfide 15 to 20 feet thick were commonly encountered. Narrow bands of massive sulfide up to one foot thick occur below the main massive sulfide horizon. In addition to the massive sulfides chalcopyrite, pyrrhotite and pyrite occur in veins up to several cms in width, cutting across and paralleling the schistosity and as disseminations in the footwall schists.

Fine grained glassy quartz is often found in the sulfide veins...

a. Hangingwall

The hangingwall, in most places, is a band of 'magnetite tuff' described as "siliceous magnetite bearing siltstones, probably tuffs, varying to purplish sugary-textured quartzitic rock" (Baragar, 1954). This layer which may be up to ten feet thick, is present throughout much of the mineralized zone. However, where it is absent the massive sulfides are in direct contact with quartz-chlorite schists or quartz-chlorite-sericite schists...

The 'magnetite tuff' horizon passes upward into fine grained chloritic rocks which Baragar (1954) called 'streaked and mottled chlorite schists' and described as 'dark green chlorite schists streaked and banded with light blue-gray patches. They are probably intensely sheared fragmental rocks...' ...Baragar found that well defined bands of acidic fragmental rocks graded into the chlorite schists in several drill holes. In addition he observed sections of acidic material up to two feet in length which he considered to be either large fragments or thin flows... Since none of Baragar's larger intersections of acidic rock could be correlated between drill holes, they probably represent intersections of large blocks of acidic rock...

The blocks of acidic material are petrographically and chemically similar to the porphyritic acidic lavas found northwest of the Rambler Mine.

A band of acidic rocks overlies the chloritic schists and underlies a horizon of fragmental acidic rocks. The acidic rocks are porphyritic with 1 to 2 mm quartz and feldspar phenocrysts in a grayish to pink aphanitic groundmass and are similar to the acidic lavas exposed northwest of the Rambler Mine.

A horizon of fine grained sedimentary rocks overlies the acidic pyroclastics. These rocks have a pronounced banding and a distinct brown colour due to an abundance of biotite. These rocks are not known to crop out at the surface.

The uppermost rock unit is basic pillow lava. These rocks crop out along the road to the East Mine and form the main occurrence of tholeiitic lavas within the Rambler map area.

b. Footwall

The ore zone and immediate footwall rocks are mainly quartz-sericite and chloritic schists. Baragar describes the quartz-sericite schist as forming a vaguely defined layer in the upper portion of the ore zone and varying in thickness as it interfingers laterally with the chlorite schist in the lower portions of the ore zone. The two rock types normally have gradational contacts; however, in places they are separated by zones of quartz-chlorite-sericite schist.

The most common footwall rock, chlorite schist, underlies the rock types described above. This chlorite schist, consisting of stratiform sedimentary rocks and basic agglomerates, crops out in the bed of Rambler Brook. These chlorite schists are in turn underlain by basic lavas.

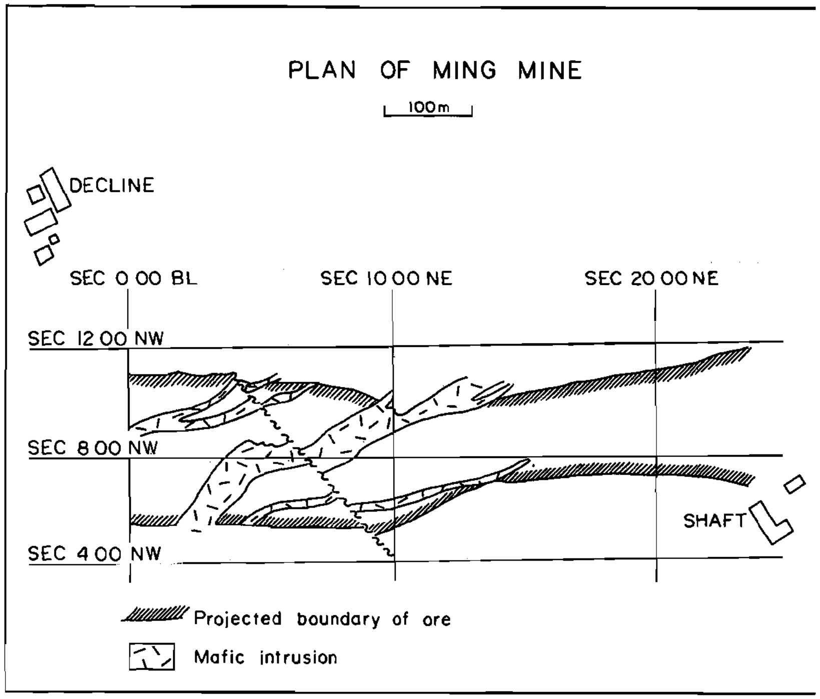
c. Structure

A conspicuous feature of some of the sulfide specimens is the separation of sphalerite-pyrite ores and pyrite-chalcopyrite ores into nearly monomineralic bands... In specimens of banded sulfides which have inclusions of the host rock, the banding in the sulfides is parallel to the weak-schistosity in the silicate inclusions...

Rock fragments enclosed in the massive sulfides have their long axes parallel to the long axes of lineated minerals in the fragments... The presence of mineral lineations in the sulfide ores indicate that the ore was deformed during the main deformation affecting the Rambler area.

The plunge of the ore body, 030/35NE, approximates the plunge of a minor F_2 fold axis (035/35NE) at Rambler Pond 2000 feet west of the mine.

Although it cannot be conclusively proven, it appears as if the ore body lies along the axis of a northeastward plunging fold.



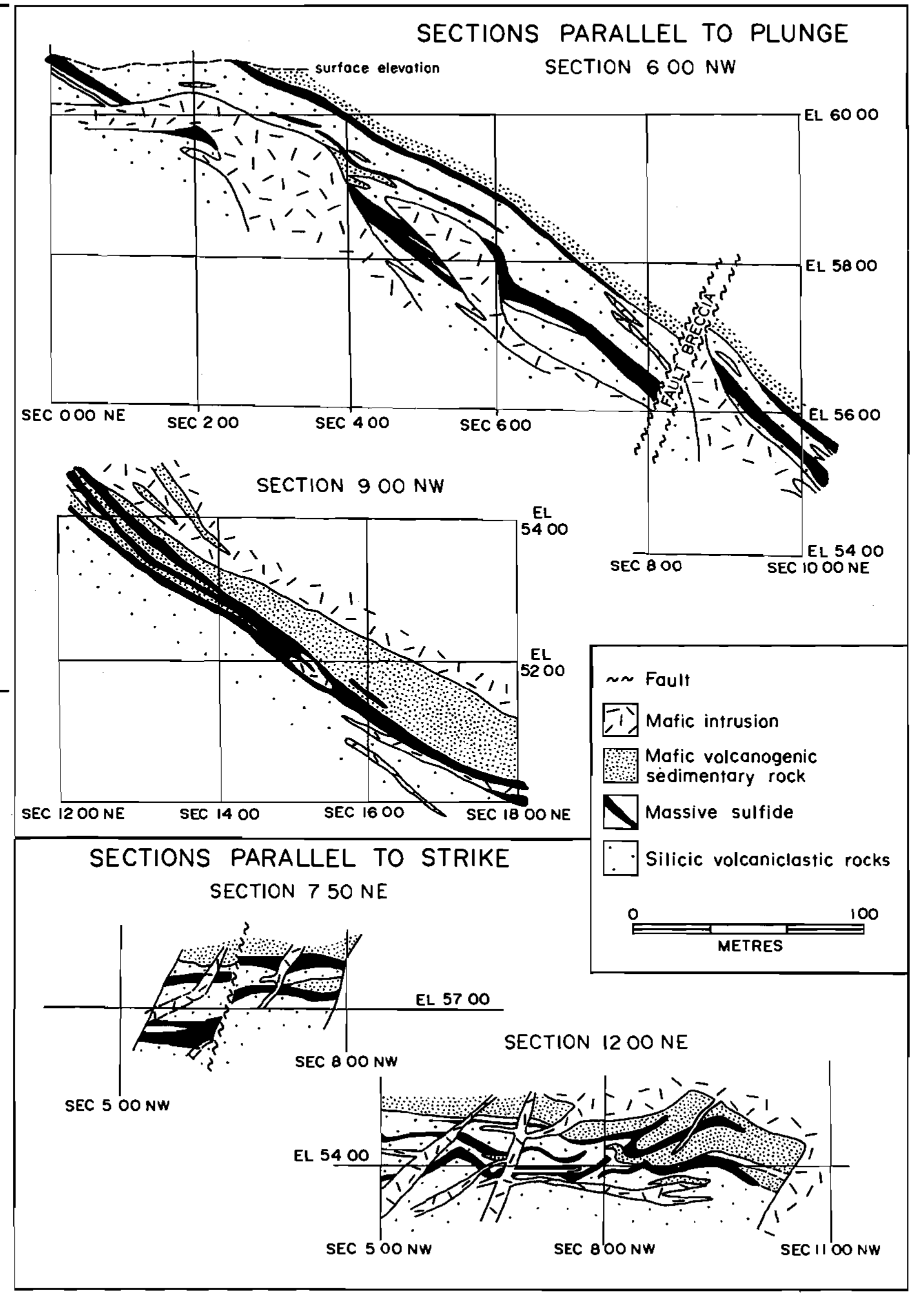
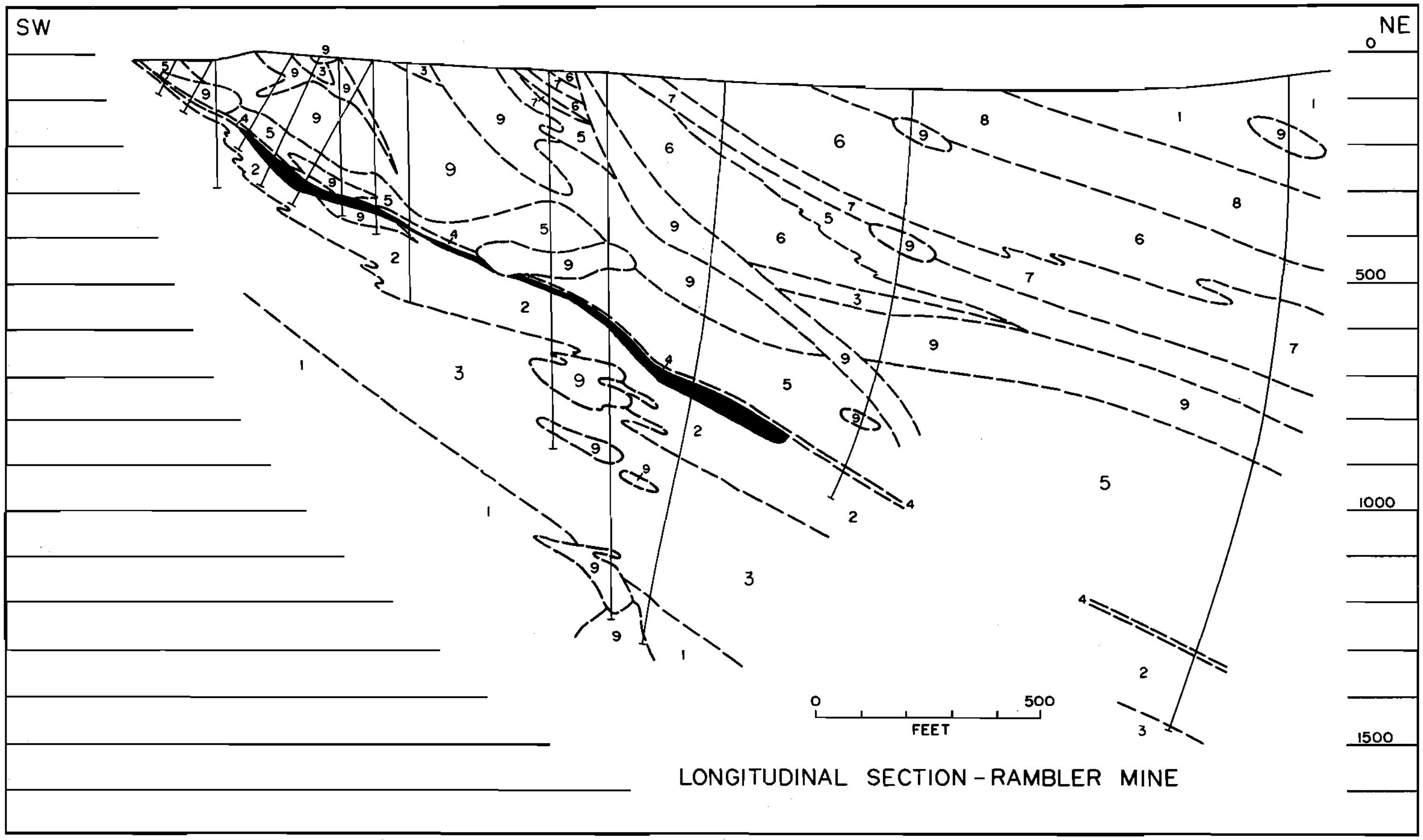


Figure 8-7: *Plan and section views of the Ming ore body (from Tuach and Kennedy, 1978).*



9	Diabase and diorite of at least two generations.
8	Graywacke and siltstone, probably tuff.
7	Felsic porphyry.
6	Coarse fragmentals composed largely of well defined felsic and intermediate volcanics.
5	Streaked and mottled chlorite schist grading into and apparently derived from massive felsite.
4	Siliceous magnetite-bearing siltstone, probably tuff, grading to purplish, sugary textured quartzitic rock adjacent to the ore zone.
3	Chlorite schist of uncertain origin.
2	Quartz-sericite schist.
1	Mafic lava.

Figure 8-8: *Longitudinal section of Rambler Mine (from Gale, 1971) based on drill data as shown.*

Tuach and Kennedy (1978) suggested that the Rambler and Ming ore bodies occupy the same stratigraphic level, but on different parts of a large D_M recumbent fold. Such a correlation is tempting considering the striking similarity of the host rocks to these bodies; at the present, though, there is no compelling evidence to verify this correlation.

The East Mine deposit was also described in detail by Gale (1971); the following is an abridgement of his description:

The East Mine ore body is mainly a disseminated sulfide deposit and unlike the Rambler Mine it does not contain horizons of massive sulfides. The main sulfide minerals are pyrite, chalcopyrite and pyrrhotite. Above the 500 level the mineralization consists mainly of pyrite which locally may constitute 75 percent of the rock. The average pyrite content of the upper levels of the mine is approximately 30 percent, whereas chalcopyrite rarely constitutes more than one percent. The pyrite content decreases below the 500 level and the lower levels of the mine, 850 and 1000 levels, contain approximately equal volumes of chalcopyrite and pyrite which together seldom amount to more than 15 percent of the rock.

The limits of the ore body are defined by copper assay values. The sulfide deposit is roughly ruler shaped with a width of about 400 feet and a thickness that is generally less than 100 feet. The deposit has been traced by diamond drilling for a distance of about 3000 feet down dip from its surface exposures...

The hangingwall rocks consist of basic agglomeratic and tuffaceous rocks, basic flows and/or sills, metadolerite dikes and acidic tuffaceous rocks. The hangingwall rocks are mineralogically and texturally similar to rocks found elsewhere in the Rambler area...

The dominant rock type in the footwall is quartz-chlorite schist which is commonly interbanded with 'quartz-eye' chlorite schists, biotite schists and actinolite schists. These schists are intruded by fine grained, often porphyritic, and medium grained metadolerite dikes and aphanitic basic dikes...

Fine grained, fine grained porphyritic and medium grained metadolerite dikes intrude the wall rocks and the ore zone. Two nearly vertical dikes, up to 50 feet thick, intruding the ore body can be traced on the 500, 625, 750, 850, and 1000 levels...

In contrast to the Rambler Mine, sphalerite and gold are present in only trace amounts. Sphalerite is rarely seen in hand specimen and gold is recovered as a by-product during smelting of the chalcopyrite concentrates.

The East Mine sulfide deposit exhibits a broad zonal pattern. Between the 375 level and the surface pyrite is the dominant sulfide and economic concentrations of chalcopyrite are rare. Below the 500 level the ratio of pyrite to chalcopyrite decreases rapidly and below the 750 level the volume of chalcopyrite and pyrrhotite equals or exceeds the volume of pyrite.

Above the 500 level pyrite may constitute more than 75 percent of the rock although the average pyrite content is less than 30 percent. The pyrite occurs mainly as disseminated to nearly massive lenses in quartz-chlorite and quartz-sericite schists. In addition pyrite occurs as veins up to six inches in width cutting across the pyritized and nonpyritized rocks. Chalcopyrite occurs as scattered blebs, 1-2 mm in length, in the disseminated pyrite and in chalcopyrite-pyrrhotite veins several cms wide.

Below the 625 level pyrite rarely constitutes more than 10 percent of the rock. The pyrite occurs mainly as disseminated porphyroblasts, 1-2 mm in diameter. Pyrite veins, several cms in width, may contain chalcopyrite and pyrrhotite.

The main chalcopyrite mineralization occurs between the 500 and 1000 levels. The chalcopyrite mineralization occurs mainly as disseminated blebs and as chalcopyrite-pyrrhotite veins.

Disseminated chalcopyrite-pyrrhotite mineralization consists of oriented blebs of chalcopyrite with variable, generally minor, amounts of pyrrhotite scattered throughout quartz-chlorite and 'quartz-eye' schists. The concentration of sulfide blebs is quite variable but rarely exceeds 15 percent of the rock by volume.

Individual sulfide blebs range in size from barely visible with a 10X hand lens up to slightly more than 1 × 2 cm. The sulfide blebs resemble the 'quartz-eyes' in their roughly ellipsoidal shape. The long axes of the sulfide blebs parallel the long axes of mineral and particle lineations in their host rock. Sulfide blebs in 'quartz-eye' chlorite schists have long axes parallel to the long axes of the quartz bodies... The contacts of the disseminated chalcopyrite-pyrrhotite mineralization are generally diffuse and often grade into disseminated pyrite mineralization...

The ore body is broadly concordant with the regional stratigraphy, i.e. the acidic tuffs, and its long axis lies parallel to the mineral and particle lineations in the host rocks.

No major fold structures have been recognized to date in the East Mine...

The main structural features of the East Mine are the mineral and particle lineations, L_1, produced by the first deformation. The most common particle lineations are rock fragments, 'quartz-eyes' and sulfide blebs. Particle and mineral lineations measured on oriented hand specimens have a mean orientation of 035/35 NE.

Both Gale (1971) and Tuach (1976) suggested that the East Mine is the distal equivalent of the Rambler deposit.

The Big Rambler Pond deposit is very small and has been only briefly described by previous workers (Gale, 1971; Tuach and Kennedy, 1978); the following description is taken from Tuach and Kennedy (1978):

This deposit is pipe shaped and consists of disseminated and stringer pyrrhotite, chalcopyrite, and pyrite mineralization in a quartz-chlorite schist with a strike width of approximately 30 m, a thickness of 15 m, and a length down plunge to the northeast of 60 m.

The country rocks are mafic pyroclastic and mafic pillowed rocks. Pyrite- and mariposite-bearing chert beds are present in the vicinity of the deposit.

Several subeconomic deposits consisting of stringer and disseminated chalcopyrite, pyrrhotite, and pyrite mineralization in quartz-chlorite schists are present approximately 300 to 600 m northeast of the Big Rambler Pond deposit. These deposits are associated with pillow lavas and mafic volcaniclastic rocks and have cherts and silicic rocks in close association with them. The deposits are small, pipeshaped, and continuous down plunge to the northeast for up to 500 m.

ORIGIN OF THE ORES: Early workers considered mineralization in the Rambler area to be concentrated in veins along sheared zones of chloritic schist (Douglas et al., 1940; Quinn, 1945; Watson, 1947). Livingston (1942) related the mineralization to shearing and the emplacement of mafic dikes in the area. More recently, the work of Gale (1971), Tuach (1976) and Tuach and Kennedy (1978) demonstrated that the Rambler area deposits, with the possible exception of the Big Rambler Pond deposit, are stratiform and spatially related to the local felsic volcanics. The nature of the Big Rambler Pond ore body is, at this time, uncertain although it has been suggested that the deposit represents stockwork type mineralization (Tuach and Kennedy, 1978). These workers also showed that ore deposition in the area predated the regional tectonism, though the bodies have been remobilized, probably accounting for their similar shapes and orientations. These characteristics indicate that the Main, East and Ming bodies were most likely deposited at an ocean floor - sea water interface (Gale, 1971; Tuach and Kennedy, 1978); the Big Rambler Pond deposit may represent part of the stockwork to this depositional system.

The tectonic setting of these deposits is unclear at this time; however, regional implications indicate that the Pacquet Harbour Group occurs in a paleo-forearc, and the local geology, therefore, displays characteristics transitional between the ocean floor and an island arc (see Chapter IX). The mineralization occurs along the northern periphery of a mass of high magnesian (boninitic) lavas, near their contact with tholeiitic lavas (Gale, 1971); the Big Rambler Pond deposit appears to be within the boninitic terrane. These boninites are correlated with those of the Betts Cove Complex, so this part of the Pacquet Harbour Group is considered as ophiolitic (see Chapter VI). The nature of the remainder of the group, including the tholeiitic lavas, is uncertain; this portion may represent either an overlying island arc, ocean island type volcanism, or possibly a rare manifestation of an ophiolite suite; in the latter case, the felsic rocks may represent an unusual case of extrusive trondhjemites in an ophiolite. However, the felsic volcanic rocks at the mine sites may be more akin to an early island arc setting; also, sedimentary and volcaniclastic rocks along strike from the mineralized zone support the concept of an island arc type setting. On the basis of the given correlations, the Big Rambler Pond deposit appears to be similar to Betts Cove type stockwork deposits, whereas the stratiform deposits in the area occupy an apparently unique depositional environment that seems to be related to boninitic lavas and associated with felsic volcanic rocks. These latter deposits are herein termed Rambler type deposits.

The Rambler deposits are unlike ophiolitic deposits in the strict sense. Instead, the Rambler deposits show characteristics of both "Kieslager or Besshi type" deposits and primitive types as outlined by Hutchinson (1980). The mineralogy of the Rambler deposits closely corresponds to that of the Kieslager deposits, in which zinc is subordinate to copper. The association of the Rambler deposits with felsic volcaniclastics most closely corresponds to the setting of primitive type deposits, which occur in differentiated volcanic sequences.

PRODUCTION AND RESERVES: Production figures for Consolidated Rambler Mines, shown in Table 8-3, have been taken from Tuach and Kennedy (1978) and updated with reports to the Department of Mines and Energy (B. Hynes, personal communications, 1981).

In both the Rambler and East Mines, subeconomic mineralization extends beyond the ore assay boundaries. Little is known about the lateral extent of the subeconomic mineralization in these deposits, since little exploratory work was carried out during mining operations.

OTHER METALLIC MINERAL DEPOSITS

Numerous other smaller metallic mineral occurrences have been noted within rocks of the Pacquet Harbour Group; these are summarized in Table 8-4 and located in Figure 8-5. (Note that only the major occurrences are shown in Figure 1-1). The table is based largely upon data from the Mineral Occurrence Data System of the Department of Mines and Energy; selected references are given for further details.

Nearly all of these occurrences are typically disseminated and stringer pyrite-chalcopyrite associations in mafic and felsic volcanic-volcaniclastic rocks with local mineralization apparently remobilized in small quartz veins and masses. The Rambler # 2 Prospect was interpreted as a stockwork deposit by Tuach (1978); it is possible that many of the stringer and disseminated occurrences in chloritic basalt and quartz-chlorite schist are also this type, although available evidence is inconclusive. The Lever-Tuach # 3 showing is the only undeveloped occurrence of massive mineralization in the group; at this showing, the massive ore forms stringers up to 0.6 m thick.

Vein-type deposits (Au, Ag, py, cp, bo) occur in close association with the above-mentioned occurrences, though locally this type occurs alone, such as in the Stuckey Vein. It is suspected that the mineralized veins represent the remobilization of metallic minerals from the host Pacquet Harbour rocks.

Hanks Occurrence appears to be atypical of mineralization in the group, as it contains disseminated and stringer galena in addition to pyrite and sphalerite. The only other occurrence of galena in the group is in the massive Rambler and Ming ores.

SUMMARY OF METALLIC MINERALIZATION

Three major zones of mineralization within the Pacquet Harbour Group, the Ming, Rambler and East Mines, are

Table 8-3: *Summary of production at Consolidated Rambler Mines.*

Deposit	Life Span	Tonnage (short tons)	Grade
Rambler (Main)	1961-1967	440,000	1.3% Cu, 2.16% Zn, 0.15 oz./ton Au, 0.85 oz./ton Ag
East	1967-1974	2,130,854	1.04% Cu
Big Rambler Pond	1969	50,000	1.2% Cu
Ming	1971-October 1981	2,121,388	<3.5% Cu, approximately 0.07 oz./ton Au, approximately 0.06 oz./ton Ag

Table 8-4. *Pacquet Harbour Group metallic mineral occurrences. (Number refers to Figure 8-5.)*

OCCURRENCE	STATUS	HOST ROCK	MINERALOGY	TYPE	COMMENTS	SELECTED REFERENCES
27. Uncle Mike	indication	quartz-sericite schist	py-cp	disseminated	zone approx. 1 m wide; drilling by Rambler Extension Group	Livingston (1942), Quinn (1945)
28. Uncle Enos	showing	quartz-sericite schist	py-cp-Au	disseminated	zone approx. 120 m long x 10 m wide; drilling by Jawtam Key Gold Zones Ltd.; avg. just over 0.5 oz./ton Au	Quinn (1945), Livingston (1942)
29. Angus Vein	indication	quartz-sericite schist	py-cp	disseminated	some drilling by Rambler Extension Group	Livingston (1942), Tuach (1978b), Quinn (1945)
30. Rambler No. 1 Prospect (Uncle Theodore vein)	showing	quartz-chlorite-sericite schist	py-cp-po-sp	disseminated - stringer	deposit approx. 600 m x 12 m, drilled to a depth of 90 m; avg. 0.4% Cu over 42 m; best intersection 1.23% Cu, and 2.4% Zn over 7.5 m	Livingston (1942), Quinn (1945), Tuach (1978)
31. Rambler No. 2 Prospect (Uncle Will Vein)	showing	silicified mafic breccia	py-po-cp	stockwork	area of 12 m x 12 m with avg. grade of 2.0% Cu	Livingston (1942), Quinn (1945), Tuach (1978b)
32. Rambler No. 3 Prospect (Uncle Bill Vein)	showing	quartz-chlorite-sericite schist	py-po-cp	disseminated - stringer	drilled to depth of 60 m; avg. grade 0.49% Cu over 17.7 m	as above
33. Hanks Occurrence	indication	chlorite-quartz-sericite schist	py-sp-gn	disseminated - stringer	over an area of 210 m^2, no vertical continuation	Tuach (1978b)
34. CR Occurrence	indication	chloritic basalt	py-cp	disseminated - stringer	basalts nearby are boninitic	Tuach (1978b)
35. Sids Pond	indication	chloritic basalt	py-cp	disseminated - stringer	nearby basalts are boninitic	Tuach (1978b)
36. Northeast Pond	indication	greenschist	py-po-cp	disseminated - stringer	within contact aureole of Burlington Granodiorite	Neale (1958a), Tuach (1978b)
37. Lever-Tuach No. 2	indication	quartz porphyry and chlorite schist	py-cp	disseminated - stringer	nearby pillow lavas are boninitic, 0.12% Cu	Tuach (1978b)
38. Lever-Tuach No. 3	showing	chlorite schist	py-cp, Au, Ag	stringers of massive ore up to 0.6 m thick	grab sample 27.6% Cu, 0.009 oz./ton Au, 0.97 oz./ton Ag	Tuach (1978b)
39. Lever-Tuach No. 4	indication	quartz-porphyry and chloritized schist	py-cp, Au, Ag	disseminated - stringer	grab sample yielded 1.53% Cu, trace Au, 0.18 oz./ton Ag	Tuach (1978b)
40. Nash Showing	indication	quartz vein in pillow lava	cp	massive vein and stringers in lava	15 cm wide, best sulfides in lava	Tuach (1978b)
41. Woodstock Rd.	indication	lineated amphibolite	py-cp	dissem. and assoc. with quartz veinlets	locally, Cu estimated at 1%	Coates (1970)
42. Belly Pond Brook No. 1	showing	mafic metavolcanics near contact with felsic metavolcanics	py-cp-po	dissem. - stringer and locally in quartz veins	representative samples of drill hole intersection yielded 0.01 - 0.1% Cu, 0.003 - 0.01% Pb, and 0.003 - 0.03% Zn	Coates (1970), Leslie (1976)
43. Phillips No. 4	showing	mafic pillow lava	cp-py	dissem.; also assoc. with quartz-epidote pods	pillow lavas along strike to west are tholeiitic	Riccio (1975)
44. South Yak Lake W.	indication	siltstone	py	sedimentary	locally up to 50% py in 0.5 cm laminated siltstone	Riccio (1975)
45. Stuckey Vein	showing	quartz vein	Au, Ag	vein type	18 m long x 24 m thick quartz vein; channel sample over 0.54 m yielded 0.13 oz./ton Au, 0.43 oz./ton Ag; channel sample over 0.6 m yielded 0.18 oz./ton Au, 3.19 oz./ton Ag	Tuach (1978b)
46. Woodstock Brook	showing	quartz veins	py-cp	vein type		Baird (1951)
47. Pacquet Harbour S.	indication	quartz vein	py-cp	vein type	nearby aplites, presumably related to Dunamagon Granite, also contain minor mineralization (sulfide)	Baird (1951), Neale (1958a), DeGrace et al. (1976)
48. N. England's Brook	indication	mafic schist near felsic volcanic rocks	py	uncertain		Neale (1958a)
49. England's Brook	indication	uncertain	py	uncertain		Neale (1958a)
50. Woodstock	indication	mafic pyroclastic rocks	py-cp	uncertain		Coates (1970)
51. Belly Pond Brook No. 1	indication	amphibolite	py-cp	uncertain		Coates (1970)

continued.....

Table 8-4: *(continued)*

52. Phillips No. 5	showing	uncertain	py-cp	uncertain		Riccio (1975)
53. Phillips No. 6	showing	uncertain	py-mag-cp	uncertain		Riccio (1975)
54. Belly Pond Brook No. 2	indication	uncertain	py	uncertain		Coates (1970)
55. Phillips No. 7	indication	felsic pyroclastics	py	uncertain		Riccio (1975)
56. Phillips No. 8	indication	mafic and felsic pyroclastic rocks	py	uncertain		Riccio (1975)
57. Phillips No. 9	indication	mafic pyroclastic rocks	py	uncertain		Riccio (1975)
58. Phillips No. 10	indication	mafic pillow lavas	py	uncertain	host lavas are tholeiitic	Riccio (1975)
59. S. Yak Lake W.	indication	uncertain	py	sedimentary	locally up to 50% py in 0.5 cm laminated siltstone	Riccio (1975)
60. S. Yak Lake S.	indication	uncertain	py	uncertain		Riccio (1975)
61. S. Yak Lake	indication	uncertain	cp	uncertain		Riccio (1975)
62. Lever-Tuach No. 1	indication	quartz vein	bo	vein	disseminated bornite in quartz vein at granodiorite margin	Tuach (1978b)

dominantly pyrite-chalcopyrite deposits associated with felsic volcanic-volcaniclastic rocks. The bodies apparently occur near the contact of boninitic and tholeiitic mafic lavas. Significant gold, silver and zinc values are found in these deposits. The disseminated and stringer East Mine deposit may be the distal equivalent of the massive Ming and Rambler ore bodies. This style of mineralization is herein termed the Rambler type; its tectonic setting is uncertain at this time. A fourth major deposit in the group, the Big Rambler Pond body, appears to be a simple pyrite-chalcopyrite stockwork-type deposit (Tuach and Kennedy, 1978) and available data indicate that it is associated with the boninitic lavas. Considering the correlation of the Pacquet Harbour Group boninitic lavas with the lower lavas of the Betts Cove Complex (see Chapters V and VI), the Big Rambler Pond deposit may be a small Betts Cove type deposit. Most of the subordinate metallic mineral occurrences in the group are similar to these major occurrences, though vein deposits such as the Stuckey Vein are locally significant, and disseminated lead-zinc mineralization is found at Hanks Occurrence.

Point Rousse Complex

The Point Rousse Complex has yielded the least amount of metallic minerals of the ophiolitic suites on the Baie Verte Peninsula. At the present time, however, it is one of the most attractive prospecting terrances among the ophiolites since it hosts the old Goldenville gold mine and other gold showings. This complex also hosts numerous small sulfide showings.

GOLDENVILLE MINE

LOCATION AND PREVIOUS STUDIES: The main shaft of the abandoned Goldenville mine is located on the Point Rousse Peninsula, approximately 2 km inland from the western shoreline of Ming's Bight (Figure 8-9). It is accessible by either an overgrown footpath from the bight or a network of footpaths that originate in the community of Ming's Bight. The mine was described by Howley (1918), Snelgrove (1935), Watson (1947) and Frew (1971); additional notes on the history of the property have been compiled by Martin (1983) and are found in directors' reports for the mine.

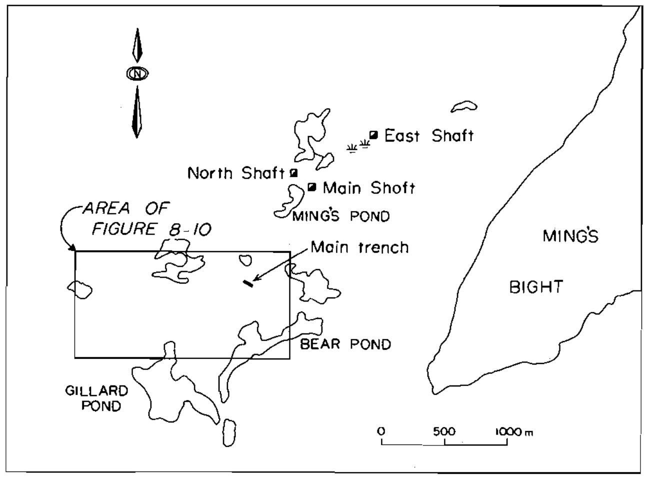


Figure 8-9: *Location of Goldenville Mine and associated trenches and shafts.*

HISTORY OF DEVELOPMENT: The occurrence of gold in quartz veins in the Ming's Bight area was known prior to 1867, as Murray notes the mineralization in his report for 1867 (Murray and Howley, 1881). Early claims in the area were staked by S. McKay and C.F. Bennett, both of Tilt Cove fame, in 1877, by A. Guzman in 1879, and by Capt. A.B. Cunningham in 1883. Prospecting in the area at that time was difficult since it was part of the French Shore; work had to be suspended periodically because of litigation as well as interference by French warships. D.J. Henderson and A.J. Harvey acquired claims in the area in 1897; following Henderson's death, these claims were optioned to J.R. Stewart, a former Little Bay mine manager. In 1902, Stewart located auriferous gravels, and by 1903 had located their source, the Goldenville prospect. Stewart formed the Goldenville Mining Company Limited in the same year. The directors' report for 1906 described the subsequent development of the gold mineralization as follows:

> The first opening made on the vein was by means of a trial shaft 50 feet deep, which was sunk on the eastern part of the deposit; later the

sinking of a working shaft was commenced at a point on the lode, half a mile west from the trial shaft. When the working shaft was attained a depth of 17 feet, a shipment of 23 tons were taken therefrom and made to Brookfield Mine, Nova Scotia, for the purpose of having a milling test made; the ore upon being treated by the amalgamation and cyanide processes, yielded 10.1/5 oz. of melted Gold...

The results of the mill run having proved satisfactory, the sinking of the working shaft was continued, until it reached a depth of 100 feet.

At a point in the shaft 80 feet below surface, levels were begun and have been driven East and West on the lode, for 80 feet and 51 feet respectively, the width of the veins being from five to thirteen feet.

Another inclined shaft was sunk into the main mineralization zone, just to the northeast of the main shaft, and a shallow shaft was lowered in a separate zone of mineralization about 190 m north of the main shaft. In 1906, a ten stamp gold mill and a Wilfley Concentrator were installed at the main shaft. By the end of 1906, the mine had produced approximately $3000 in gold, approximately one-tenth of the original investment in the property; as a result, Stewart and the other directors decided to shut down the mine.

In 1935, the property was optioned to the N.A. Timmins Corporation, who dewatered and sampled the main shaft. Trenching was carried out in 1937 by the Newfoundland Prospecting Syndicate to the southwest of the main shaft. In 1961, M.J. Boylen interests carried out geological surveys and some diamond drilling on the property. Presently, the Goldenville property is still being held as a fee simple mining grant by the heirs of A.J. Harvey. There has been some renewed exploration interest in the area over the past few years by Noranda.

LOCAL GEOLOGY: The Goldenville mine area is located in the Point Rousse cover sequence, in the core of the major syncline that folds the Point Rousse Complex (Figure 1-1; see Chapter VII). The rocks in the area of the old mine include mafic volcanic and volcaniclastic rocks, graywacke, iron formation, and unseparated greenstone. It is uncertain if the volcanic rocks and chemical sediments form the top of the Point Rousse ophiolite or if they compose part of the overlying sequence of rocks. The sequence strikes east-northeast and dips moderately to the northwest. Detailed descriptions of the sequence in the area immediately surrounding the mine are lacking, but Fitzpatrick (1981) described the stratigraphy of the same sequence immediately to the southwest of the property (Figure 8-10). Here, the sequence is repeated in a "Y" shaped pattern opening to the west, indicating that the regional syncline tightens to the east (Fitzpatrick, 1981). The stratigraphy of both limbs appears to be similar, and the sequence at the mine site seems to be an extension of the southern limb; the northern limb appears to be either missing or unexposed in the mine area. The overall stratigraphy consists of four major members, including a basal pillow lava and flow breccia member, a medial mafic tuff member overlain by ferruginous chert and iron formation, and a capping unit identical to the basal member (Fitzpatrick, 1981). Fitzpatrick (1981) described the mafic members as follows:

<u>Pillow Lava and Flow Breccia</u>

The rocks range from virtually unbrecciated flows to highly brecciated varieties where the fragment to matrix ratio exceeds 2:1. The size and angularity of clasts in these volcanics is variable. The clasts range from silt sized particles to blocks measuring 30 cm across and are typically sub-rounded.

Flows are locally amygdaloidal with vesicle infillings of quartz and calcite; rarely epidote amygdules are seen.

<u>Mafic Tuff</u>

The unit has a variable thickness, tending to pinch out in some areas and swell in others along strike. The average thickness of the unit is approximately 50 m. This tuffaceous horizon always occurs proximal to the ferruginous chert and iron formation unit.

The unit is devoid of clasts such as those found in the volcanic flows. However, some exposures contain variable amounts of particles usually less than 1 mm in diameter. The rocks also contain small lenses of magnetite and/or red chert. Where both are present the latter is usually less abundant. Considerably more epidote alteration than is found in any other rock type is associated with this unit.

The unit thus contains both fine-grained, distal, mafic pyroclastics and minor chemical sediments. It is likely that this horizon developed during the waning stages of volcanism.

The ferruginous chert and iron formation member, like the mafic tuff, is thin and discontinuous (Figure 8-10); the two units do not necessarily occur together. The ferruginous chert and iron formation member appears to be highly variable along strike. Fitzpatrick (1981) described a good exposure of this member as follows:

... the unit is approximately 7 m thick. At the base, directly above the tuff, 1.5 m of red chert nodules up to 5 cm across along with other detritus are set in a matrix of chlorite and magnetite. The bed is likely reworked material from ferruginous chert of higher relief. Five metres of massive ferruginous chert overlies the reworked zone. The bed is inhomogeneous and contains random lenses of quartz and magnetite... areas within this bed exhibit signs of hydraulic fracturing...

The ferruginous chert is overlain by 0.5 m of magnetite iron formation. This bed contains up to 30% sulfides (mainly pyrite) in places.

In addition to these members, Fitzpatrick (1981) noted that the sequence is intruded locally by small bodies of diabase and diorite.

MINERALIZATION: The Goldenville deposit consists mainly of gold- and sulfide-bearing quartz veins and disseminated gold and sulfides within a well defined zone of interbanded ferruginous chert and iron formation. Rock analyses (Frew, 1971; Fitzpatrick, 1981) indicate that the gold is concentrated in these chemical sedimentary rocks in the area; in these rocks it averages 15 to 20 times the background of 3 to 4 mg/t in the surrounding rocks. In addition, fractured and veined chert locally contains up to four times more gold than the mainly massive chert. Watson (1947) described the mineralization in the mine area as follows:

White quartz and minor amounts of carbonate, pyrite, chalcopyrite, specularite, chlorite, and gold have been deposited in the zone of chert and iron oxides from hydrothermal solutions. The quartz occurs principally as lenticular veins parallel to the foliation, but is also present as irregular veins of smaller size which are definitely cross-cutting. The individual veins are generally less than 6 inches thick, but commonly occur in groups. For example, at the No. 1 Shaft, there is a zone at least 4 feet wide consisting almost entirely of lenticular quartz veins which range from 4 inches to 0.25 inch in thickness.

A little rusty-weathering yellowish-brown carbonate and also pink and white carbonates occur in the quartz veins and as small veinlets in the mineralized zone.

Pyrite is present in the magnetite-hematite-quartz rock and less commonly in the associated chlorite schist as disseminations and distinct veinlets. It also occurs along the margins of, and locally within, the quartz veins. The pyrite forms euhedral or subhedral cubic crystals averaging 1 to 2 mm in diameter. The crystals commonly contain numerous small relict grains of magnetite which owe their preservation to incomplete replacement.

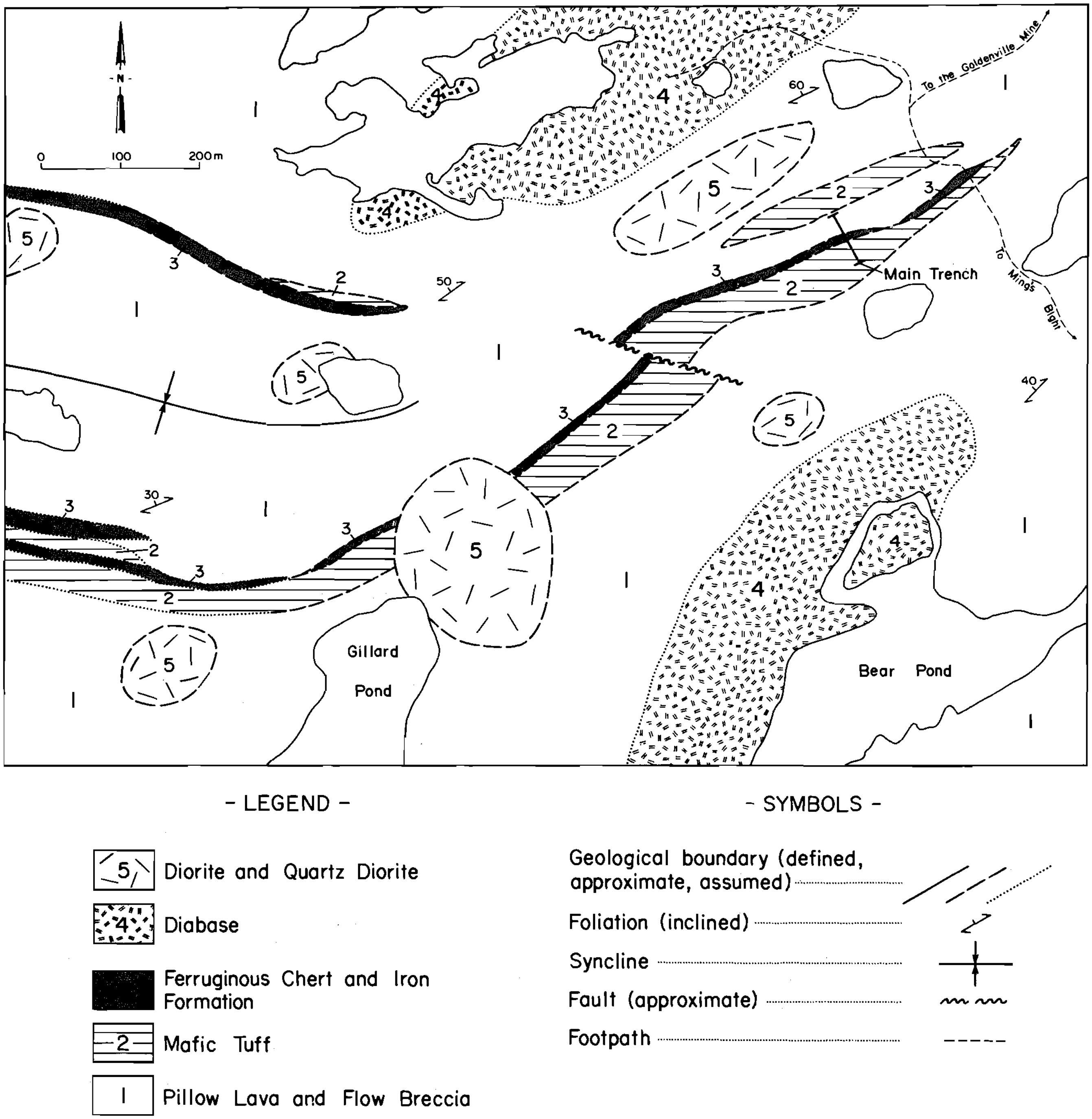


Figure 8-10: *Geology of the southwestern extension of the Goldenville horizon (from Fitzpatrick, 1981).*

Chalcopyrite occurs as minute blebs and veinlets in the pyrite. It was also observed in small amounts within well-defined veinlets of specularite and chlorite which cut the other minerals of the ore zone.

The specularite-chlorite veinlets are most commonly about 0.1 inch thick. They consist of large plates of hematite arranged normal to the walls, which are separated by fine-grained chlorite.

Gold is present as minute grains in narrow silicate veinlets which cut the pyrite.

The North Shaft was sunk on a narrow zone of white quartz veinlets in chlorite schist. The veinlets contain thin sericitic partings and sparsely disseminated, small pyrite grains.

The mineralized strata are apparently extensive, as noted by Watson (1947):

The mineralized zone is best exposed at the East Shaft, the Main Shaft, and the Main Trench. Immediately south of Big Head, on the western shore of Ming's Bight, are a number of lenses and nonpersistent beds of dark red ferruginous chert containing magnetite and hematite. The character, position, and attitude of these rocks suggest that they are the continuation of the Goldenville zone.

In addition, the zone has been traced to the southwest by workers from Noranda Exploration Co. Ltd. (Fitzpatrick, 1981) (Figure 8-10). It has also been reported that several outcrops of ferruginous chert occur near the west end of Green Cove Pond (Tuach, 1978b); these may also be extensions of the Goldenville chert zone. The thickness and character of the mineralized strata are variable; three sections through the zone are compared in Figure 8-11. The two sections to the southwest of the main shaft are similar, but the sequence nearest the main shaft is much thinner and is characterized by more magnetite-hematite chemical sedimentary rocks and much thicker quartz veins than the other sections. Detailed sections for other parts of the mineralized zone are unavailable, though Snelgrove (1935) noted that the mineralized zone at the main shaft is approximately 2.75 m thick, one-third of which is quartz, and at the east shaft the zone is approximately 1.50 m thick with only rare quartz.

ORIGIN OF THE ORE: Snelgrove (1935), Watson (1947), Frew (1971) and Fitzpatrick (1981) all noted that the concentration of gold in the chemical sediments indicates that the mineralization is strata-bound and related to the deposition of these sedimentary rocks. Fitzpatrick (1981) noted that the host sediments were deposited during a hiatus in volcanism and suggested that the gold may have been either (i) inherited from the volcanic pile by fumarole activity or (ii) scavenged from seawater during deposition of the chert and iron formation.

Gold mineralization is most highly concentrated in the extensively fractured and veined ferruginous chert, indicating that the gold was remobilized from the sediments into the veins. This is particularly evident near the main shaft, where quartz veins attain their maximum development in the area and gold values are highest. Since the veins locally crosscut the regional foliation, remobilization of the gold appears to have been a late event.

PRODUCTION AND FUTURE POTENTIAL: The total gold production recorded from the Goldenville mine is 158 oz. (Snelgrove, 1935). In 1904, the mine sent a sample shipment from the main shaft, which produced an 11 oz. gold brick, the first gold brick produced from Newfoundland gold (W. Martin, personal communication, 1979). The remainder was produced during the operation of the local plant in 1906.

Surface exposures in the mine area were sampled and assayed by Snelgrove (1935) and the results are shown in Table 8-5. From this table, it appears that the gold is commonly associated with pyrite. A grab sample of the concentrate tailings obtained during the present study was determined to contain 104 g/t Au (assay by A. Lee, government geochemist).

Considering that the mineralized zone is discontinuous for over 2 km strike length and that most of this area is covered by glacial drift, it appears that much more work is necessary before the potential of the area is determined. The most useful guides to gold mineralization within the zone appear to include (i) a concentration of magnetite-hematite layers, (ii) extensive quartz veining, and (iii) the presence of pyrite.

OTHER OCCURRENCES

Other metallic mineral occurrences in the Point Rousse Complex are summarized in Table 8-6 and located in Figure 8-5. Essentially, two major types of mineralization are observed within the complex, including volcanogenic sulfide and vein type mineralization. In addition, there are incidental showings of sedimentary sulfides at Deer Cove and a chromite occurrence at Kidney Pond.

The volcanogenic sulfide deposits all appear to be simple py-cp-po disseminated, stringer, and less commonly massive deposits in greenschist; the most significant is at Mud Pond. Locally, such as at Gillard Pond, minor sphalerite and galena are associated with the deposits. Remobilization of mineralization into veins appears to have been significant in the complex and is best typified by deposits at Penny Cove and the Barry-Cunningham property. In both cases, the presumed source of the vein mineralization is disseminated to massive sulfides in nearby greenstones. This source is different from that of the Goldenville deposit, where the ferruginous chert and iron formation are the source rock for gold mineralization in veins.

Advocate Complex

The Advocate Complex is host to the oldest known mining concern on the Baie Verte Peninsula, the Terra Nova Mine. The complex is a highly unorthodox prospecting terrane because of the extreme tectonic dismemberment of the ophiolite, its structural interleaving with younger cover rocks and the presence of olistostromes in the cover sequence (see Chapter V).

TERRA NOVA MINE

LOCATION AND PREVIOUS STUDIES: The Terra Nova massive sulfide body is located on the Baie Verte River directly opposite the junction of the Seal Cove road and the Baie Verte highway. The following description of the mine is compiled largely from the work of Murray and Howley (1881), Douglas et al. (1940), Watson (1947) and Purcell (1961).

HISTORY OF DEVELOPMENT: It has been reported that Smith McKay found the Terra Nova prospect in 1857, the same year as the Tilt Cove discovery (Martin, 1983). In a letter to Governor Bannerman dated October 10, 1860, Frederick Gisbourne listed the mineral deposits of Newfoundland that were worked between 1855 and 1860; included in this list was the Terra Nova prospect, which was being exploited for mundic (pyrite) and copper (Martin, 1983). In April of 1860, McKay, C.F. Bennett and other financiers from St. John's and Boston formed the Terra Nova Mining Company. The mine apparently operated full time from 1860 until 1864, and only intermittently thereafter. Murray described the mine workings in 1864 (Murray and Howley, 1881); he reported that the massive sulfide ore was discovered in the stream bed and that five shafts had been sunk "to prove the mine." He noted that only two of these shafts, the Bell (#5) shaft and the #4 shaft, intersected the "metalliferous band." Production was mainly through the Bell shaft. Murray also noted an abrupt termination of the ore body on the south side of the brook, which led him to think that either the ore body twisted sharply or there was an overlap in the local stratigraphy.

Table 8-5: *Assays of samples from Goldenville deposit (from Snelgrove, 1935).*

Sample Number	Location	Kind	Width (in feet)	Gold (per long ton)	Gold (g/t)
G 5	Dump, East Shaft	Grab		25 dwt	29.3
G 6	East collar, East Shalt	Grab		trace	
G 7	10 ft. west of Main Shaft	Channel	3.0	1 oz., 4 dwt, 19.84 gr	34.6
G 8	(continuous section)	Channel	4.5	trace	
G 9		Channel	1.5	1 dwt, 7.36 gr	1.4
G 10	East of collar, East Shaft	Channel	4.3	1dwt, 7.36 gr	1.4
G 11	No. 1 Shaft	Selected	8.0	nil	
G 12	Concentrate dump	Grab		1oz., 19 dwt, 4.8 gr	54.7
G 14	Dump, North Shaft	Selected		1 dwt	1.4

Determination of iron in samples G 8 and G 9 was 36.19% and 31.35% respectively. Result of an assay for silver in sample G 10 was nil. Assays were by J.F. Newman.

Character of Samples

G 5 Quartz and sulfides
G 6 Magnetite-hematite-quartz
G 7 Quartz and pyrite
G 8 6 inches pyritized greenstone 4 feet magnetite-hematite-quartz
G 9 Magnetite-hematite-quartz rocks with pyrite
G 10 Pyrite in chlorite schist, quartz and iron oxides
G 11 Pyrite in chlorite schist and quartz
G 12 2/3 iron oxides, 1/3 pyrite
G 14 Pyrite in chlorite schist and quartz

An American firm, the Newfoundland Exploration Syndicate, revived the Terra Nova mine workings in 1901. It formed the Terra Nova Company and proceeded to work the deposit from 1902 to 1906. Subsequently, the property was optioned to the Cape Copper Company Limited, which also controlled the Tilt Cove Mine at this time. There are discrepancies in reports concerning the closing of the Terra Nova Mine, although most reports indicate that the workings ceased in 1915. The reason for the demise of the mine is uncertain, though market conditions and freight charges escalated by World War I are thought to have been major factors. In addition, it has been suggested that ore was being sent indirectly to Germany through European ports, which hastened closure of the mine. A substantial quantity of ore had been stockpiled at the mine by the time it closed. This stockpile caught fire in 1946 and was heavily oxidized.

In 1938, the Geological Survey of Newfoundland commissioned Geophysical Explorations Limited of Toronto to make geophysical surveys over major copper deposits in Newfoundland, including the Terra Nova Mine. There were no significant results from this work.

Brunsman Mines of Toronto undertook geophysical surveys and surface diamond drilling on the Terra Nova property in 1953 and 1954. M.J. Boylen Engineering Company dewatered the mine in 1960, and carried out underground

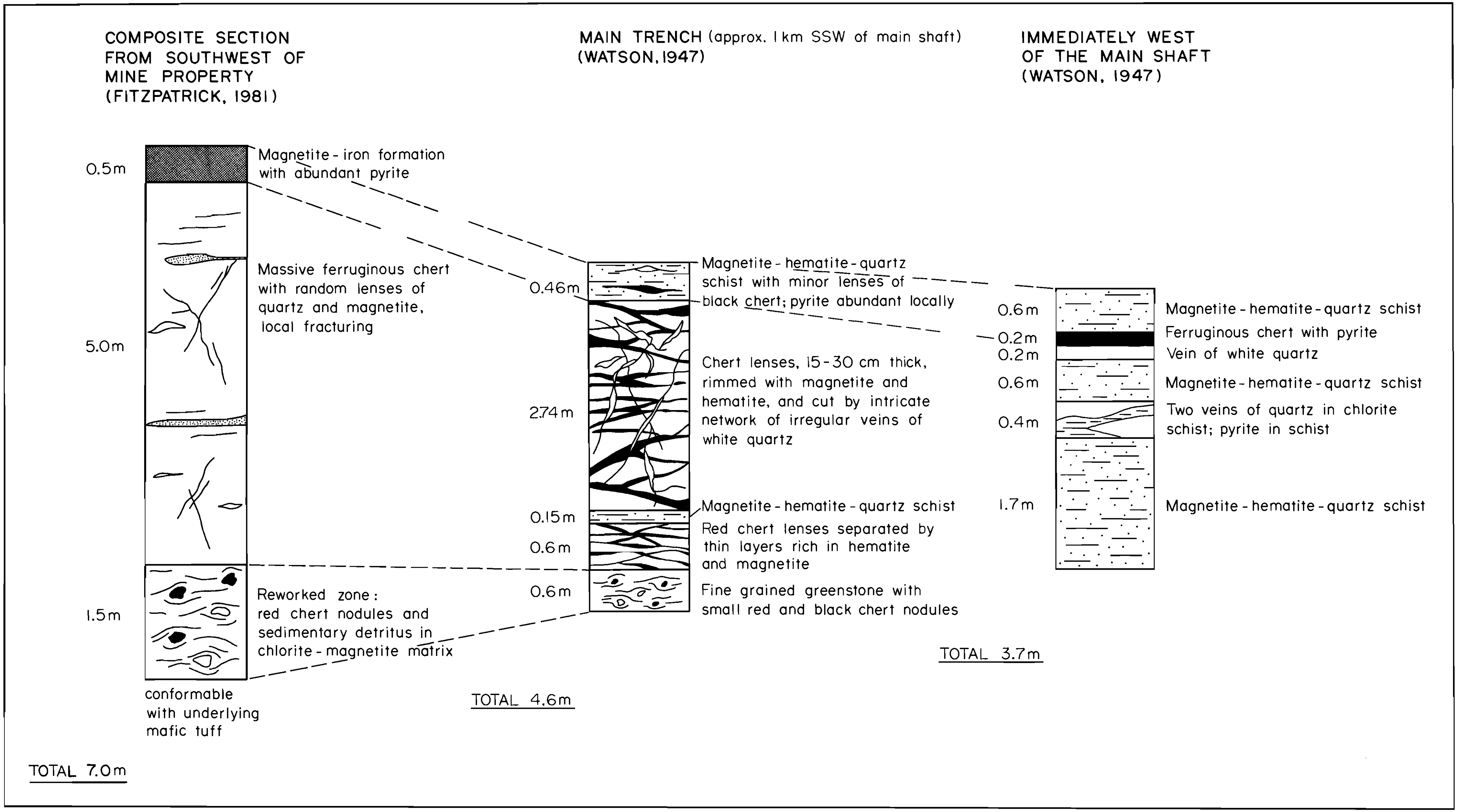


Figure 8-11: *Comparison of sections through the Goldenville mineralized horizon. Gold mineralization is concentrated in the chemical sedimentary zones; quartz veins within these sediments are particularly rich in gold.*

Table 8-6. *Point Rousse Complex metallic mineral occurrencs. (Number refers to Figure 8-5.)*

OCCURRENCE	STATUS	HOST ROCK	MINERALOGY	TYPE	COMMENTS	SELECTED REFERENCES
63. Barry-Cunningham	showing	chlorite schist zones in pillow lava and quartz veins	Au-cp-py sp, bo	vein and disseminated, and massive sulfides	two chip samples from quartz veins assayed 284 g/t Au and 4.2 g/t Au; grab sample of massive ore assayed 11% Cu, 0.10% Zn, 50 g/t Ag, 1 g/t Au; numerous trenches, pits, shafts and adits	Howse & Collins (1978), Frew (1971), Watson (1947), Snelgrove (1935)
64. Mud Pond	prospect	chlorite-quartz schist	py-cp-po	disseminated and stringer, minor massive pockets	drilling has suggested a maximum of 90,000 metric tonnes grading 1% Cu with trace of Au and Ag	Fitzpatrick (1981), Watson (1947)
65. Green Cove	indication	silicified zones in fine grained gabbro	po-cp-py	disseminated	up to 10% observed mineralization; 1 drill hole barren	Neale (1958a), Advocate Concessions Exploration Co. Ltd. (1965), Tuach (1978b)
66. Penny Cove	showing	quartz vein and greenstone	cp-py-sp	vein type and disseminated	mineralization also in silicified and chloritized greenstone; vein barren of Au, trace of Au in greenstone	Watson (1947), Neale (1958a)
67. Northwest of Gillard Pond	indication	?	py, po	?		Neale (1958a)
68. Green Cove Pond	showing	mafic pyroclastic rocks and chlorite schist	py-cp-po	volcanogenic	4 drill holes; best intersection 1.44% Cu over 2 feet	Neale (1958a), Tuach (1978b), Advoc. Concess. Expl. Co. Ltd. (1965)
69. East Pine Cove	indication	?	py	?		Neale (1958a)
70. Green Cove Brook	indication	chlorite schist	py-cp	disseminated		Neale (1958a)
71. West of Goldenville	indication	?	py	?		Neale (1958a)
72. Gillard Pond	indication	chlorite schist	cp-py-sp-gn	disseminated	2 drill holes and trenches	Tuach (1978b)
73. Kidney Pond, north leg of K. Pond	indication	serpentinite	cr-mag	podiform	lenses approximately 1 x 2 inches recorded	Schryver (1959)
74. Deer Cove Occurrence	indication	chert in pillow lavas	py	sedimentary(?)	up to 50% py over area 10 x 2 feet; no gold in assay	Watson (1947), Tuach (1978b)

and surface diamond drilling in 1961, with only minor intersections of mineralization. In the process of dewatering the mine, Boylen workers showed that the Bell shaft, reportedly 640 feet, bottomed at 400 feet and that lower levels had been mined by means of a winze.

The property, as a fee simple mining grant, has since been obtained by B. Phillpot, a Baie Verte entrepreneur, who at the present time utilizes the mine workings as part of a sewage treatment system for his motel, situated directly across the street from the mine.

LOCAL GEOLOGY: The mine site is located in the southern part of the Duck Island Cove sequence (see Chapter V), within the cover rocks of the complex. The mine is in the Baie Verte River valley, which is bounded by highlands to the north and south, formed dominantly of ophiolitic gabbro of the Duck Island Cove and Sisters Cove sequences, respectively. The sequence in the immediate mine area is composed predominantly of interlayered mafic lavas, gray-green volcaniclastic rocks, clastic sedimentary rocks, and irregular masses of serpentinite and diorite (Figure 8-12). It is mainly moderately to steeply southeast dipping, though steep dips to the northwest are present locally; the younging direction of the sequence is uncertain due to the lack of geopedal features and the structural complexity of the area. All of the rocks except the more massive igneous ones are affected by a single, strong, penetrative cleavage. Minor faults and high strain zones are common in the area.

The rocks in the mine area appear to be, in part, of slump origin. The occurrence of irregular masses of serpentinite and diorite within slate, graphitic slate, and fine grained volcaniclastic rocks as well as the somewhat chaotic outcrop pattern (Figure 8-12) are highly reminiscent of olistostromal zones containing large ophiolitic blocks, within the complex and within the nearby Flat Water Pond Group (see Chapter V). In addition, diamond drill logs by M.J. Boylen Company indicate that conglomeratic black slate occurs in the subsurface just to the northwest of the mine site; these rocks are typically associated with olistostromal horizons in the complex and in the Flat Water Pond Group.

MINERALIZATION: The sulfides at the Terra Nova mine occur in two deposits that are oriented nearly vertical; one body outcrops at the surface and the other occurs at depth. Rove (Douglas et al., 1940) compiled old mine maps and formulated the diagrammatic section in Figure 8-13. The upper deposit extends from the surface to the "35 fathom" level[1],

[1] Note that fathom levels were not necessarily equivalent to their depth in fathoms (see Figure 8-13).

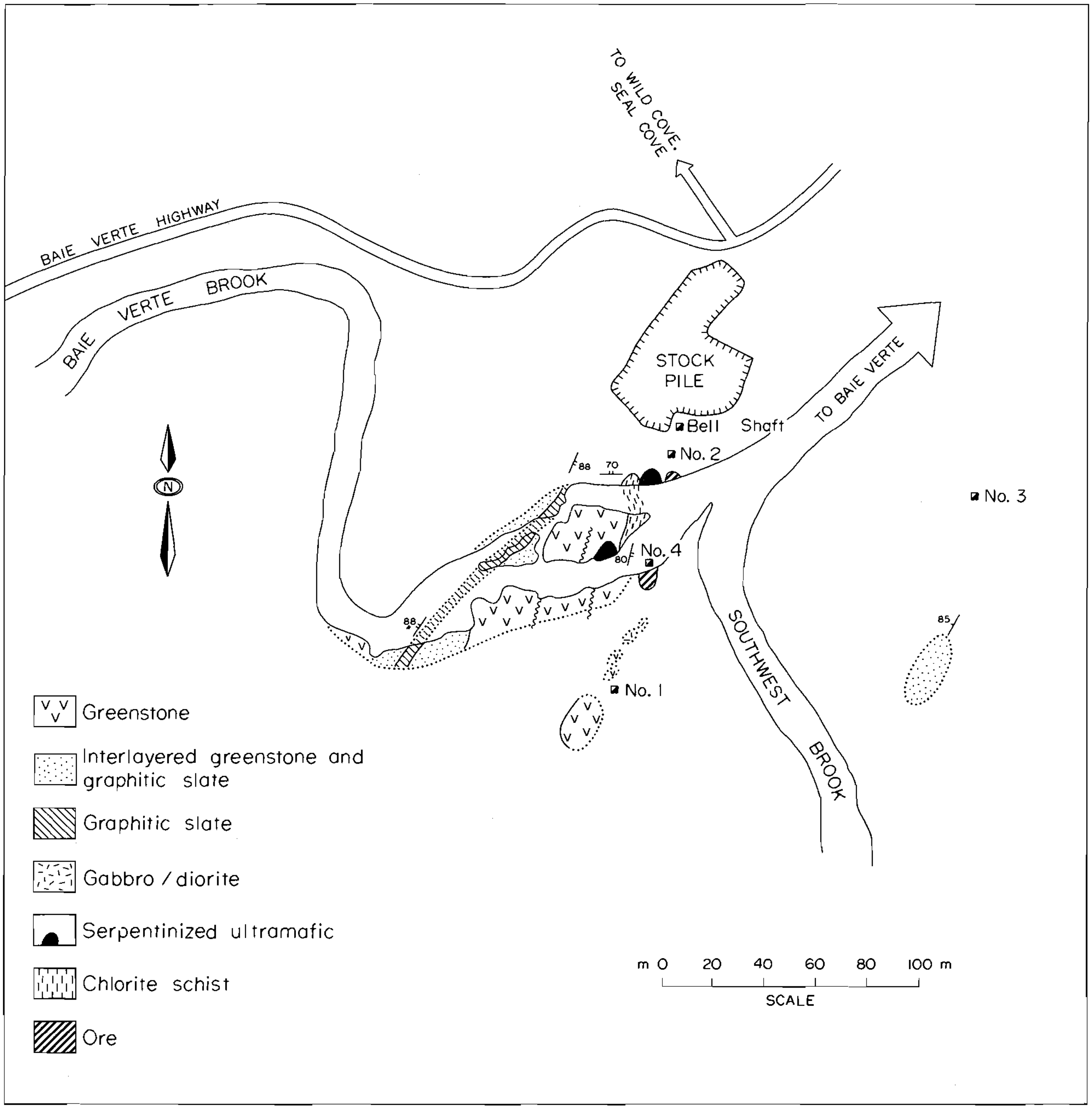


Figure 8-12: *Geology of the Terra Nova Mine area (from Watson, 1947).*

where it apparently pinches out. The lower deposit begins at the "40 fathom" level and extends at least as far down as the "80 fathom" level, though its full extent in this direction is uncertain. Both bodies strike approximately 020° and dip 80° to the southeast. The surface outcroppings of the upper deposit are found near the #4 shaft and on the stream bank below Bell's shaft (Figure 8-12).

The footwall of the deposits consists of chlorite schist, mafic volcaniclastic rocks, graphitic slate, massive greenstone, and local blocks of nodular serpentinite; the hanging wall is formed, in large part, of massive serpentinite, with subordinate chlorite schist. In addition, small masses of diorite and mafic dikes occur in the mine workings. Watson (1947) made special note of the lack of alteration and mineralization in the immediate wall rocks and of the sharpness of the contact between the sulfide zones and host rocks.

The mineralization consists mainly of pyrite, pyrrhotite and chalcopyrite with lesser sphalerite and arsenopyrite; gold and silver are present in minor amounts but are not discernible in polished sections. Murray (Murray and Howley, 1881) reported native copper within calcite veins that cut serpentinite in the #3 shaft. Watson (1947) estimated that pyrite

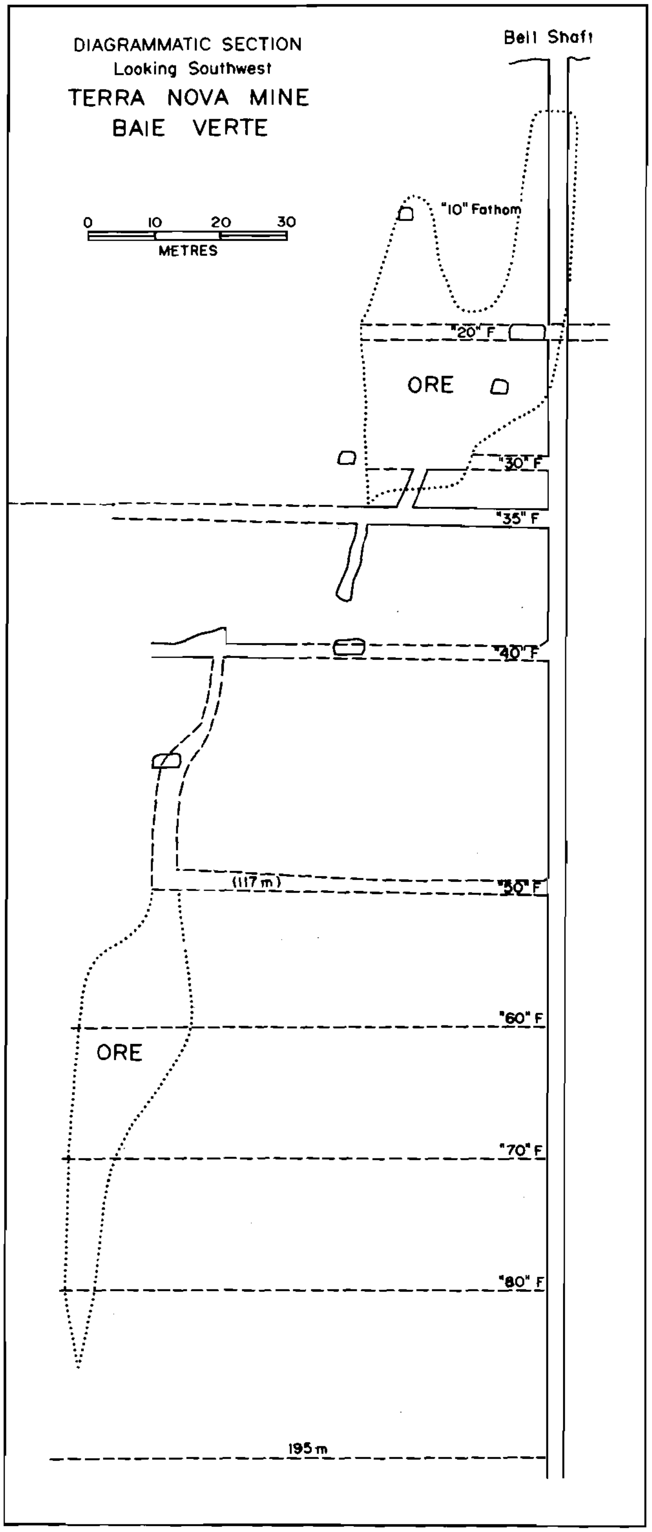


Figure 8-13: *Section through the Terra Nova Mine shaft as compiled by Rove (Douglas et al., 1940).*

forms about 80 to 90% of the deposits. Chalcopyrite and pyrrhotite occur in approximately equal amounts, each constituting 5 to 10% of the sulfides, and sphalerite is generally present in small amounts, though rarely composing as much as 10% of the deposit; arsenopyrite is rare (Murray and Howley, 1881; Purcell, 1961). Murray (Murray and Howley, 1881) noted the colloform and locally banded nature of the ore. Watson (1947) described the colloform nature of the ore as follows:

> In approximately half of the polished surfaces examined, colloform structures were visible in the pyrite. A structure consisting of parallel scalloped bands is the most abundant type but well-developed, concentrically banded, minute spherules were also observed in several places. Although the structures are generally made apparent by differences in the grain size of the pyrite of adjacent bands, there are cases in which pyrrhotite, or less commonly sphalerite or chalcopyrite, occur concentrically interbanded with the pyrite.

ORIGIN OF THE ORES: Previous workers interpreted the Terra Nova deposits as replacement type, within the crests of drag folds associated with local minor faults (Rove, *in* Douglas et al., 1940; Purcell, 1961). Based on more modern concepts and a fuller understanding of the local geology, a different origin for the sulfides is proposed.

The distinct lack of alteration and mineralization in the immediate host rocks strongly suggests that the sulfide deposits have been transported. The apparently olistostromal character of the host rocks supports this concept and further suggests that the transport mechanism was slumping. The sulfides probably originated as stratiform deposits as indicated by their locally banded nature. Also, their source appears to have been ophiolitic, judging from the mafic-ultramafic composition of nearby blocks. Based on this evidence, I suggest that the Terra Nova deposits represent slumped portions of a stratiform massive sulfide that was originally deposited within an ophiolite suite. Detailed studies are necessary to document the distribution of blocks in the mine area as well as to establish if the deposit was in one or more blocks. Since olistostromes within the Advocate Complex may be related to obduction processes, final deposition of the Terra Nova bodies may have been at the time of obduction. Subsequently, the bodies were deformed with the Advocate Complex.

PRODUCTION: Data on production for the Terra Nova mine are very limited. Approximately 200 tons of ore were mined between 1860 and 1864 (Martin, 1983). The only other known production figures are for the years 1902 through 1904, when 38,812 tons of ore were recovered. Based on calculations from old plans and sections of the mine, at least 250,000 tons of ore and waste were extracted from the mine (Tuach, 1978, *quoting* M.J. Collins). Douglas et al. (1940) estimated the grade of the body as 2.41% Cu with significant values of gold and silver.

The stockpile that was left on the property following shutdown was extensively sampled and assayed by Douglas et al. (1940); they reported an average grade of 2.5% Cu, 1% Zn, 0.49 oz./t Au and 0.29 oz./t Ag. This stockpile burned in 1946 and later sampling of the ore revealed a considerable loss of copper (Purcell, 1961). Purcell (1961) also reported assays from samples collected from underground during the M.J. Boylen Company survey of the area (Table 8-7).

Table 8-7: *1961 assays from underground samples at Terra Nova Mine (Purcell, 1961).*

		% Cu	oz./ton Au	oz./ton Ag
Sample No. 1	60 foot chip of massive sulfides from 50 fathom level	4.00	0.05	0.355
Sample No. 2	20 foot chip of disseminated sulfides in chloritic schist from 40 fathom level	0.68	0.01	0.01
Sample No. 3	representative grab of massive sulfides from 10 fathom level	4.20	0.085	0.470

OTHER OCCURRENCES

Five other metallic mineral occurrences have been discovered in the Advocate Complex, and are located in Figure 8-5 and listed in Table 8-8. Four of these are sulfide deposits, each of which is of a distinct type (Table 8-8); the drill hole intersection at Southwest Five Mile Brook appears to have host rocks similar to those at the Terra Nova mine, but the mineralization is a disseminated stringer type deposit. The fifth occurrence is a small podiform chromite deposit within the Flat Water Pond ultramafic body.

Summary of Ophiolitic Metallic Mineral Deposits

The economic metallic mineral deposits of the ophiolite suites on the peninsula are summarized in Table 8-9. These major deposits have been divided into three types, including the Betts Cove type, the Ramber type, and the Goldenville type.

Betts Cove type mineralization is common to all of the ophiolitic units on the peninsula, though the Point Rousse Complex is the only suite lacking such a deposit of economic proportions. The major deposits of this type on the peninsula include the Tilt Cove and Betts Cove bodies and probably the Big Rambler Pond and Terra Nova deposits; most of the less significant occurrences in the area are also of this type. These deposits are characterized by the following features:

(i) simple py – cp – po ± sp mineralogy;

(ii) occurrence in either sheeted dike or pillow lava host rocks, in many cases at the contact of these two members (locally, mineralization in gabbro member), except at the Terra Nova deposit;

(iii) mineralization is in disseminated stockwork and strata-bound massive modes, and is volcanogenic in origin;

(iv) in many cases, zones of mineralization are apparently weak zones, susceptible to later, intense deformation and remobilization.

The Terra Nova deposit is peculiar in setting compared to the others, since it appears to be a leviathan block, transported from its original depositional setting by slumping and mass wastage. Otherwise, the Betts Cove type deposits on the peninsula display the same features as other cuprous-pyrite deposits that have formed at an oceanic ridge (Hutchinson, 1980). The mineralization is closely associated

Table 8-8: *Advocate Complex metallic mineral occurrences. (Number refers to Figure 8-5.)*

OCCURRENCE	STATUS	HOST ROCK	MINERALOGY	TYPE	COMMENTS	SELECTED REFERENCES
75. Priests Prospect	prospect	mafic volcanic and volcaniclastic rocks	py-po-cp-sp	massive sulfide	beneath Roman Catholic school in Baie Verte; drilling indicates 40,000 tons at 1% Cu, 2% Zn, trace of gold and silver	Neale (1958a), Tauch (1978b)
76. Duck Island	indication	quartz vein in shear zones in metagabbro	py-cp	vein	lenses of quartz up to 0.6 m wide and 9 m long	Watson (1947), Neale (1958a)
77. Southwest 5 Mile Brook	showing	chloritic and graphitic schist	py-po-cp-Ni	stringer and disseminated	drill hole intersection; spotted on EM anomaly and geochemical anomaly; over 25 feet 0.49% Cu, 0.13% Ni, 0.02 oz./ton Au, 0.08 oz./ton Ag	Tuach (1978b)
78. Marble Cove	indication	calc-silicate schist	cp	disseminated	very small occurrence	Tuach (1978b)
79. Flatwater Pond	indication	serpentinite	cr	podiform	pod is 2.4 x 0.7 m	Neale & Nash (1963), Neale (1958b)

Table 8-9: *Summary of economic metallic mineral deposits in ophiolitic rocks of the Baie Verte Peninsula.*

OPHIOLITE	NOTABLE FEATURES OF OPHIOLITE	MINE	MINERALOGY	HOST	TYPE
Betts Cove Complex	Thin ophiolite with ultramafic sheeted dikes and lavas; dikes intrude upper part of cumulate section. Ultramafic and mafic lavas of 2 geochemical varieties - lower lavas mostly boninitic, upper lavas mainly tholeiitic.	Tilt Cove	py-cp + po + sp + mag	between sheeted dike and pillow lava members	Betts Cove type (disseminated, stockwork and massive)
			nickel sulfides	in talc-carbonate schist at contact of pillow lavas and ultramafics	remobilized from ultramafic
		Betts Cove	py-cp + sp	sheeted dike and pillow lava members	Betts Cove (disseminated, stockwork and massive)
Pacquet Harbour Group	Only the pillow lava member of suite is present; geochemically the lavas are predominantly high magnesia with some tholeiitic lavas. The ophiolitic member is, at present, inseparable from probable overlying arc-related volcanic and volcaniclastic rocks. Main mineralization occurs near boninite-tholeiite contact and a felsic volcaniclastic zone.	Big Rambler Pond	py-cp	mafic pillow lava and pyroclastic rocks	Betts Cove type (stockwork)
		Rambler	py-cp-sp-po + asp + Au + Ag	mafic and felsic volcaniclastic rocks	Rambler type (massive)
		Ming	py-cp-sp + gn + asp + Au + Ag	mafic and felsic volcaniclastic rocks	Rambler type (massive)
		East Mine	py-cp-po	mafic and felsic volcaniclastic rocks	Rambler type (disseminated)
Point Rousse Complex	Upper member of ophiolite, at present, unseparated from cover volcanic-volcaniclastic sequence. Mineralization most likely at the top of ophiolite. Minor boninitic lavas.	Goldenville	py-Au	quartz veins in ferruginous chert and iron formation	Sedimentary, with remobilization into veins
Advocate Complex	Very highly dismembered ophiolite; cover sequence unconformably overlies ophiolite and is locally composed of olistostromes. All rocks are tectonically interleaved.	Terra Nova	py-cp-po + sp + asp + Au + Ag	apparently an olistostrome with ophiolitic blocks in slate	transported Betts Cove type? (massive)

with boninitic lavas of the pillow member, apparently near the contact between tholeiite and boninite. In the case of the Betts Cove Complex, the boninite apparently forms the base of the pillow member (Coish and Church, 1978); data on the Big Rambler deposit are insufficient for stratigraphic positioning. In contrast, most other occurrences in the world are in tholeiitic pillowed units that form the basal pillow member. More detailed work is necessary to clarify the relationship between the lavas and mineralization and to assess the significance of this association.

Rambler type deposits are peculiar to the Pacquet Harbour Group. They are associated with boninitic lavas, interpreted to be ophiolitic (see Chapter V), and are spatially related to tholeiitic lavas and quartz keratophyre. The Rambler and Ming bodies represent massive deposits of this type and the East Mine is probably their distal equivalent; numerous similar minor occurrences are also found in the group. The Rambler type deposits are characterized by the following:

(i) py – cp – sp + Au + Ag mineralogy in the massive mode; simpler py – cp – po in disseminated deposits;

(ii) immediate host rocks include felsic and mafic volcaniclastic rocks;

(iii) deposits located on periphery of boninitic lava terrane near contacts with tholeiitic lavas.

These deposits are different from typical ophiolitic deposits; instead they are similar to both "Kieslager or Besshi type" deposits and primitive types as outlined by Hutchinson (1980). The tectonic setting of the Rambler deposits is uncertain, but regional implications indicate that the Pacquet Harbour Group occurs in a paleoforearc and the local geology, therefore, displays characteristics transitional between ocean floor and island arc. The localization of mineralization near the boninitic terrane suggests a close affinity to other ocean floor deposits of the peninsula; possibly, the keratophyric rocks represent an unusual manifestation of the ophiolite related to the high magnesian magmatism in the area (see Chapter IX). However, the felsic volcanic rocks at the mine sites may be more akin to an early island arc setting. More detailed stratigraphic and geochemical work is needed to determine the exact nature of the Rambler type deposits.

The Goldenville deposit is the sole known deposit of its type on the peninsula. It is characterized by disseminated gold and pyrite in ferruginous chert and iron formation in a mafic volcanic sequence. It is uncertain whether this deposit is at the top of the Point Rousse ophiolite, or forms part of the cover sequence to the ophiolite. The gold was remobilized into quartz veins in the chert during postdepositional tectonization. This deposit appears to be most akin to Algoma type iron formations, characterized by thin lenses of thinly banded chert, iron oxides and sulfides associated with volcanic rocks of eugeosynclinal affinity (Gross, 1965).

In summary, sulfide mineralization within the ophiolites of the peninsula, though of two forms (the Betts Cove and Rambler types), is in all cases spatially associated with boninitic magmatism. In Betts Cove type mineralization, the deposits were generated in the oceanic crust. The Rambler type deposits are more enigmatic, and may have formed in either an unusual manifestation of oceanic crust or in a volcanic sequence immediately overlying oceanic crust. Gold mineralization in the ophiolites and closely associated sequences is significant in chemical sediments and appears to be related to submarine fumarolic activity.

INDUSTRIAL MINERAL DEPOSITS

Advocate Complex

The Advocate Complex is the only producer of industrial minerals on the Baie Verte Peninsula; it is host to an asbestos mining operation. Limited quantities of the ornamental stone virginite are extracted from altered ultramafic rocks of the complex.

BAIE VERTE MINE

LOCATION AND PREVIOUS STUDIES: The asbestos mine, operated by Baie Verte Mines Inc. since the latter took over from Advocate Mines Limited in 1982, is approximately 6 km north of the town of Baie Verte, along the Fleur de Lys highway. The following description has been summarized from the work of Hutcheson (1965), Blagdon (1974), Bursnall (1975) and Young (1978).

HISTORY OF DEVELOPMENT: The following history of the mine is a compilation and abridgement of Hutcheson (1965) and Blagdon (1974):

Advocate Mines Limited was formed in 1954, with an Ontario charter. In April, 1955, M.J. Boylen obtained a concession on the Baie Verte Peninsula from the Newfoundland Government. Because of the numerous occurrences of copper in this area the first search was mainly for these minerals. However, in August of 1955 the asbestos occurrence was discovered by prospectors and in December of that year Boylen concluded an agreement with Advocate Mines Limited whereby Advocate assumed control of the concession.

The first diamond drilling program was started in November of 1955 and extended through to early 1957 with very encouraging results. During this period approximately 21,000 m of diamond drilling was done and 235 m of underground work.

In September 1958, an agreement was concluded between Advocate Mines Limited and a financing group comprising Canadian Johns-Manville Company Limited, Pafino Mining Corporation, Amet Corporation, and Financière Belge de l'Asbest-Ciment L. A.. Management control was vested in Canadian Johns-Manville Company Limited.

In the same month, a detailed examination program was begun and completed in December, 1959. This program included some 10,700 m of diamond drilling, deepening of the shaft and approximately 1,300 m of lateral underground work. A small test plant was constructed and a bulk sample of 5,000 tons of ore was processed. The fibre produced was shipped to various manufacturing plants for production line testing.

Ore reserves as determined by this latter program stood at 21,000,000 tons. In August 1960, it was decided to carry out additional drilling in an attempt to increase the ore reserves and this program was completed in April 1961. An additional 11,000 m of drilling was done and ore reserves increased to approximately 40,000,000 tons. Further drilling in 1966-67 increased ore reserves to approximately 67,000,000 tons.

In September 1960, the decision to bring the property into production was made. Originally scheduled at a rate of 3,000 tons per day, this was increased to 5,000 tons per day. Construction which was started in mid-1961, was completed during the early summer of 1963. On June 30, 1963 the first ore was processed through the Mill, and by September 1 of that year Advocate Mines was in full production. During the first few months of production, the mine workings were underground. Due to the heterogeneous nature of the host rock, the open pit was subsequently formed [Plate 8-2].

In 1981, Advocate Mines Limited ceased mining because of reported financial difficulties and lack of markets, despite the presence of large reserves projected as adequate for another 15 years of operation. Subsequently, with intervention and assistance from the provincial and federal governments, the property was assigned to Transpacific Asbestos Inc., which formed a subsidiary, Baie Verte Mines Inc.. It reopened the mine in September, 1982.

LOCAL GEOLOGY: The asbestos mine is located in the Advocate ultramafic body, which forms the base of the Duck Island Cove sequence (see Chapter V). The following description of the mine geology is abridged from Young (1978), and accompanied his geologic map of the area (Figure 8-14):

Four basic rock units are recognized in the mine area and are briefly described as follows:

Fleur de Lys Supergroup

Greenschist: This well-foliated, moderately-fractured rock outcrops along the south wall and just west of the perimeter of the west pit.

Advocate Complex

Marble Cove Sequence: This sequence outcrops along the west wall of the mine and from west to east consists of: cream to light gray, moderately fractured, poorly foliated leucocratic schist; moderately fractured, well foliated, greenstone; irregularly foliated mélange consisting of banded mylonite, mylonitic metabasalt and contorted graphitic shale with frequent small to medium sized saucer to egg-shaped blocks of serpentinite and gabbro.

Central Ultrabasic Sequence: This serpentinized peridotite (harzburgite) mass consists of a waste cap in the center made up of very competent lightly to moderately serpentinized rock. Serpentinization and fracturing increase outwards from this core grading into increasingly sheared ultramafic with frequent picrolite slip faces. A highly sheared "fish scale" zone defines the contact on all sides with the country rocks, and this zone thickens with depth, encasing the unsheared material at the center. Chrysotile asbestos ore is largely confined to a basin-shaped zone surrounding the core 'waste cap' within moderately to highly fractured serpentinized peridotite adjacent to the sheared rock.

The eastern contact of the ultramafic body is defined by a partially recrystallized serpentine mylonite with frequent large tabular blocks of rodingitized gabbro forming what has been misnamed the "Dyke Zone".

Cover Sequence: Moderately fractured, mafic volcanics with minor diabase and apparently pillowed rocks in the west pit are more intensely fractured than the rocks occurring along the east wall of the north pit. This sequence may be unconformable on the ultramafic (see Advocate Complex).

STRUCTURE

The most prominent structural features are the shear contacts which surround the ultrabasic body and dip towards it at 90 to 55 deg.. Numerous cross faults cut the ultrabasics and influence variations in the ore body. Faulting is common in the non-ultrabasic country rocks as well, and most rock units are tectonically emplaced against each other.

Plate 8-2: *View of the Advocate pit from the Fleur de Lys road, July, 1978.*

Jointing within the mine is complex and the patterns vary considerably. All major joint sets show clockwise swing of their dip direction from south to north corresponding with the slightly arcuate shape of the ultramafic outcrop.

The most prominent joint system in the ultrabasics is a steeply inclined north trending set which is roughly parallel to a pseudo-bedding. Secondary sets cut this system obliquely in northeast and northwest directions.

One of the most striking structural features of the mine area is the arcuate shape of the Advocate body in plan view. It is the only "flexed" body in the complex and it is the most highly mineralized body; this overall structural feature, thus, may be ultimately responsible for the formation of an environment favorable for asbestos mineralization (see below).

MINERALIZATION: Descriptions of the ore are brief; the following is taken from Young (1978):

The asbestos ore zone is in the form of an elongated basin on which the east and west limbs dip 70 deg. inwards while from the north and south ends, the ore plunges beneath the waste cap at approximately 35 deg.. It is some 1200 meters long, 400 meters wide and of chrysotile crossfiber veins which average 6 to 8 mm in width although veins as wide as 25 mm or more are common [Plate 8-3].

ORIGIN OF THE ORE: Detailed analyses of the ore and its origins are lacking. The origin of asbestos ore in comparable deposits in southern Quebec has been discussed by Riordan (1975). He viewed the process as depending upon an extensive fracture system to act as a conduit system for serpentinizing fluids. Riordan (1975) considered the deposition of ore to involve both wall rock replacement along fractures and fracture filling by mineralizing fluids. During serpentinization processes, wall rock along fractures is gradually replaced by colloidal material. The removal of silica and magnesia from the host rock accompanies this process; these materials accumulate in fracture fillings and are deposited as a vein filling. This vein filling eventually crystallizes as either picrolite or chrysotile (Riordan, 1975).

At Baie Verte Mines, both an extensive fracture system and extensive serpentinization are evident. The key factor in the formation of this deposit appears to have been the arcuate shape and position of the host body in the "nose" of the Baie Verte Flexure; this tectonic situation, where the ultramafic body was molded and buckled against the Fleur de Lys Belt, is most likely responsible for the generation of the fracture system. Numerous other ultramafic bodies on the peninsula apparently lack the highly fractured and disturbed nature of the Advocate body; I attribute this to their tectonic position along the "limbs" of the flexure. A cursory review of deposits in the Quebec serpentinite belt (described by Riordan, 1975) indicates that the major deposits there are also located in the crooks or doglegs of the serpentinite belt (e.g. at Asbestos, Jeffrey). I suggest that, as at Baie Verte, the localization of the deposits along these flexures is in response to the development of highly fractured serpentinite bodies.

PRODUCTION AND FUTURE POTENTIAL: During the mining by Advocate Mines Limited from 1963 to 1981, approximately 33,500,000 tons of ore was milled and 1,066,802 tons of fiber was produced. Baie Verte Mines Inc. produced 4,316 tons of fiber in 1982 and has projected the production of 50,000 tons of fiber in 1983.

The future of asbestos mining at Baie Verte depends on the success of the new operator in marketing its asbestos. Proven reserves available for mining are about equal to the total mined to date, so mining can continue for another twenty years if the operation remains economically feasible.

Other Occurrences in the Ophiolites

Numerous other occurrences of asbestos occur in ultramafic rocks of the ophiolite suites, as well as other industrial minerals, including talc and virginite. These occurrences are compiled in Table 8-10. Most asbestos showings are located

Plate 8-3: *Typical ore-grade asbestos from the east pit of the Advocate Mine.*

in the Advocate Complex and are of minor significance, though the West Pond area and Micmac Lake prospect may have potential for further development. Talc zones occur in ultramafic rocks of the Point Rousse Complex and Betts Cove Complex; mineralization is most abundant at the Trimms Brook prospect in the Point Rousse Complex. The virginite occurrences are found in ultramafic rocks of the Advocate and Point Rousse Complexes (see Chapter V). The virginite at Middle Arm Brook is being quarried for use as an ornamental stone by the Rock Shop of Springdale. Kidd (1974) estimated that virginite forms up to 40% of the ultramafic rocks along the Baie Verte Road Fault system between Micmac Lake and Flat Water Pond. In addition, there is a small occurrence in ultramafics of the Point Rousse Complex along the Scrape Thrust, as well as minor virginite in other fault-bounded ultramafics of the complex.

VOLCANIC COVER ROCKS

The metallic and industrial mineral deposits of the Snooks Arm, Flat Water Pond, Micmac Lake and Cape St. John Groups are summarized in Table 8-11 and Figure 8-5. Sulfide deposits within the Snooks Arm Group are disseminated in sedimentary rocks and lavas and dikes: the only occurrence of an industrial mineral, asbestos, in the cover sequence is in the Balsam Bud Cove Formation.

Mineral occurrences in the Flat Water Pond Group are mainly of stringer and disseminated modes associated with felsic volcaniclastics, although some of these bodies are associated with chloritic and graphitic schist. Massive deposits within the group appear to be stratiform and in mafic volcaniclastic rocks. Although olistostromes are common throughout the group, massive deposits similar to the Terra Nova Mine have not been reported; an apparent block of iron formation in the Kidney Pond conglomerate along the highway approximately 1 km south of Flat Water Pond revealed no gold in an assay done for the present study (assay by A. Lee, government geochemist).

Only minor pyritic occurrences, mostly disseminated, have been found in the late volcanic arc cover rocks of the Micmac Lake Group and the Cape St. John Group. Squires (1981) described the most significant of the Cape St. John occurrences (Beaver Cove Pond) as pyrite and chalcopyrite remobilized from mineralized Ordovician rocks; however, he noted that some disseminated pyrite in nearby Cape St. John chert indicates primary Siluro-Devonian sulfide mineralization.

Overall, the showings in the volcanic cover sequences appear to be insignificant compared to deposits in the ophiolitic rocks. However, each of the units in the cover has potential for different types of deposits. The Snooks Arm Group has a potential for volcanogenic sulfide deposits; pyrite in sediments of the group indicates potential for larger stratabound deposits of this type. Also, other early volcanic arc sequences in central Newfoundland are host to significant sulfide deposits (see Dean, 1978), thus enhancing the potential of the Snooks Arm terrane. The olistostromal nature of much of the Flat Water Pond Group makes it an interesting and a sporting target. The olistostromes appear to have been shed from westward moving ophiolites; hence, there is potential for any of the ophiolite-type mineral deposits in the Flat Water Pond olistostromes. This concept is enhanced by the Terra Nova massive sulfide deposit within an olistostrome of the Advocate Complex (probably equivalent to the Flat Water Pond olistostromes). In addition, the presence of submarine felsic and mafic volcanic rocks indicates a potential for volcanogenic sulfide deposits. The subaerial late volcanic arc sequence including the Micmac Lake and Cape St. John Groups may be worthy of exploration for ancient fumarolic deposits; fluviatile conglomerate and sandstone of these units have potential for uranium deposits. Definitely, the potential of these units is high. Since exploration interest in these units has been low, in the past, renewed exploration activity could yield positive results.

INTRUSIONS IN THE BAIE VERTE BELT

Intrusive rocks of the Baie Verte Belt have not been a major target for mineral exploration. However, mineralization has been reported from both the older and the younger intrusive rocks, with more attention having been directed toward the older Dunamagon and Burlington bodies. Interest in the older intrusions has stemmed largely from high background levels of uranium in stream and lake geochemistry surveys of both units (Davenport et al., 1976, 1981) and from anomalous molybdenum values in the Dunamagon Granite in the same survey. Tuach (1978) reported local disseminated molybdenite mineralization in the Dunamagon Granite near its west end and also near Pacquet Harbour. Exploration work has failed to reveal any significant uranium mineralization in either body.

The Burlington Granodiorite appears to host other sulfide mineralization; mineralized boulders of the granodiorite were reported by Livingston (1942) and were noted during the present program. A large angular boulder (2 m by 1.3 m by 0.6 m) of vein quartz with small fragments of fine grained granodiorite, typical of the Burlington marginal phases, contains signficant stringers and veins of molybdenite and chalcopyrite (Livingston, 1942). It is located approximately 5 km south of Rambler Mines on the brook connecting the two ponds immediately north of Gull Pond (Figure 1-1). A grab sample taken during the present study assayed 0.8% MoS_2 and 0.26% Cu. A second boulder, found approximately 2 km north of

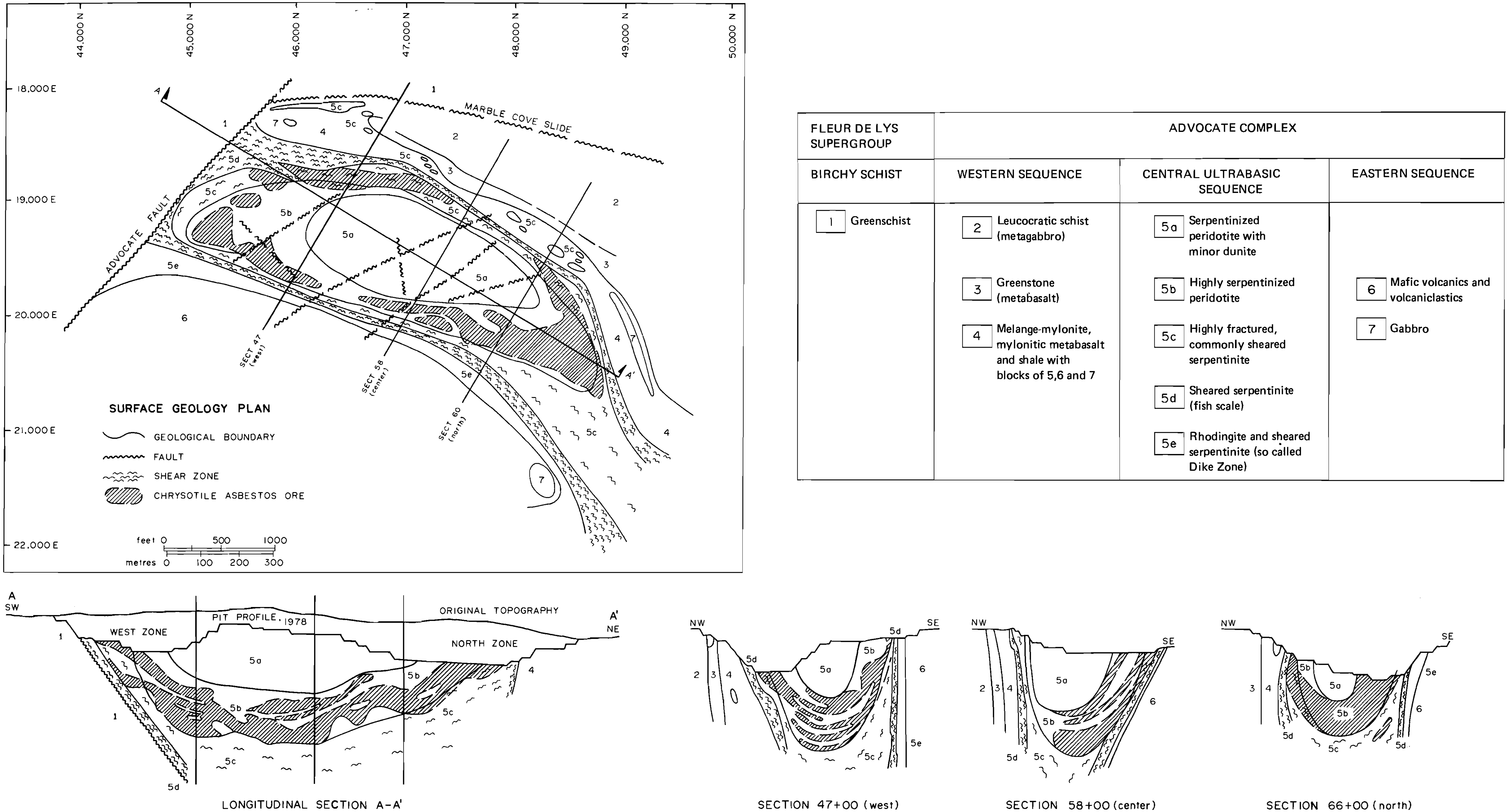


Figure 8-14: *Geology of the Advocate Mine (from Young, 1978). Same scale for all figures; no vertical exaggeration.*

Table 8-10: *Industrial mineral occurrences in ophiolitic rocks of the Baie Verte Peninsula. (Number refers to Figure 8-5).*

STRATIGRAPHIC UNIT	OCCURRENCE	COMMODITY	STATUS	COMMENTS	SELECTED REFERENCES
	80. Flat Water Pond No. 2	asb	indication	one of several small occurrences on a map by Evelegh	Evelegh (1975)
	81. Flat Water Pond No. 3	asb	indication	veins up to 0.47 cm with partings	Evelegh (1975)
ADVOCATE COMPLEX	82. Middle Arm Brook Rapids	asb	indication	lenticular veins up to 0.45 cm with partings	Evelegh (1975)
ultramafic member	83. Red Cliff Pond				
	84. NE Red Cliff Pond	asb	indication	veins up to 0.47 cm with partings	Evelegh (1975)
	85. Upper Red Cliff Pond				
	86. Micmac Lake North	asb	prospect	vein widths up to 2 cm, up to 0.7 m long; high talc content in fiber. One hole drilled in 1964.	Butt (1964, 1965)
	87-89. Jacobs Lake (Wolverine Pond)	asb	showing	veins range from 1/32 to 1/8 inch wide and less than 2 feet long	Butt (1964)
	90-92. Moose Antler Pond	asb	showing	fiber widths up to 3/8 inch; visible only in float and drill core	Butt (1963)
	93. North Brook	asb	showing	cross fiber up to 5 mm wide over 1.4 m; mineralized area 278 x 870 m. In drill core, fiber development best above intersection with dikes of light gray granite.	Butt (1964)
	94. Indian Brook	asb	showing	one fiber vein 1-3 mm wide over 5 m long; minor fiber in core	Butt (1964)
	95. Gillards Lake	asb	showing	two 1.5 mm wide fiber veins >46 cm long. At least 6 drill holes.	Butt (1963, 1964)
	96. North Pond	asb	showing	fiber in lenticular veins within ladder vein complexes of serpentinite. Average fiber length 7 mm, average vein length 15-60 cm. Fiber zone approximately 0.5 km long. Best fiber 16 mm wide over 1 m.	Dodge (1962)
	97. North Brook	asb	showing	fiber length to 7 mm over 5 m; 6 drill holes	Koski (1961)
	98. West Gillards Lake	asb	indication	cross fiber to 2 mm wide, slip fiber to 5 mm long	Czamanske (1956), Conn (1959)
	99. West Pond and nearby occurrences	asb	prospect	at least 18 drill holes in numerous isolated fiber occurrences. Fiber 5.08-10.16 cm thick.	Evelegh and Kelly (1977)
	100. Middle Arm Brook	vir	producer	quarried sporadically by Rock Shop, Springdale, for ornamental uses. Virginite occurs extensively along Baie Verte Line.	Kidd (1974)
POINT ROUSSE COMPLEX	101. North Scrape Pond	vir	indication	poorly developed virginite along Scrape Thrust	Watson (1947), Hibbard and Gagnon (1980)
ultramafic member	102. Trimms Brook	tlc	prospect	small zone; talc forms 75%, in masses up to 10.16 cm long	Watson (1947)
	103. North Yak Lake	asb	showing	one of many showings noted by Neale in the area	Neale (1958a)
BETTS COVE COMPLEX	104. North Yak Lake E.	asb	showing	small cross fiber veins	DeGrace et al. (1976)
ultramafic member	105. Noble Pond N.	asb	showing	small cross fiber veins	DeGrace et al. (1976)
	106. Red Cliff Pond	tlc	indication	associated with significant amounts of carbonate	Snelgrove (1931)

Table 8-11: *Mineral occurrences in the volcanic cover sequences. (Number refers to Figure 8-5.)*

OCCURRENCE	STATUS	HOST ROCK	MINERALOGY	TYPE	COMMENTS	SELECTED REFERENCES
SNOOKS ARM GROUP						
107. Snooks Arm	indication	shale and minor graywacke, Balsam Bud Cove Fm.	py	sedimentary	small isolated grains discernible	Upadhyay (1974)
108. Round Harbour Rd. No. 3	indication	shale and minor graywacke, Balsam Bud Cove Fm.	py	sedimentary	on strike with Snooks Arm showing	Upadhyay (1974)
109. Round Harbour Rd. No. 1	indication	pillow lava and diabase sills, Venams Bight Fm.	py	disseminated	maximum concentration along fractures	Upadhyay (1974)
110. Round Harbour Rd. No. 2	indication	pillow lava and diabase sills, Venams Bight Fm.	py	disseminated	maximum concentration along fractures	Upadhyay (1974)
111. Round Harbour Pond	indication	diabase sill in Balsam Bud Cove Fm.	asb	vein	approximately 300 cm wide vein	Upadhyay (1974)
FLAT WATER POND GROUP						
112. NW Flat Water Pond No. 5	showing	felsic volcaniclastics	py-bo-cp-gn ± Ag	disseminated	mineralized pods up to 40 feet long; best assay - 7-foot chip sample, 0.14% Cu, 1.58% Pb, 0.05% Zn, 0.38 oz./ton Ag	Tuach (1978b)
113. Six Mile Brook	indication	mafic volcanic and volcaniclastic rocks	py-cp	uncertain		Neale (1958a)
114. East Hill	showing	chlorite-sericite-quartz schist	py-po-cp-sp	stringer and disseminated	between 23.4 and 35.3 m depth assayed 0.5% Cu, 0.19% Zn over 3 m	Tuach (1978b)
115. Flat Water Pond Hwy. 411	indication	felsic schist	py	disseminated		Neale (1958a)
116. Flat Water Pond W.	indication	mafic and felsic volcaniclastics	py	disseminated		Tuach (1978b)
117. Flat Water Pond NW	showing	chert and pillow lava	py-cp	disseminated		Tuach (1978b)
118. NE Flat Water Pond No. 1	indication	mafic tuff	py-cp	massive	bed is 15.2 to 20.3 cm thick with 50% visible py	Tuach (1978b)
119. NE Flat Water Pond No. 2	showing	felsic volcaniclastics	py-cp	disseminated	grab sample 0.60% Cu; Au and Ag not detected	Tuach (1978b)
120. NE Flat Water Pond No. 3	indication	sheared mafic and felsic volcaniclastics	py-cp	disseminated		Tuach (1978b)
121. E. Highway 410 No. 1	indication	felsic volcanogenic sediment	py	disseminated		Tuach (1978b)
122. E. Highway 410 No. 2	indication	felsic volcanogenic sediment	py	disseminated		Tuach (1978b)
123. S. Baie Verte	indication		py			Neale (1958a)
124. 5 Mile Brook N.	indication	felsic volcaniclastics	py	disseminated	10-30% disseminated and stringer py, assayed 1.5 g/t Ag; Au not detected	Tuach (1978b)
125. 5 Mile Brook S.	indication	mafic volcaniclastics	py	massive	30 cm thick bed, assayed for Au - nil	Tuach (1978b)
MICMAC LAKE GROUP						
126. Black Lake Brook	indication	felsic volcanic rocks	py	disseminated		Dean & Strong (1975a)
CAPE ST. JOHN GROUP						
127. La Scie	indication	amphibolite	py-gn			Neale (1958a)
128. Brent's Cove and Big Cove	indication	metafelsite	py			Neale (1958a)
129. Gooseberry Cove	indication	mafic pyroclastic rocks	py			Coates (1970)
130. Beaver Cove Pond	indication	chert, sandstone, mafic lava	py-cp	sedimentary and fault mobilized	py-cp dissem. in chert; sandstone and mafic lava near Betts Cove mineralization are mineralized along faults	Squires (1981)
131. Rattling Brook	indication	felsic pyroclastics	py	disseminated		Neale & Nash (1963)
132. Rattling Brook S.	indication	rhyolite porphyry and felsic pyroclastics	py	disseminated		Neale & Nash (1963)
133. Corner Brook E.	indication	quartz-feldspar porphyry	py	disseminated		Neale & Nash (1963)
134. Corner Brook	indication	quartz-feldspar porphyry	py	disseminated		Neale & Nash (1963)
135. Paddys Brook	indication	quartz-feldspar porphyry	py	disseminated		Neale & Nash (1963)

the first boulder, is also composed of quartz veins and fine grained granodiorite, but the visible mineralization is chalcopyrite and sphalerite; on assay, a part of the boulder yielded 2.64% Cu, 0.03% Zn, 0.06% Pb, 0.007% Mo, and 1 g/t Au. The sample was assayed for tin, but with negative results. Minor sulfide mineralization was reported in outcrop by Neale (1958a) in the area just north of the La Scie highway.

A meager amount of mineralization has been reported from the younger intrusive rocks. The most promising deposit in these rocks appears to be the La Scie Quartz Vein, in the La Scie intrusive suite (Figure 1-1). This vein has been evaluated as a silica resource by the provincial and federal governments (Butler and Greene, 1976); it has been estimated to contain 824,000 tons of material averaging 98.0% SiO_2, 1.2% Al_2O_3 and 0.33% total Fe (as Fe_2O_3). In addition, minor sulfide occurrences have been reported from the Cape Brulé porphyry and the Reddits Cove Gabbro; ilmenite has also been reported from the Reddits Cove occurrence.

It should be stressed, again, that the intrusive rocks of the belt have received only minor exploration attention, though they constitute more than half of the Baie Verte Belt on the peninsula. In addition to such targets as uranium and molybdenum, it may be worth considering new targets in these rocks. For example, the Cape Brulé porphyry contains numerous large xenoliths of ultramafic rock; the contact zones of these rock types may be favorable environments for gold mineralization (R.K. Stevens, personal communication, 1981). In addition, the younger intrusions of the belt, including the Cape Brulé porphyry and its southerly equivalents, may be favorable terranes for tin-tungsten mineralization.

FLEUR DE LYS SUPERGROUP

The Fleur de Lys Supergroup is host to numerous metallic and industrial mineral deposits (Table 8-12, Figure 8-5). The most common metallic mineral deposits include strata-bound sulfides in marble, vein type sulfides with minor disseminated pyrite in greenschist, and pods of rutile in association with the metaclastic rocks. Industrial mineral occurrences include marble, garnet and graphite.

The strata-bound sulfide deposits in marble contain mainly chalcopyrite although minor sphalerite and galena have been noted at two occurrences. They all occur in marble of the Rattling Brook Group; it is uncertain if this marble represents one member or many separate members (see Chapter IV). The most significant of these occurrences is the Hodder Prospect, located less than 1 km northwest of Fleur de Lys. Here, cp – bo – po – py ± sp ± ga mineralization forms stringers and disseminated zones crudely parallel to nearly horizontal layering and schistosity in the impure marble. Locally, the sulfides have been remobilized into fractures and offsets that crosscut layering and schistosity. The best assay of the Hodder mineralization is approximately 17% Cu (Watson, 1947) in grab sample. The origin of this mineralization is uncertain; however, based on its strata-bound nature, it likely predates the regional deformation and metamorphism. It may have been related to rifting of the continental margin or later adjustments of this margin, or it may be detrital in origin. Somewhat similar deposits occur in the Whitebrook dolomite in the Eastern Townships of Quebec (Harron, 1976), though mineralization there is confined to the matrix of carbonate breccias. These breccias appear to be in a depositional setting analogous to that of the marble in the Fleur de Lys Supergroup.

Vein type deposits are common in the Fleur de Lys Supergroup in the area of Fleur de Lys (Parrell, Traverstown, Pigeon Cove) and one occurrence has been noted in the Ming's Bight Group (Clark). The vein type deposits are mostly polymetallic sulfide deposits, all of which contain some galena and chalcopyrite commonly with sphalerite and molybdenite; the Clark occurrence is only known to contain pyrite. The deposits around Fleur de Lys are all in calcite-quartz veins situated along fault zones in the Rattling Brook Group, though at the Parrell prospect two types of veins are present, including dolomite-actinolite and feldspar-quartz. The faults all appear to postdate the regional metamorphism and deformation of the group, since the veins reportedly contain blocks of the country rock and "penetrate into" the host rock (Fuller, 1941). The Clark occurrence is in quartz veins in the Ming's Bight Group.

The Parrell prospect is distinct from the Traverstown and Pigeon Cove deposits as it is a molybdenite deposit, whereas the others are mainly lead-zinc. The detailed geology around the Parrell prospect was outlined by Fuller (1941), who demonstrated that the mineralization there is clearly along fault zones. The Traverstown prospect is located along the east trending Lead Mine Fault (Figure 1-1), and Fuller (1941) reported the Pigeon Cove occurrence to be similar to, though of smaller scale than, the Traverstown. Considering the limited distribution of these deposits near Fleur de Lys and their disposition along late faults, it is possible that they are related to the postkinematic Partridge Point Granite (Fuller, 1941).

Less significant sulfide deposits are found in greenschist of the Birchy Complex; these minor occurrences consist mainly of disseminated pyrite and, locally, chalcopyrite. The recent revelation that the Birchy Complex is ophiolitic (Bursnall, 1975) gives new significance to these showings and the mineral potential of the complex. All of the other ophiolitic units on the peninsula host either active or abandoned mines; this greatly enhances the significance of any occurrences in the complex and indicates that the Birchy Complex may have high mineral potential.

Minor rutile showings are also found in the Fleur de Lys Supergroup. The largest occurs in association with quartz veins on Pigeon Island, White Bay (Betz, 1948; Papezik and de Wit, 1973). Many other smaller occurrences are found in the Old House Cove Group, particularly near the common amphibolites in the group.

Marble in the White Bay Group has been prospected for industrial purposes (*summarized in* DeGrace, 1974). The deposits at Clay Cove and at Purbeck's Cove have both been quarried; marble from Clay Cove was exported to London, England for unknown purposes (Edgar, 1927) and that at Purbeck's Cove has been used locally as a source for lime (DeGrace, 1974). The Bear Cove marble has also been analyzed (see Table 8-12).

Other industrial mineral occurrences in the area include garnet, graphite and serpentinite. The graphite was first noted in schists on Slaughterhouse Cove Brook by Watson (1947) who stated "... the slates almost invariably cleave along a

Table 8-12: *Mineral occurrences in the Fleur de Lys Supergroup. (Number refers to Figure 8-5).*

OCCURRENCE	STATUS	HOST ROCK	MINERALOGY	TYPE	COMMENTS	SELECTED REFERENCES
136. Coachman's Hr.	indication	greenschist, Birchy Complex	py-cp	disseminated	small occurrence	Neale (1959a)
137. South Cove	indication	greenschist, Birchy Complex	py	disseminated	small occurrence	Neale (1959a)
138. Slaughter House Cove	indication	greenschist, Birchy Complex	py	disseminated, lenses	small occurrence	Fuller (1941)
139. Slaughter House Block	indication	biotite-graphite schist, Birchy Complex	gf	disseminated, lenses and layers	small occurrence	Fuller (1941), Tuach (1978b)
140. Clark's "Prospect"	indication	quartz veins in semipelite, Ming's Bight Gp.	py	disseminated vein type	shaft sunk over mineralized zone	Watson (1947)
141. Hodder Prospect	prospect	marble, Rattling Brook Gp.	cp-mag-bo, po-py-ma ± sp + gn	strata-bound stringer remobilized	extensive testing by pits, trenches, shafts and drilling; estimated reserve of 30,480 tonnes with minimum of 2% Cu	Fuller (1941), Grant (1956)
142. Bishie Cove	showing	marble, Rattling Brook Gp.	bo-cp	strata-bound disseminated and stringer	chip sample over outcrop (this report) - 0.66% Cu, 0.24% Zn, 80 g/t Pb, 5 g/t Ag	Fuller (1941)
143. Ford's Point	showing	marble, Rattling Brook Gp.	cp	strata-bound disseminated	chip sample over outcrop (this report) - 0.40% Cu, 0.01% Zn, 50 g/t Pb, 4 g/t Ag	Fuller (1941)
144. Pigeon Cove	indication	calcite and quartz vein in the Rattling Brook Gp.	cp-bo-sp-gn	vein type		Fuller (1941)
145. Traverstown	prospect	vein quartz and silicified breccia	gn-sp-cp-py	vein type	1 sample yielded 7.98% Pb; maximum Pb from many channels = 2.38%, maximum Zn = 1.3%, one sample yielded 0.16	Fuller (1941)
146. Parrell Prospect	prospect	in and near quartz-feldspar veins in Rattling Brook Gp.	mo-po-gn-cp	vein	ore left on dumps = 18 tons at 10% MoS_2, 0.3% Ni and greater than 100 tons at 1.5% MoS_2	Fuller (1941)
147. Ruth's Occurrence	showing	marble, Rattling Brook Group	bo-cp-sp	strata-bound stringer	chip sample across outcrop - 1.04% Cu, 0.15% Zn, 0.46 oz./ton Ag, trace Au	Tuach (1978b)
148. Jim's Occurrence	indication	marble, Rattling Brook Group	bo-mal	strata-bound stringer	noted across approximately 1 m	Tuach (1978b)
149. Seal Cove Road	indication	graphite schist, Rattling Brook Gp.	gn(?)		reported on old maps of M.J. Boylen Eng. Ltd.	Tuach (1978b)
150. W. Coachman's Cove	indication	quartz-muscovite schist, Rattling Brook Gp.	gt	porphyroblasts	up to 30% garnet, individual crystals to 1 cm diameter	Tuach (1978b)
151. Pigeon Island	indication	Pigeon Island Fm., White Bay Group	ru, gt	veins, porphyroblasts	up to 1 m wide lenses with up to 40% rutile	Papezik & de Wit (1974)
152. Bear Cove	showing	marble, White Bay Group	marb	strata-bound	analyses - 43.50% CaO 8.22% MgO 0.70% SiO_2 2.56% Fe_2O_3 0.21% Al_2O_3 0.03% S	DeGrace (1974)
153. Purbeck's Cove	past producer - dormant	marble, White Bay Group	marb	strata-bound	quarried for time prior to 1941. Two analyses - 50.71% 54.60% $CaCO_3$ 1.00% 0.70% MgO 2.35% 1.20% SiO_2 0.32% 0.90% Fe_2O_3 0.83% 0.90% Al_2O_3 0.04% 0.00% S	DeGrace (1974)
154. Clay Cove	past producer - dormant	marble, White Bay Group	marb	strata-bound	inclined railroad track from quarry to pier	Edgar (1927), DeGrace (1974)
155. Fleur de Lys	indication	serpentinite	ant		facing stone	Fuller (1941)

foliation plane containing considerable graphite in preference to others, [thus] one is apt to greatly overestimate the amount of graphite present." The garnets occur in garnetiferous schist of the Rattling Brook Group; they are generally fractured, but may be useful as abrasive material. The serpentinite occurs in Fleur de Lys as pods and lentils within the Rattling Brook Group; Fuller (1941) first noted that the rock was mainly antigorite and may be useful as a facing stone.

The Fleur de Lys Supergroup has received very little mineral exploration attention in the past, mainly because it has been ominously labelled a "metamorphic terrane." The potential of such terranes is beginning to be realized, as a better understanding is attained of the protoliths to such belts. Similar metamorphic terranes in the Appalachian-Caledonide System are hosts to major deposits. Most notably, the Ducktown sulfide deposits in Tennessee (Addy and Ypma, 1977) occur in rocks indistinguishable from those of the Old House Cove Group (personal observation). The Aberfeldy barite-sulfide deposit in Scotland (Coats et al., 1980) occurs in rocks very similar to the Rattling Brook Group. Clearly, the Fleur de Lys Supergroup has potential for such deposits.

WILD COVE POND IGNEOUS SUITE

There are no reported mineral occurrences within the suite; however, reconnaissance stream sediment surveys by Northgate Exploration in 1978 revealed minor molybdenum and uranium anomalies in the southeastern part of the batholith. Considering the composite nature and lack of exploration in the suite, it has potential for many different types of mineralization. The most prominent targets are the two-mica granitoids and alaskites for tin-tungsten deposits, and the pegmatite-rich areas for molybdenum mineralization. Also, vein type deposits may be expected around the periphery of the suite, much the same as those deposits spatially associated with the Partridge Point Granite. Extensive zones of obvious alteration commonly associated with porphyry deposits were not noticed in the suite.

SUMMARY OF MINERALIZATION AND EXPLORATION TARGETS

The following points summarize the major concepts and possible exploration target areas outlined in this chapter:

(i) The highest potential in the area is within the ophiolitic rocks, where economic grade deposits have been discovered in the past. A spatial association has been noted in this study that may belie a genetic link between boninitic (high magnesian) lavas and Betts Cove type sulfide mineralization. This is particularly obvious in the Betts Cove Complex and Pacquet Harbour Group.

(ii) The peculiar Rambler type sulfide deposit is polymetallic and occurs close to submarine felsic pyroclastics that are near the contact of tholeiitic and boninitic lavas. It is uncertain whether this is a modified ophiolitic deposit or a volcanic cover deposit.

(iii) Iron formation in the Point Rousse Complex hosts the Goldenville gold mineralization; the gold was deposited in the chemical sediments and remobilized during later tectonism.

(iv) Major deposits such as the sulfides at Terra Nova may have formed huge slump blocks within olistostromal horizons; such deposits may occur in the Advocate Complex and Flat Water Pond Group.

(v) Extensive asbestos development resulted from hydrothermal activity within ultramafic rocks of the peninsula, producing at least one deposit of economic size and grade, the Baie Verte deposit. The ultimate control on the development of the fracture system necessary for such a deposit may have been the position of the ultramafic body in the 'crook' of the Baie Verte Flexure.

(vi) Showings within the volcanic cover sequences indicate that sulfide generation and deposition were active during the volcanism; this implies that larger deposits may be present in these rocks. Sulfides in sediments of the Snooks Arm Group suggest that sedimentary sulfide deposits are significant in the clastic sequences.

(vii) Emplacement of the older intrusive rocks of the Baie Verte Belt, i.e. the Dunamagon Granite and Burlington Granodiorite, was accompanied by molybdenite mineralization; the Burlington Granodiorite also contains minor copper-lead-zinc mineralization. The younger intrusions of the belt may be prospective for gold in areas where the Cape Brulé porphyry intrudes ultramafic rocks; they may also be good targets for tin-tungsten deposits.

(viii) The Fleur de Lys Supergroup contains strata-bound copper deposits and vein type sulfide deposits. The copper deposits appear to be confined to extensive bands of marble in the Rattling Brook Group. Vein type deposits are found near the Partridge Point Granite and appear to be related to it. Similar deposits may be expected to occur around the periphery of the Wild Cove Pond Igneous Suite.

(ix) The ophiolitic Birchy Complex may be host to deposits similar to those found in other, less tectonized, ophiolites on the peninsula.

(x) The Fleur de Lys terrane is a likely target area for deposits analogous to those at Ducktown, Tennessee (sulfides) and at Aberfeldy, Scotland (barite).

(xi) The Wild Cove Pond Igneous Suite may have potential tin-tungsten or sulfide deposits.

The numerous mines and major prospects on the Baie Verte Peninsula attest to the mineral wealth of the area. Exploration concerns would be wrong to assume that the area has been "milked dry." It is clear that there is room for expansion of targets using newly developed ideas about mineralization and new exploration technology. The Baie Verte Peninsula still has significant potential for the discovery of new economic mineral deposits.

CHAPTER IX
REGIONAL SYNTHESIS

"There is, of course, a correct way to put all of the pieces together, but it is very elusive."
Marshall Kay (1976)

INTRODUCTION

This final section summarizes and attempts to integrate the major concepts developed in preceding chapters. It is the most interpretive section of the report. My approach is to provide one internally consistent model for the geological evolution of the area with the understanding that it does not represent the only possibility. At the end of this section, I present a summary of the major findings of this study plus a list of outstanding geological problems on the peninsula. Hopefully, viewing the area as presented in this section will help to generate new paths of geologic thought.

THE MODEL

The model (Figure 9-1) is based almost entirely on the geology of the Baie Verte Peninsula, though certainly the processes responsible for the formation and juxtaposition of rocks in the area operated at a much larger scale. Many previous workers have demonstrated the similarity of rocks on the Baie Verte Peninsula to those in analogous positions elsewhere in the Appalachian-Caledonian system. Correlations have been made between the rocks and structures of the peninsula with those of Ireland and Scotland, including comparisons of the Cabot Fault with the Great Glen Fault of Scotland (Wilson, 1962), and of the Fleur de Lys Supergroup with the Dalradian of Ireland and Scotland (Church, 1965a,b, 1969; Neale and Kennedy, 1967; Dewey, 1969b; Kennedy et al., 1972; de Wit, 1972; Kennedy, 1975a,b). These correlations were summarized by Williams (1978b). Similarities also have been noted between Baie Verte Peninsula geology and rocks in the Eastern Townships of Quebec (Church, 1965a,b; Dewey, 1969b; Bird and Dewey, 1971; Kennedy, 1971). More recently, these striking correlations were outlined in detail by St-Julien et al. (1976) and Williams and St-Julien (1978, 1982) and summarized by Williams (1978b). Other correlations have been suggested with rocks in Spitzbergen and eastern Greenland (de Wit, 1972). Thus, the scale of processes that formed rocks of the peninsula, as well as those in analogous positions elsewhere in the Appalachian-Caledonide Orogen, was on the order of thousands of kilometres; however, the processes were not necessarily consistent over the full length of the orogen.

In reconstructing the evolution of Paleozoic lithotectonic systems, the modeller is susceptible to unavoidable pitfalls. Foremost amongst these is our scant knowledge of Paleozoic tectonic processes, which necessitates that the model be based on the premise that these processes were the same as present-day processes. Also, the basis of the model, the rock record, is incomplete due to both inherent depositional gaps in the stratigraphic sequence and tectonism and erosion of the rock record. Thus, perhaps it is best to view the following model of the Baie Verte area as a mental construct to help grasp overall lithotectonic relationships of rocks preserved rather than to consider it a totally accurate account of the evolution of the area. For simplicity, the model is depicted in Figure 9-1 in only two dimensions, thus neglecting a major feature of the peninsula, the Baie Verte Flexure, which is however discussed in this section.

PRE-TACONIC (Figure 9-1, I)

Rifting of the ancient North American continent at about 600 Ma (Williams and Stevens, 1974) gave rise to an oceanic tract, Iapetus. The North American continental mass to the west of the rift developed a typical passive type continental margin (Williams and Stevens, 1974) that comprised rift facies volcanics (protoliths of Oody Mountain and Garden Cove Formations, Rattling Brook Group and Fleur de Lys amphibolites) and clastic rocks derived from the continental area (protoliths of most of the East Pond Metamorphic Suite and the Fleur de Lys Supergroup, except the Birchy Complex). The portion of the ancient North American continent that was east of the rift has yet to be identified. The shape of the rifted margin was most likely orthogonal, as reflected by the present Baie Verte Flexure; this shape was probably inherited from transform faults formed during rifting. By approximately 500 Ma (Figure 9-1, I), the ancient North American margin was well established and Iapetus had likely reached oceanic proportions. The location and orientation of the Iapetan spreading ridge is uncertain, since most of the oceanic realm was later consumed along a destructive margin. The ophiolitic Bay of Islands Complex of western Newfoundland and possibly the ophiolitic Birchy and Advocate Complexes had been generated by about 500 Ma. Subduction of Iapetus had also probably begun by then, since Tremadocian conodonts have been recovered from island arc rocks (Catchers Pond Group) in central Newfoundland (Stouge, 1980).

During this phase of evolution, rift-associated and some ophiolite-associated mineral deposits formed; these include deposits such as the Hodder prospect and possibly sulfide deposits in the Birchy and Advocate Complexes.

SYN- TO LATE TACONIC (Figure 9-1, II)

The Taconic Orogeny was marked by an westward structural polarity, with the western edge of Iapetus being obducted onto the ancient continental margin (Figure 9-1, II) (Stevens, 1974). During this event, a dynamothermal aureole probably formed beneath the Bay of Islands - Advocate Complexes oceanic slab, while oceanic crust of the Birchy Complex, and possibly the Marble Cove sequence of the

Advocate Complex, as well as the clastic Fleur de Lys Supergroup protoliths were subducted beneath the continent. The formation of this imbricate wedge of oceanic crust led to the generation of ophiolitic mélanges in front of (to the west) and beneath it. Probably during later stages of formation of this wedge, ophiolitic detritus was shed eastward onto the "backs" of other ophiolitic slices, thus forming the conglomerates of the Flat Water Pond Group and Advocate cover sequence.

$^{40}Ar/^{39}Ar$ cooling dates on metamorphic minerals in the basal aureole of ophiolites on the west coast of Newfoundland (Dallmeyer, 1977) indicate that subduction, imbrication and aureole formation proceeded until at least 470 Ma. A 488.6 $^{+3.1}_{-1.2}$ U/Pb zircon age (Dunning and Krogh, 1983) on gabbro from the Betts Cove Complex as well as circumstantial evidence for the age of the Pacquet Harbour Group (see Chapter V) indicate that new oceanic crust was being formed in the fore-arc area at the same time as subduction and formation of the aureoles. The oceanic crust formed in this environment contains high magnesian lavas as the upper member of the ophiolite suite. The generation of these unusual ophiolitic lavas may be directly related to subduction, since major faults forming during subduction could either have acted as conduits for sea water to reach the mantle or have allowed entrance of water-laden sediments into the mantle. Such hydrous conditions would lower the solidus of a melt, allowing the formation of a high magnesian magma. The heat source for such a melt is uncertain, but may have been a leaky transform fault, since the geometry of the Baie Verte Flexure suggests that a transform fault bounded the peninsula to the north. Alternatively, some form of spreading ridge may have been present in the fore-arc area. The relatively small amount of magnesian lavas in the Point Rousse Complex compared with the Betts Cove and Pacquet Harbour ophiolites indicates that the Point Rousse Complex probably formed during the earlier stages of subduction, whereas the latter units formed at a later stage, when more water had reached the mantle. Since the Betts Cove lavas are more magnesian than those of the Pacquet Harbour Group, they probably represent the products of a more mature melt and, thus, are probably slightly younger.

Island arc activity continued through the Taconic events and was marked by deposition of the Snooks Arm Group and cover sequences on the Pacquet Harbour Group and Point Rousse Complex, as well as intrusion of the Burlington Granodiorite into the Pacquet Harbour Group. The unique rare earth element geochemistry of the Snooks Arm Group (Jenner and Fryer, 1980) may be an inherited feature from the preceding, high magnesian regime.

Rocks of the ancient North American continent and the Iapetan oceanic crust were structurally telescoped and juxtaposed during the Taconic to form the embryonic Baie Verte Line. The shape of the line reflects the original shape of the interface of these two realms. The Dunamagon Granite was intruded at about 460 Ma and welded the east-west portion of the line. The source of the granite is uncertain; since it spans the east-west portion of the line, it may be related to strike-slip movements along this structure during the juxtaposition of the continental and oceanic rocks. The ophiolitic rocks and cover sequences appear to have been folded and faulted in latest Taconic times, but the mechanism of this deformation is unclear at this time.

The East Pond Metamorphic Suite and Fleur de Lys Supergroup underwent major deformation and metamorphism during subduction. Also, high strain schists were formed at the margins of the East Pond Metamorphic Suite, and the basement within the suite was remobilized along its contact with cover rocks of the suite.

Mineral deposits which formed during this time included the deposits at Tilt Cove and Betts Cove and the unusual Rambler-type deposits. In the Point Rousse Complex, gold was deposited within chemical sediments closely associated with the ophiolite. Since the Advocate ultramafic body was emplaced in the nose of the Baie Verte Flexure at this time, it is possible that a structural environment suitable for the formation of asbestos developed in the body at the same time. However, formation of the deposit may well have taken place any time up to the end of Acadian tectonics.

PRE-ACADIAN (Figure 9-1, III)

Taconic subduction and the obduction of ophiolites marked the closure of Iapetus. Post-Taconic and pre-Acadian events were marked mainly by isostatic re-equilibration. During the initial isostatic uplift, the tectonics of the Baie Verte area began to reverse polarity to an eastward transport direction. Minor horizontal telescoping probably continued through the Taconic into post-Taconic times due to the impetus of the plates involved. Thus, the Betts Cove Complex and Snooks Arm Group were deformed prior to the ensuing deposition of Siluro-Devonian strata.

Silurian - Early Devonian deposition is marked by the dominantly felsic volcanics of the Micmac Lake Group and the bimodal volcanics of the Cape St. John Group. Deposition of these units was accompanied by faulting and intrusion of calc-alkaline to peralkaline plutons such as the Cape Brulé porphyry, the La Scie intrusive suite, the Middle Arm Ridge suite and the Gull Pond Ridge intrusion. The source of the magma for these units is uncertain. It may have been the result of melting of either the subducted oceanic slab or the lower portion of the subducted North American continental crust (Figure 9-1, III). The bimodal nature of the Cape St. John Group compared to the largely felsic volcanism of the Micmac Lake Group suggests that both mechanisms may have been operative.

Uplift of the East Pond Metamorphic Suite and the Fleur de Lys Supergroup was contemporaneous with the Siluro-Devonian magmatism. Rocks at the present erosional level of these metaclastic units along the western part of the Baie Verte Flexure had cooled below the blocking temperature of argon in hornblende, muscovite and biotite by this time.

Related deformation appears to have been mainly faulting, such as in the Micmac Lake Group and along the margins of the Fleur de Lys Belt. Major transcurrent faults such as the Green Bay Fault may have been initiated at this time.

LATE ACADIAN (Figure 9-1, IV)

Intense Acadian deformation appears to be localized along the southeastern side of the Baie Verte Line where it is characterized by southeastward structural polarity. Along the western limb of the Baie Verte Flexure, deformation appears to be associated with reverse faulting and southeastward imbrication of the Advocate Complex, Flat Water Pond and Micmac Lake Groups, and the part of the Burlington Granodiorite near the Baie Verte Line.

Rocks on the eastern limb of the flexure were overthrust, in part, by the Point Rousse Complex and polydeformed by Acadian events. The mechanism of Acadian deformation is unclear. Deformation may have been either a part of the initiation of a global scale Variscan megashear as envisaged by Arthaud and Matte (1977), or the result of lithosphere detachment and crustal shortening as outlined by Colman-Sadd (1982). Both mechanisms may have been operative and interdependent.

Acadian polydeformation along the eastern limb of the flexure can be viewed as the product of colliding blocks during early movements on transcurrent faults in the megashear model (Arthaud and Matte, 1977). The Baie Verte Belt can be viewed as a structural block, bounded to the south by the right-lateral Green Bay Fault and to the north by the deformed eastern segment of the Baie Verte Line; to the west, the belt is bounded by the western segment of the Baie Verte Line, the Baie Verte Road Fault system. The most obvious movement on this latter system was uplift of the western side, though strike-slip movement may also have occurred. If left-lateral movement had occurred along this system in conjunction with the known right-lateral movement of the Green Bay Fault, then the Baie Verte Belt would have been jammed against the rocks of the Fleur de Lys Belt to the north, thus causing deformation along the contact between the two belts.

Further strain on these rocks may have been caused by overthrusting of the Point Rousse Complex to the south and east. The complex (not depicted in part IV of Figure 9-1 since it lies to the north) was probably originally transported westward of the present Baie Verte Line during Taconic events and may have overlain the Fleur de Lys Belt. With subsequent uplift and unroofing of this belt as well as the possible jamming of the stratigraphic belts by transcurrent faulting, the complex was backthrust over the Baie Verte Line, thus adding to the deformation. Metamorphism in this zone of Acadian polydeformation may be attributable to the interaction of high heat flow in this volcanic arc area and deformation as outlined above.

Reverse faulting along the western section of the Baie Verte Line may be the result of lithosphere detachment (Colman-Sadd, 1982) and eastward advancement of the North American continent over the lithosphere. Alternatively, continued uplift of the continental margin may have caused this back-shuffling of rock units.

Acadian deformation has not been recognized regionally in the main Fleur de Lys Supergroup outcrop belt, but may be represented as some minor late structures. In addition, some K/Ar and $^{40}Ar/^{39}Ar$ dates in the area between the communities of Fleur de Lys and Baie Verte range from uppermost Silurian to Middle Devonian (approximately 360 Ma). These dates are very similar to those on the east limb of the Baie Verte Flexure. Thus, the rocks north of Baie Verte may have originally lain further to the southeast along the east limb of the flexure; they were possibly faulted into their present position during dextral movement along the Little Lobster Harbour Fault in post-Acadian times. The Fleur de Lys Belt was intruded by the Wild Cove Pond Suite and the Partridge Point Granite, late in the Acadian Orogeny.

The cause of Acadian deformation on the peninsula appears to have been more abstract and localized than Taconic deformation and may have involved the interplay of a megashear system and lithosphere detachment, forming many small structural plates with complex structural interrelationships.

Regional deformation appears to have ceased by the Carboniferous as Carboniferous sedimentary rocks at the southern end of the peninsula are undeformed. Locally, though, fault movements may have continued.

SUMMARY OF NEW RESULTS AND OUTSTANDING PROBLEMS

Having reviewed the possible geological evolution of the area, I conclude with a summary of the major results of this report and the most outstanding problems of the area that confront us.

MAJOR RESULTS OF THIS STUDY

1. Mapping of the Baie Verte Peninsula and resultant erection of a formal stratigraphy, with refinements in the definition of local stratigraphic successions.
2. Recognition that large parts of the area previously designated as Grenville basement in the Fleur de Lys Belt (de Wit 1972, 1980) may be highly deformed equivalents of the Fleur de Lys Supergroup.
3. Recognition that the stratigraphy of the Fleur de Lys Supergroup is more complex than previously considered; in particular, that the White Bay Group directly overlies Grenville basement.
4. Determination of lithic and geochemical features of the Wild Cove Pond Igneous Suite.
5. Determination of geochemical features of Fleur de Lys Supergroup amphibolites.
6. Elucidation of the geological setting of the Pacquet Harbour Group; in particular, recognition of its ophiolitic nature, its similarity to the Betts Cove pillow lava member, and its distinction from the Cape St. John Group.
7. Elucidation of the tectonic setting of ophiolitic high magnesian (boninitic) lavas on the peninsula; in particular, the apparent synchronism of subduction beneath the oceanic plate and the formation of the high magnesian lavas on that plate.
8. Recognition that the Baie Verte Line extends into the Pacquet Harbour area, and tectonically separates the Pacquet Harbour and Ming's Bight Groups. This extension is represented solely as an early zone of disruption, whereas the established part of the line, to the west, is delineated by late faults as well as earlier structural zones.
9. Use of the extended Baie Verte Line to define the major structural geometry of the peninsula, including the Baie Verte Flexure.
10. Resolution of the traditional problem of the age of deformation of polydeformed rocks on the peninsula by recognition of the flexure in conjunction with the use of $^{40}Ar/^{39}Ar$ dates. Essentially, rocks on the western limb were deformed mainly in the Taconic Orogeny whereas those on the eastern limb display Acadian polydeformational features.
11. Delineation of a major thrust, the Scrape Thrust, that separates the Pacquet Harbour Group from the Point Rousse Complex. Structural relationships indicate that it is of broadly Acadian age.
12. Establishment of new targets for mineral exploration, including the Fleur de Lys marbles and psammites for massive sulfides, the Fleur de Lys metaclastics for barite, and the boninitic lavas for massive sulfides.

OUTSTANDING PROBLEMS

Stratigraphy

1. Age of the East Pond Metamorphic Suite and the nature of its migmatites.
2. Relative and absolute ages of units in the Fleur de Lys Supergroup.
3. Age of Advocate and Point Rousse Complexes and Pacquet Harbour Group.
4. Age of Flat Water Pond Group.
5. Age of presumed Siluro-Devonian igneous suite.
6. Age of Granby Island Formation.
7. Detailed stratigraphy of the Advocate Complex in the Baie Verte area.
8. Nature of the Pacquet Harbour - Cape St. John Group contact (this may be best sought inland, southeast of the Rambler area).
9. Nature of the Burlington Granodiorite, especially the possibility of its being a composite pluton.
10. Detailed stratigraphy of the Cape Brulé porphyry - how much of this unit is actually intrusive?
11. Stratigraphy of the Middle Arm Ridge Suite.
12. Nature of the Gull Pond Ridge intrusive body, particularly with respect to mafic dikes in the area and the bimodality of some of the presumed Siluro-Devonian rocks.

Geochemistry

1. Analyses of the Pelée Point schist and comparison with the Birchy Complex and other ophiolitic rocks.
2. Further investigation into geochemical differences between the northern and southern portions of the Wild Cove Pond Igneous Suite.
3. Determination of why potassic feldspar is lacking in the Old House Cove Group metaclastic rocks, assuming that these were continentally derived sediments.

4. Analyses of the Advocate Complex, Flat Water Pond Group, Micmac Lake Group, Burlington Granodiorite, Dunamagon Granite and Middle Arm Ridge Suite to determine general character; in particular, to further document some of the peralkaline rocks in the Middle Arm Ridge suite.
5. Further investigation of the genesis of boninites.
6. Determination of petrogenesis of the bimodal Cape St. John Group. A key to this may be the Gull Pond Ridge pluton and mafic dikes in the Middle Arm Ridge area.

Structure and Metamorphism

1. Structural analyses and possible correlation from eastern Western Orthotectonic block into Transition and Paratectonic structural blocks.
2. Documentation of structural correlation across the Fleur de Lys Belt.
3. Uplift and cooling history of the Fleur de Lys Belt north of Little Lobster Harbour Fault and in the area south of Purbeck's Cove.
4. Detailed nature of the Big Brook Slide and further delineation of D_E slides in the belt.
5. Cooling history of the northerly Ming's Bight Group and the Horse Islands Group.
6. Sense and amount of movement on the Baie Verte Road Fault system.
7. Nature of Taconic orogenesis in the Baie Verte Belt.
8. Mechanism of Acadian deformation along the eastern limb of the Baie Verte Flexure.
9. History and role of strike-slip movements in the area, such as those along the Green Bay Fault and the Baie Verte Road Fault system.

Mineral Deposits

1. Nature of Hodder-type deposits.
2. Origin of vein-type Mo and Pb-Zn deposits in the Fleur de Lys Belt.
3. Potential of serpentinite and other local plutonic rocks as facing stone.
4. Nature of Rambler-type deposits; in particular, detailed structural, stratigraphic and geochemical work is needed.
5. Nature of Terra Nova deposit and the potential of olistostromes.
6. Development of new targets such as Cyprus-type deposits in the Birchy Complex, gold in quartz veins in and around ultramafic rocks, and fumarolic deposits in Siluro-Devonian volcanic rocks.

CLOSING REMARK

The breadth of the subject may have caused the reader to lose focus on the objective of this study, which has been to determine the geologic settings of mineral deposits on the peninsula. The results have met this objective in a regional sense, but have generated more questions about the detailed geology than were obvious at the outset. Thus, refinements are now needed. Future efforts at resolving the detailed geology may be toward the present objective or some unforeseeable goal. My only hope is that this work proves competent as an introduction to the geology of the peninsula despite the many stones still unturned.

"...and so there ain't nothing more to write about and I am rotten glad of it, because if I'd a knowed what a trouble it was to make a book I wouldn't a tackled it and I ain't agoing to no more."

Mark Twain

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APPENDIX I

ROCK, MINERAL AND ELEMENT ABBREVIATIONS

ant = antigorite
asb = asbestos
asp = arsenopyrite
bo = bornite
ca = calcite
cp = chalcopyrite
cr = chromite
gf = graphite
gn = galena
gt = garnet
hb = hornblende
mag = magnetite
marb = marble
mal = malachite

mar = marcasite
mlr = millerite
ptl = pentlandite
py = pyrite
po = pyrrhotite
q = quartz
ru = rutile
sp = sphalerite
tlc = talc
vir = virginite

Au = gold
Ni = nickel
Ag = silver

APPENDIX II

COMPARISON OF PREVIOUS WORK AND STRATIGRAPHIC DIVISIONS

Table AII-1: *Comparison of previous work on the Baie Verte Peninsula.*

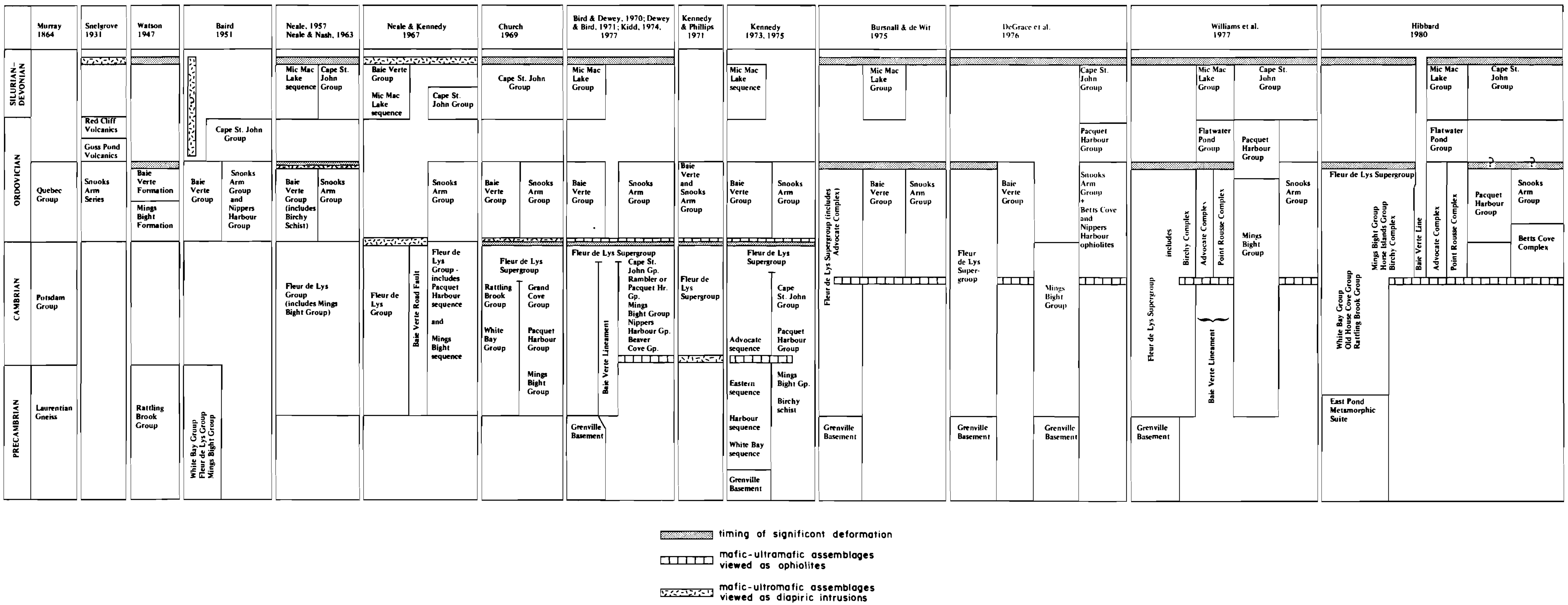

Table AII-2: *Evolution of stratigraphic division of the Fleur de Lys Supergroup.*

	Fuller 1941	Church 1969	De Wit 1972	Kidd 1974	Kennedy 1971, 1975	Bursnall 1975	Hibbard 1980
L. ORD.			White Bay Group White Pt. Fm. Back Cove Fm. Pigeon Island Formation Walkers Cove Formation Stuckless Cove Formation Garden Cove Formation Rocky Pt. Formation Ratding Brook Group Birchy Schist Formation Unit 5 Unit 4 Unit 3 Unit 2 Unit 1			Ratding Brook Group Drovers Formation Flat Pt. Formation	White Bay Group Pigeon Island Formation Rattling Brook Group and Birchy Complex
CAMBRIAN		Rattling Brook Group Birchy Schist Formation Semipelitic formation Middle Arm Brook Formation Western Arm Formation White Bay Group Garden Cove Formation Bear Cove Fm. Westport Fm.		Rattling Brook Group Birchy Schist Formation Semipelitic schist unit	Greywaeke-marble sequence of south-eastern White Bay Advocate sequenee oceanic crust	Coaebmans Formation Long Pond Formation Birchy Group Beaver I Fm. Beaver II Fm. Deep Cove Fm. South Cove Schist oceanic crust	Garden Cove Formation Unseparated schists Old House Cove Group oceanic crust
PRECAMBRIAN	Fleur De Lys Group Pardee Gneiss Birchy Schist Shoal Rocks Gneiss Lngy Formation Starboard Gneiss Headlands Gneiss		Seal Cove Group Southern Arm Formation Middle Arm Formation Middle Arm Pond Metaconglomerate GRENVILLE BASEMENT	Western Sequence Psammite-amphibolite unit Leucocratic quartzofeldspathic gneiss ?	Eastern sequence: Flat Pt. Formation, Birchy Schist, South Cove Schist, Slaugherhouse Schist Harbour sequence: Pardee Formation, Pigeon Formation, Caplin Cove Formation, Starboard Formation White Bay sequence: Bishie Cove Formation, Flat Bay Formation GRENVILLE BASEMENT		Oody Mountain Amphibolite GRENVILLE BASEMENT East Pond Metamorphic Suite ?

APPENDIX III

GEOCHEMICAL ANALYTICAL METHODS AND DATA

Samples were prepared by fragmenting rocks to 2 cm chips either with a prospector's pick and selecting unweathered material by hand, or by trimming and using a steel jaw crusher. The chips were subsequently pulverized to −100 mesh in a tungsten carbide swing mill.

Major element analyses were undertaken at two separate laboratories; samples 1340001 through 1340057 and 1340083 were analyzed by the Geochemical Laboratory of the Newfoundland Department of Mines and Energy whereas samples 1340058 through 1340082 were analyzed by the Earth Sciences Department of Memorial University of Newfoundland.

The following excerpt from a report by H. Wagenbauer (personal communication, 1982) describes the government laboratory techniques:

0.1000 g of rock powder (<150 μm in size) is dissolved at 90°C in a combination of 1 mL of concentrated HCl and 5 mL of concentrated HF acid in a 250 mL Nalgene Polycarbonate bottle. Samples which do not totally dissolve with this procedure are rerun in a teflon-lined steel bomb, and allowed to digest overnight at 100°C under pressure. The excess of HF acid is then complexed under 50 g of Boric acid per litre and the contents are once more digested at 90°C to dissolve any fluorides which may have precipitated in the first digestion. Samples are subsequently cooled and diluted to 100 mL.

All major oxides except FeO and P_2O_5 are determined by atomic absorption spectrophotometry. For SiO_2, Al_2O_3, total Fe, Na_2O, K_2O and MnO the final 100 mL solution from the digestions described above is used. For CaO, MgO and TiO_2, 25 mL of the final solution is pipetted to a 50 mL polyethylene volumetric flask, to which 15 mL of concentrated HCl and 5 mL of 5% lanthanum solution are added, and the solution made up to 50 mL and mixed.

Phosphorus is determined spectrophotometrically as the reduced phosphomolybdate complex (Riley, J.P., 1958).

Ferrous iron is determined on a separate sample by decomposing at room temperature in an HF solution containing a known amount of quinquevalent vanadium, which oxidizes the ferrous iron as it passes into solution. After standing until decomposition is complete the excess V^{5+} is reduced by a known quantity of ferrous ammonium sulphate, the excess ferrous iron being titrated with standard potassium dichromate using barium diphenylamine sulphomate indicator. Boric acid is used to complex the HF, and phosphoric acid the ferric iron (Wilson, 1955).

Loss on ignition at 1000°C is determined as an estimate of the water of crystallization, carbon dioxide, sulphur and organic matter present in rocks (Hillebrand et al., 1953).

Samples 1340058 through 1340082 were determined by G. Andrews using a Perkin Elmer 403 digitized atomic absorption spectrophotometer. Ferrous iron was determined by titration (Wilson, 1955), P_2O_5 by colorimetry (Maxwell, 1968), and 'loss on ignition' by weighing samples before and after heating in porcelain crucibles to 1050°C for two hours in a muffle furnace.

All of the trace element analyses were determined by Dave Press of Memorial University using a Phillips 1450 computerized X-Ray fluorescence spectrometer calibrated against international standards. The samples were run as pellets prepared by subjecting a mixture of 10 g of rock powder and 1.25 g of phenol formaldehyde to 50 MPa pressure in a 40 mm diameter die for 1 minute, then baking at 200°C.

Table AIII-1: *Major element data.*

STRATIGRAPHIC UNIT	SAMPLE NUMBER	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	TiO_2	MnO	P_2O_5	LOI	TOTAL
WHITE BAY GROUP	1340019	50.3	13.20	14.56	0.00	6.97	5.99	4.54	0.69	0.99	0.28	0.00	0.83	98.35
amphibolites	1340021	51.3	12.75	12.72	0.00	5.81	7.22	2.65	0.96	1.63	0.21	0.16	1.28	96.69
	1340022	50.4	13.55	12.84	0.00	6.84	5.38	4.97	0.32	1.89	0.15	0.19	1.72	98.25
	1340023	45.2	13.40	1.99	8.80	6.03	8.53	4.29	1.43	1.40	0.20	0.00	7.47	98.74
	1340046	48.8	12.80	12.09	4.06	4.50	7.05	3.04	1.03	3.15	0.27	0.49	1.70	98.98
	1340050	51.8	13.60	5.97	8.29	4.71	5.63	4.68	0.48	2.36	0.23	0.33	0.74	98.82
	1340051	47.0	13.65	9.47	6.07	5.85	6.61	3.45	0.46	3.02	0.23	0.32	1.93	98.06
	1340053	49.4	14.35	2.62	7.62	8.08	11.35	2.54	0.55	1.46	0.19	0.15	1.62	99.93
	1340057	49.2	13.40	2.97	8.89	6.86	11.51	2.12	0.46	1.41	0.21	0.16	0.91	98.10
RATTLING BROOK	1340030	38.4	11.40	7.45	6.17	5.60	15.66	1.09	0.22	1.70	0.24	0.00	10.31	98.24
GROUP amphibolites	1340054	48.4	15.25	5.61	5.84	5.30	8.48	3.58	0.97	2.38	0.20	0.27	1.82	98.10
	1340055	50.1	12.95	7.04	7.54	5.98	9.26	2.01	0.15	1.51	0.24	0.14	2.47	99.39
	1340056	46.4	15.55	5.10	7.19	6.90	7.93	3.02	1.08	2.15	0.25	0.22	2.52	98.31
Amphibolites in the	1340020	50.4	13.10	2.88	10.45	6.39	9.27	3.86	0.68	2.12	0.19	0.00	1.28	100.62
OLD HOUSE COVE	1340031	49.2	13.50	14.14	0.00	7.07	11.07	2.26	0.17	1.44	0.22	0.08	0.03	99.18
GROUP and EAST POND	1340033	51.6	14.10	2.01	9.75	7.88	8.61	3.22	1.37	1.51	0.21	0.00	0.89	101.15
METAMORPHIC SUITE	1340047	48.1	14.75	2.39	9.10	6.95	10.24	3.46	0.22	1.25	0.22	0.12	1.72	98.52
	1340048	51.8	14.00	2.49	8.99	6.32	7.01	4.59	0.31	1.62	0.14	0.15	1.12	98.54
	1340049	50.4	13.50	2.29	9.66	6.46	9.60	2.67	0.46	2.20	0.21	0.20	1.02	98.67
	1340052	51.1	13.55	2.28	9.74	6.12	8.94	2.68	0.48	2.30	0.20	0.11	1.01	98.51
BIRCHY COMPLEX														
metagabbro	1340024	50.2	14.70	1.98	4.66	10.55	13.27	1.83	0.12	0.36	0.14	0.00	2.16	99.97
metagabbro	1340025	50.5	14.30	2.11	4.06	10.08	14.11	1.82	0.00	0.37	0.12	0.00	1.88	99.35
metagabbro	1340026	49.0	14.50	1.20	5.07	10.21	13.10	1.78	0.04	0.34	0.12	0.00	1.93	97.29
metagabbro	1340027	43.7	13.60	3.92	10.04	8.40	12.42	2.11	0.56	2.10	0.24	0.00	1.37	98.46
metagabbro	1340028	47.0	18.00	1.28	4.27	9.06	13.10	2.15	0.03	0.25	0.10	0.00	2.52	97.76
mafic schist	1340035	49.0	12.50	3.36	9.85	7.72	8.59	3.12	0.05	1.85	0.19	0.00	2.49	98.72
mafic schist	1340036	48.0	14.30	3.74	9.39	7.70	10.20	2.34	0.28	1.42	0.20	0.00	2.54	100.11
MING'S BIGHT GROUP														
amphibolite	1340083	46.3	16.70	1.34	2.64	13.23	16.68	0.60	0.22	0.09	0.08	0.00	2.48	100.36
Prekinematic ultramafic rock	1340029	37.2	0.67	3.74	3.79	39.58	0.14	0.00	0.00	0.03	0.12	0.00	15.36	100.65
PACQUET HARBOUR GROUP														
massive lava	1340001	52.6	11.95	1.07	7.19	14.85	9.12	2.13	0.14	0.17	0.13	0.04	2.16	101.55
mafic dike	1340002	48.4	18.75	2.43	6.05	7.18	11.57	2.63	0.56	1.30	0.17	0.13	2.15	101.32
mafic dike	1340003	47.5	17.95	1.77	6.94	8.14	11.64	2.39	0.54	1.34	0.15	0.14	1.85	100.35
pillow lava	1340005	51.4	9.40	0.89	7.32	17.03	8.54	1.53	0.06	0.07	0.17	0.02	3.28	99.71
pillow lava	1340006	48.3	15.35	2.95	8.06	8.39	6.89	3.39	0.10	1.59	0.15	0.16	2.54	97.87
mafic tuff	1340007	51.9	14.65	2.17	5.82	9.56	8.70	3.06	0.35	0.63	0.16	0.08	2.80	99.88
felsic tuff	1340008	74.1	13.10	0.81	1.45	2.04	0.98	5.58	1.28	0.12	0.03	0.05	1.11	100.65
pillow lava	1340010	49.9	10.55	12.75	5.27	13.35	4.17	0.21	0.05	0.12	0.18	0.05	4.80	101.40
pillow lava	1340013	53.8	9.90	1.87	7.68	12.79	9.18	3.28	0.09	0.17	0.13	0.07	2.40	101.36
pillow lava	1340014	49.7	10.00	1.57	7.36	16.75	9.42	1.57	0.06	0.18	0.18	0.07	3.36	100.22
pillow lava	1340015	51.7	8.80	1.06	7.47	16.88	9.53	1.73	0.08	0.10	0.12	0.04	2.75	100.26
pillow lava	1340016	52.0	9.25	1.42	7.68	15.00	10.80	1.92	0.18	0.09	0.12	0.05	1.88	100.39
pillow lava	1340017	51.3	7.90	1.24	7.84	16.58	11.86	1.08	0.33	0.07	0.16	0.06	2.01	100.43
pillow lava	1340018	60.6	9.65	1.27	9.30	9.86	3.80	1.00	0.06	0.10	0.16	0.08	3.71	99.59
keratophyre	1340045	15.3	4.40	0.19	2.58	2.72	47.20	0.20	0.62	0.33	0.45	0.00	25.32	99.31
ADVOCATE COMPLEX														
pillow lava	1340037	45.4	15.60	6.25	3.88	4.75	13.36	3.28	0.49	1.67	0.15	0.00	3.94	98.77
POINT ROUSSE COMPLEX														
mafic dike	1340011	48.1	14.70	1.95	9.23	7.62	11.08	3.78	0.22	1.55	0.19	0.18	0.88	99.48
FLAT WATER POND GROUP														
felsic tuff	1340038	73.3	14.60	0.69	1.98	0.71	1.08	5.73	1.94	0.22	0.05	0.00	1.08	101.38
felsic tuff	1340039	71.7	14.70	0.39	2.48	0.86	1.33	6.46	1.60	0.29	0.05	0.00	1.45	101.31
felsic tuff	1340040	75.7	14.00	0.26	1.24	0.41	0.90	6.58	1.79	0.09	0.01	0.00	0.46	101.44
mafic tuff	1340041	50.4	17.70	3.65	5.73	5.87	7.43	4.56	0.84	1.57	0.13	0.00	2.51	100.39
felsic tuff	1340042	74.7	14.10	0.48	1.28	0.70	0.82	5.90	1.44	0.10	0.03	0.00	1.00	100.55
felsic tuff	1340043	64.6	16.60	1.26	3.39	1.78	2.51	6.37	1.81	0.69	0.07	0.00	1.90	100.98
felsic tuff	1340044	71.5	14.40	0.16	2.06	0.40	1.83	5.53	1.85	0.11	0.04	0.00	2.38	100.26
CAPE BRULE PORPHYRY														
granite porphyry	1340004	76.3	11.90	0.39	0.53	0.18	0.58	3.38	5.03	0.05	0.03	0.04	0.59	99.00
felsic tuff	1340009	77.4	12.25	2.06	1.16	0.98	0.45	1.55	3.77	0.20	0.05	0.04	1.85	101.76
felsic tuff	1340012	67.9	14.15	4.18	0.75	1.91	2.08	2.95	4.32	0.89	0.07	0.16	1.63	100.99

continued .....

Table AIII-1 *(cont.)*

STRATIGRAPHIC UNIT	SAMPLE NUMBER	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	TiO_2	MnO	P_2O_5	LOI	TOTAL
HORSE ISLANDS GROUP														
mafic dike	1340034	44.4	14.80	1.36	8.98	6.61	7.74	3.16	0.32	1.44	0.18	0.00	9.10	98.09
WILD COVE POND IGNEOUS SUITE														
granite	1340058	68.1	15.70	1.48	1.65	1.13	2.81	3.99	3.85	0.49	0.05	0.17	0.94	100.36
diorite	1340059	55.1	17.00	1.96	4.59	4.42	8.52	3.45	1.98	1.54	0.13	0.30	1.57	100.56
monzodiorite	1340060	54.3	17.50	0.35	7.30	4.09	6.35	3.83	2.98	1.37	0.15	0.42	2.06	100.70
granodiorite	1340061	71.1	14.40	0.00	1.31	0.60	1.44	3.40	4.64	0.26	0.05	0.10	1.03	98.33
granite	1340062	73.7	14.40	0.78	0.84	0.26	1.11	3.60	4.71	0.14	0.04	0.14	0.76	100.48
monzodiorite	1340063	69.6	15.20	0.35	0.73	0.80	1.91	3.62	4.96	0.48	0.05	0.14	0.72	98.56
diorite	1340064	64.2	17.40	1.12	2.50	1.53	3.24	3.87	3.63	0.64	0.07	0.28	1.07	99.55
granite	1340065	71.2	14.90	1.66	1.17	0.49	1.76	3.43	4.44	0.27	0.03	0.08	0.83	100.26
granite	1340066	71.0	16.20	0.62	0.66	0.29	2.05	4.57	3.89	0.18	0.02	0.03	0.53	100.04
granite	1340067	75.2	13.20	0.42	0.36	0.00	0.24	2.39	7.63	0.00	0.00	0.01	0.57	100.02
monzodiorite	1340068	56.3	20.80	0.08	2.99	1.51	3.98	5.96	3.25	0.76	0.08	0.78	1.12	97.61
granodiorite	1340069	67.3	17.50	2.68	1.18	0.63	2.92	4.96	3.14	0.31	0.02	0.10	0.89	101.63
granite	1340070	71.8	14.20	0.49	1.10	0.49	1.27	3.36	4.47	0.18	0.04	0.07	0.88	98.35
granite	1340071	71.7	14.40	0.65	0.97	0.43	1.35	3.37	4.47	0.19	0.05	0.06	0.86	98.50
granite	1340072	70.0	15.60	1.03	0.97	0.57	1.90	4.31	4.62	0.32	0.05	0.09	1.17	100.63
diorite	1340073	61.5	15.00	0.00	0.00	3.34	4.19	3.73	2.86	1.29	0.12	0.26	2.45	94.74
granite	1340074	71.3	14.20	0.43	0.95	0.45	1.46	3.58	5.00	0.23	0.04	0.05	0.68	98.37
granodiorite	1340075	73.2	13.90	0.25	1.20	0.49	1.43	3.59	4.42	0.24	0.05	0.08	0.65	99.50
granite	1340076	71.0	14.80	0.33	1.33	0.50	1.31	3.76	4.60	0.24	0.05	0.11	0.73	98.76
granodiorite	1340077	74.4	13.90	0.13	2.69	0.20	0.93	3.41	5.20	0.01	0.03	0.03	0.47	101.40
granodiorite	1340078	67.5	15.70	0.68	2.03	0.94	2.26	3.72	4.35	0.40	0.04	0.20	0.84	98.66
granodiorite	1340079	68.8	16.40	0.68	1.09	0.65	2.87	4.69	2.61	0.28	0.03	0.08	0.65	98.83
granite	1340080	72.8	13.70	0.27	0.79	0.18	0.91	3.53	5.28	0.11	0.03	0.05	0.37	98.02
granite	1340081	67.8	14.90	1.18	1.48	1.21	3.00	3.64	3.66	0.32	0.04	0.13	0.79	98.15
granite	1340082	70.3	14.20	0.67	2.10	0.64	1.59	3.22	4.71	0.37	0.05	0.17	0.71	98.73
PARTRIDGE POINT GRANITE														
muscovite granite	1340032	74.0	14.20	0.12	0.72	0.27	0.94	4.29	4.51	0.11	0.03	0.00	0.62	99.81

Table AIII-2: *Trace element data.*

STRATIGRAPHIC UNIT	SAMPLE NUMBER	Ba	Sr	Rb	Zr	Cr	V	Pb	Cu	Zn	Ni	Y	Nb	La	Ag	Ga	Ce
WHITE BAY GROUP	1340019	147	92	18	165	27	382	3	134	143	52	29	13	-	26	24	-
	1340021	311	242	30	169	58	328	2	84	116	58	33	15	-	27	26	-
	1340022	140	123	8	131	167	353	10	23	138	46	21	8	-	29	23	-
	1340023	279	360	46	115	66	278	14	133	105	44	24	13	-	29	25	-
	1340046	334	189	21	239	-	392	9	32	142	55	71	16	21	-	19	46
	1340050	94	91	12	245	5	401	4	123	113	57	62	21	23	-	19	54
	1340051	175	206	11	221	11	473	5	9	113	33	65	14	15	-	-	38
	1340053	243	388	19	172	97	297	7	71	89	58	43	11	18	-	11	53
	1340057	136	211	5	81	32	329	2	125	81	66	31	-	8	-	18	27
RATTLING BROOK GROUP	1340030	131	236	3	117	51	389	-	75	121	39	29	7	-	25	19	-
	1340054	89	237	12	94	350	258	6	53	89	78	31	10	12	-	13	45
	1340055	44	169	2	87	-	421	8	175	82	41	44	3	15	-	1	28
	1340056	526	218	27	133	110	375	9	38	100	79	41	7	15	-	15	45
OLD HOUSE COVE GROUP and EAST POND METAMORPHIC SUITE	1340020	56	218	12	126	71	411	13	189	122	56	23	11	-	26	23	-
	1340031	52	97	10	77	124	347	1	174	114	72	22	7	-	25	23	-
	1340033	326	198	49	102	54	288	3	87	103	38	26	9	-	27	25	-
	1340047	13	261	2	57	139	321	15	115	73	64	31	-	2	-	10	9
	1340048	9	201	3	87	131	391	16	127	86	59	36	3	12	-	13	19
	1340049	75	221	8	142	64	303	6	37	91	56	37	18	17	-	16	35
	1340052	114	31	9	130	42	329	8	31	90	48	36	16	15	-	15	39
BIRCHY COMPLEX	1340024	10	96	3	20	530	158	-	82	42	131	11	4	-	32	17	-
	1340025	15	97	-	23	569	170	-	63	38	107	23	4	-	32	17	-
	1340026	11	88	3	20	320	149	6	59	45	110	10	4	-	32	18	-
	1340027	45	169	9	107	152	404	1	70	133	67	27	7	-	24	22	-
	1340028	-	124	-	22	422	113	-	112	41	153	9	4	-	33	18	-
	1340035	20	79	3	101	88	415	-	57	92	43	25	8	-	27	22	-
	1340036	31	164	7	70	78	373	-	76	97	35	21	5	-	27	18	-
MING'S BIGHT GROUP	1340083	25	157	14	-	1162	90	7	2	-	243	5	-	-	-	12	5
Prekinematic ultramafic rock	1340029	-	-	2	6	2653	38	-	17	40	2054	3	4	-	43	6	-
PACQUET HARBOUR GROUP	1340001	37	5	3	21	1048	246	5	-	43	232	6	-	1	-	10	32
	1340002	128	281	15	85	293	227	10	49	68	78	33	-	1	-	16	42
	1340003	79	326	28	90	216	237	3	76	67	86	27	-	9	-	18	47
	1340005	16	68	-	11	1894	202	2	-	88	314	4	-	2	-	11	35
	1340006	12	226	-	90	227	373	5	38	68	52	35	3	9	-	11	45
	1340007	54	93	4	42	424	232	8	7	53	133	18	-	12	-	16	47
	1340008		no data														
	1340010	21	10	-	25	1158	265	6	-	48	212	5	-	2	-	11	21
	1340013	17	69	1	14	1543	194	58	-	71	342	8	-	-	-	11	30
	1340014	17	91	-	17	1816	199	12	-	74	346	5	-	5	-	13	32
	1340015	10	57	1	11	1885	189	16	-	32	367	6	-	5	-	6	29
	1340016	22	103	3	8	1571	190	7	-	62	317	4	-	6	-	8	33
	1340017	65	62	6	8	1551	165	6	-	77	407	3	-	-	-	6	35
	1340018	17	57	1	15	1040	186	6	-	92	171	5	-	8	-	6	29
	1340045	247	411	27	143	5	33	57	4018	1715	29	20	20	-	25	12	-
ADVOCATE COMPLEX	1340037	102	247	9	129	149	228	1	86	82	116	17	15	-	28	22	-
POINT ROUSSE COMPLEX	1340011	78	248	1	98	223	342	3	-	75	67	39	2	6	-	18	51
FLAT WATER POND GROUP	1340038	250	149	44	207	-	11	17	32	61	3	27	13	-	46	24	-
	1340039	245	109	32	218	6	24	19	40	58	1	26	12	-	45	23	-
	1340040	105	107	25	152	4	3	1	22	24	-	23	13	-	49	21	-
	1340041	136	295	9	167	110	211	-	57	71	40	23	7	-	32	20	-
	1340042	125	127	32	178	-	-	14	21	52	-	26	12	-	49	23	-
	1340043	209	188	26	206	12	75	-	38	53	7	22	10	-	40	23	-
	1340044	211	65	30	204	2	6	6	31	24	3	27	12	-	45	27	-
CAPE BRULE PORPHYRY	1340004																
	1340009		no data														
	1340012																

continued.....

Table AIII-2: *(cont.)*

STRATIGRAPHIC UNIT	SAMPLE NUMBER	Ba	Sr	Rb	Zr	Cr	V	Pb	Cu	Zn	Ni	Y	Nb	La	Ag	Ga	Ce
HORSE ISLANDS GROUP dike	1340034	109	374	15	208	53	305	7	53	96	51	22	13	-	30	24	-
WILD COVE POND IGNEOUS SUITE	1340058	909	370	90	185	-	52	13	-	34	8	26	18	43	-	14	72
	1340059	392	513	55	139	66	180	7	19	47	44	31	28	23	-	14	39
	1340060	664	563	91	240	-	175	10	18	71	31	35	28	23	-	16	46
	1340061	847	189	156	189	-	28	19	5	32	12	35	24	34	-	15	62
	1340062	547	108	215	95	-	10	27	-	30	14	20	40	23	-	18	47
	1340063	802	213	145	196	-	33	19	-	38	10	35	24	52	-	17	93
	1340064	1517	454	108	259	-	72	12	9	58	8	17	11	52	-	15	101
	1340065	1602	379	107	149	4	16	18	-	27	5	12	8	9	-	12	56
	1340066	1614	741	74	106	-	10	21	-	21	-	4	3	3	-	14	47
	1340067	699	254	100	63	-	4	30	-	-	1	2	5	-	-	8	33
	1340068	2937	2002	47	488	-	50	-	16	107	7	16	9	53	-	18	144
	1340069	1753	1070	57	122	-	24	15	-	23	3	4	-	18	-	15	60
	1340070	768	210	146	127	-	21	28	-	18	30	28	19	17	-	19	228
	1340071	1038	230	169	126	-	16	25	-	16	38	34	18	25	-	20	210
	1340072	613	193	129	140	-	29	17	-	23	7	22	23	32	-	14	61
	1340073	253	206	164	166	13	155	6	12	58	38	52	50	18	-	15	35
	1340074	784	388	93	84	-	21	27	3	11	10	11	9	26	-	17	211
	1340075	552	300	106	93	-	19	27	-	14	13	10	17	5	-	17	225
	1340076	498	163	168	98	-	3	35	-	-	29	36	15	8	-	18	248
	1340077	1040	361	119	163	-	18	26	-	23	13	14	16	19	-	16	206
	1340078	1704	655	102	181	-	44	20	-	30	16	14	7	35	-	17	219
	1340079	1239	853	74	95	-	26	14	-	23	7	5	-	5	-	20	198
	1340080	1359	675	70	89	-	49	14	2	21	17	6	3	7	-	19	180
	1340081	842	199	142	93	-	10	35	-	-	23	27	16	20	-	18	217
	1340082	730	186	129	210	-	29	25	-	23	20	26	25	33	-	18	211
PARTRIDGE POINT GRANITE	1340032	485	108	141	78	1	5	15	24	45	1	31	16	-	47	25	-

APPENDIX IV

ANALYTICAL TECHNIQUES AND DATA FOR ISOTOPIC AGE DATES UNDERTAKEN DURING THE PRESENT STUDY

(taken from D. Dallmeyer, personal communication, 1981)

The decay constants and isotopic abundance ratios used in all age calculations are those recommended for adoption by Steiger and Jager (1977). All uncertainties in calculated ages are quoted at a two sigma level.

U-Pb Zircon Analyses

Zircons were extracted from rocks crushed to −80 mesh using a super-panner, magnetic separator and heavy liquids. Final purification was achieved by hand-picking beneath a microscope. Each bulk zircon concentrate was split into two magnetic fractions. These were individually cleaned in HNO_3 and H_2O. (All reagents were first purified by sub-boiling-point distillation in quartz and/or teflon containers.) The procedures used for zircon dissolution and U-Pb separations closely follow those described in detail by Krogh and Davis (1974). Uranium concentrations were separately determined for each zircon fraction by isotope dilution analysis using triple Re filament, thermal ionization mass spectrometry. Lead isotopic analyses were performed using a silica gel technique. Both unspiked and isotopically spiked (with ^{208}Pb aliquots of zircon solutions were analyzed. Mass fractionation effects were monitored through replicate analyses of the NBS-SRM Pb isotopic standard, for which the following ratios were measured at 1280°-1340°C Re filament temperatures during analyses of rocks of the Baie Verte Peninsula: $^{208}Pb/^{206}Pb = 2.1676$; $^{207}Pb/^{206}Pb = 0.91443$; $^{204}Pb/^{206}Pb = 0.05906$. Isotopic ratios used for common lead corrections were: $^{208}Pb/^{204}Pb = 36.00$; $^{207}Pb/^{204}Pb = 15.7$; $^{206}Pb/^{204}Pb = 18.5$.

Rb-Sr Whole-Rock Analyses

Crushed samples were prepared for isotopic analysis by dissolution with hydrofluoric and perchloric acid followed by ion-exchange separation of Rb and Sr. Whole-rock samples were analyzed for Rb and Sr concentrations by stable isotope dilution using the NBS SRM-981 enriched ^{84}Sr spike. During analysis of the Baie Verte Peninsula samples, the $^{87}Sr/^{86}Sr$ ratio measured on the Eimer and Amend Sr-carbonate standard was 0.70797 ± 0.00011. The analytical uncertainty estimated for the Rb and Sr concentrations is approximately $\pm$ 1%. The slope and intercept of the isochron were determined according to the methods outlined by York (1969). A mean square of weighted deviates (MSWD) was also calculated because this statistical parameter indicates how closely the analytical data approximates a linear (isochronous) distribution (Brooks et al., 1972). In the ideal case, the MSWD approaches a value of one or less if the analytical data fit an isochron within experimental error. However, in practice, an isochron interpretation for whole-rock Rb-Sr analytical data is generally accepted if an MSWD is less than approximately 5.0.

$^{40}Ar/^{39}Ar$ Analyses

The $^{40}Ar/^{39}Ar$ incremental-release dating technique has been described in detail elsewhere (Dalrymple and Lanphere, 1971; Dallmeyer, 1979). Pure mineral concentrates were prepared from crushed and sieved rock powders using heavy liquid and magnetic separation procedures. Mineral concentrates were individually wrapped in aluminum packets and encapsulated in quartz vials. These together with mineral age standards were irradiated for 40 hours at 1000 kW in the central thimble position of the U.S. Geological Survey TRIGA reactor in Denver where they received a total dose of approximately 4×10^{18} nvt. Following irradiation, the quartz vials were returned to the laboratory where they were opened and loaded in an ultra-high vacuum argon extraction system connected to a mass spectrometer. Each sample was incrementally heated until fusion by ratio frequency induction. Each heating increment was maintained for one hour. Temperatures were monitored and controlled with a noncontact, infrared thermometer. Gases liberated during each temperature increment were collected and individually analyzed with a Neir-type 6-inch radius, 60° sector mass spectrometer operated in a static mode. A total of eight sweeps across the 40-36 mass range was carried out during analysis of each increment. During each sweep, a ten-second integrated signal was measured for each mass peak and associated base lines with a digitizing voltmeter. Uncertainties in isotopic ratios were directly calculated through linear and/or exponential regression of zero-time drift extrapolations following the statistical methods proposed by Higbie (1978). The estimated uncertainty in the apparent age of each temperature increment is based on possible errors in: (1) the measured isotopic ratios; (2) various irradiation parameters; (3) the decay constant; and (4) monitor ages. The combined effects of these errors on the uncertainty in age of each temperature fraction was estimated through calculations proposed by Dalrymple and Lanphere (1971) using a modified version of a U.S. Geological Survey program.

Total-gas ages have been computed by appropriate weighting of the age and uncertainty of each temperature fraction. Criteria for definition of a 'plateau age' are those suggested by Fleck et al. (1977). A 'plateau' is therefore defined as the part of an $^{40}Ar/^{39}Ar$ age spectrum in which contiguous gas fractions together constituting more than 50% of all the gas liberated from a sample have apparent ages which are mutually similar at a two sigma level of uncertainty. Gas fractions containing less than 5% of the total ^{39}Ar liberated from a sample have been ignored in calculation of plateau ages.

Table AIV-1: *Isotopic data for zircon separates from the Burlington Granodiorite.*

Sample	Concentrations (ppm)		Observed Isotopic ratios of Pb			Ages (Ma)		
	U	Pb	208/206	207/206	204/206	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
HGC-8-77^{L}	836	60.1	0.2054	0.0662	0.0007	420	425	452
HGC-8-77	1020	82.6	0.4171	0.0625	0.0005	400	404	426
HGC-48-77^{LM}	494	35.2	0.3385	0.0709	0.0010	380	390	447
HGC-48-77^{MM}	646	40.3	0.2136	0.0639	0.0006	364	371	424

L = leached in cold HF

LM = least magnetic fraction

MM = most magnetic fraction. Common lead correction: 208/204 = 36.00; 207/204 = 15.7; 206/204 = 18.5 .

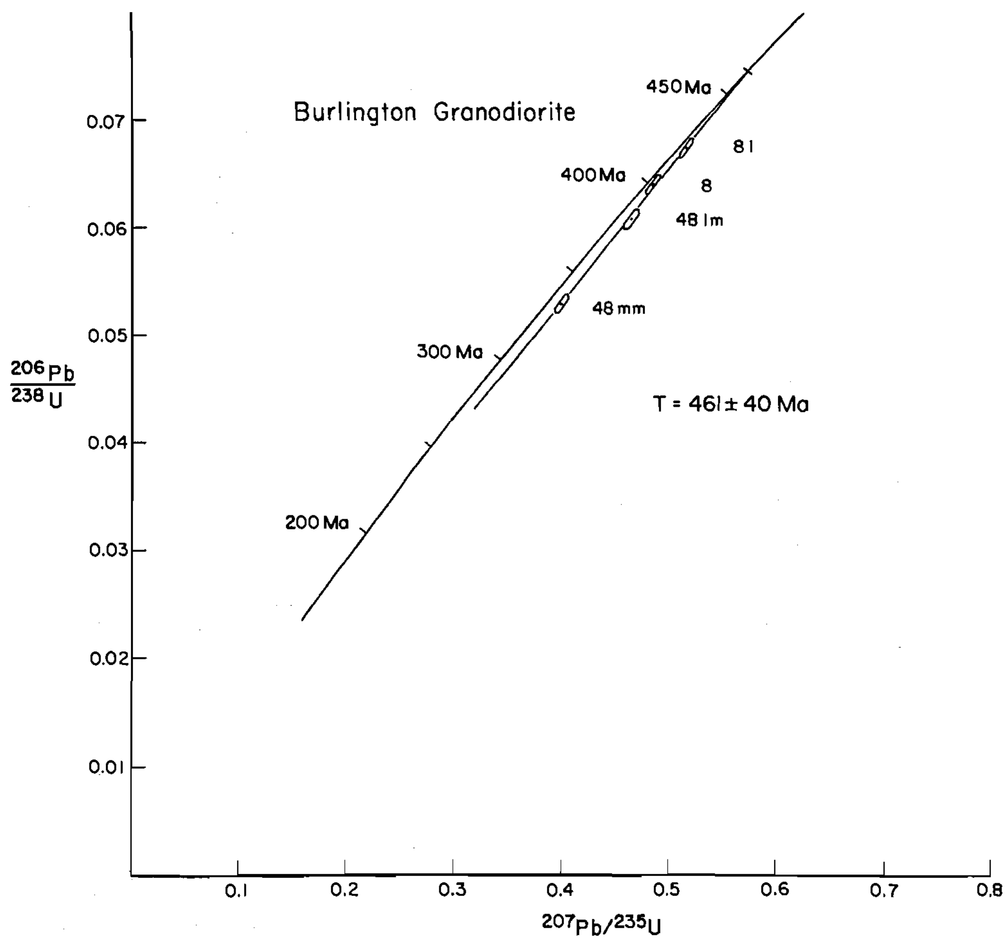

Figure AIV-1: *Zircon concordia plot for Burlington Granodiorite samples.*

Table AIV-2: *Rb-Sr analytical data for whole rock specimens of the Burlington Granodiorite.*

Sample	^{87}Rb (ppm)	^{86}Sr (ppm)	^{87}Rb/^{86}Sr (atomic)	^{87}Sr/^{86}Sr (atomic)
HGC-8-77	13.47	62.25	0.214	0.7063
HGC-47-77	20.37	60.90	0.331	0.7073
HGC-48-77	24.63	34.82	0.669	0.7095
HGC-51-77	15.50	66.04	0.232	0.7062
HGC-52-77	27.92	61.33	0.450	0.7078
HGC-55-77	19.53	70.22	0.275	0.7066

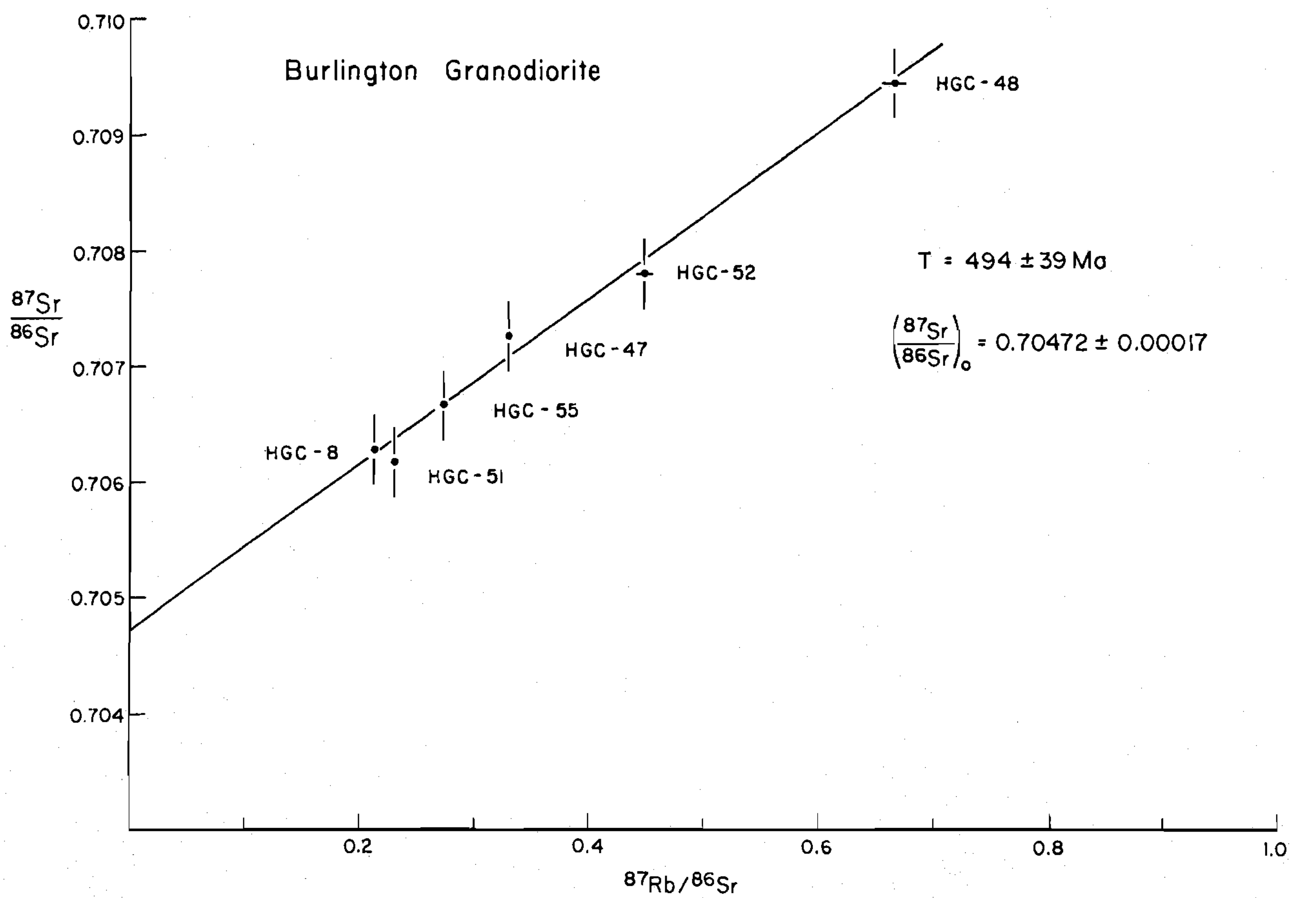

Figure AIV-2: *Rb/Sr whole rock isochron for the Burlington Granodiorite.*

Table AIV-3: $^{40}Ar/^{39}Ar$ *analytical data for incremental heating experiments on hornblende from the Burlington Granodiorite.*

Release Temp. (°C)	$(^{40}Ar/^{39}Ar)^m$	$(^{36}Ar/^{39}Ar)^m$	$(^{37}Ar/^{39}Ar)^c$	^{39}Ar % of Total	Radiogenic %	% $^{36}Ar_{Ca}$	Age (Ma)$^+$
Sample HGC-7-77, J = 0.008632:							
950	125.05	0.31186	8.781	1.77	24.49	0.85	461 ± 20
1000	38.57	0.03640	10.737	6.03	74.33	8.02	401 ± 10
1050	32.00	0.01039	9.336	54.81	92.74	24.45	414 ± 5
1100	31.82	0.00914	9.276	20.88	93.83	22.59	416 ± 5
Fusion	31.79	0.00932	10.571	16.52	93.99	30.02	416 ± 5
TOTAL				100.00			414 ± 10
Plateau (without 950°C)				98.23			418 ± 5
Sample HGC-12-77, J = 0.008632:							
550	97.16	0.26813	1.764	4.51	18.59	0.18	262 ± 20
700	57.23	0.09276	1.585	3.41	52.32	0.46	415 ± 15
800	53.40	0.08553	0.930	4.26	52.80	0.30	393 ± 10
900	51.64	0.08027	3.579	3.86	54.61	1.21	394 ± 10
1000	36.27	0.02457	12.257	13.27	82.68	13.57	418 ± 5
1050	31.66	0.00981	10.431	23.33	93.47	28.93	413 ± 5
1100	31.83	0.01008	9.416	24.39	93.00	25.41	413 ± 5
1150	32.79	0.01348	11.461	16.47	90.64	23.12	415 ± 5
Fusion	37.11	0.02720	12.777	6.50	81.09	12.78	420 ± 10
TOTAL				100.00			406 ± 10
Plateau (without 550°C)				95.49			414 ± 5
Sample HGC-13-77, J = 0.008632:							
600	87.16	0.19632	2.789	2.16	33.69	0.39	408 ± 15
850	62.02	0.11377	9.198	2.43	46.97	2.20	407 ± 10
1000	39.22	0.03289	12.238	10.79	77.71	10.12	424 ± 10
1050	31.87	0.01031	9.815	42.30	92.90	25.89	413 ± 5
1100	32.58	0.01306	9.589	25.82	90.50	19.97	411 ± 5
1150	35.40	0.02287	11.940	11.90	83.61	14.20	413 ± 5
Fusion	42.69	0.05215	12.272	4.60	66.20	6.40	397 ± 10
TOTAL				100.00			413 ± 5
Sample HGC-14-77, J = 0.008632:							
700	199.05	0.53094	3.813	0.98	21.33	0.20	564 ± 20
800	63.79	0.12093	2.479	1.14	44.28	0.56	394 ± 15
900	54.78	0.10318	8.341	1.20	45.56	2.20	354 ± 15
1000	33.33	0.01768	11.842	31.78	85.22	18.21	411 ± 5
1050	33.47	0.01380	9.446	6.10	90.07	18.61	420 ± 5
1100	31.14	0.00736	8.761	37.56	95.26	32.37	413 ± 5
1150	33.24	0.01298	9.926	14.47	90.85	20.81	420 ± 5
Fusion	36.17	0.02386	14.568	6.77	86.36	16.61	423 ± 10
TOTAL				100.00			416 ± 10
Plateau (without 700, 800 and 900°C)				96.68			414 ± 5
Sample HGC-31-77, J = 0.008632:							
600	231.96	0.21726	0.962	0.91	72.35	0.12	1616 ± 40
875	65.44	0.05839	1.076	3.65	73.75	0.50	629 ± 15
925	51.92	0.07552	3.366	2.58	57.53	1.21	415 ± 15
1025	36.36	0.01024	8.984	39.42	93.65	23.87	467 ± 5
1075	35.55	0.00787	8.240	25.97	94.64	28.47	465 ± 5
1125	38.84	0.02232	9.782	9.89	85.03	11.92	458 ± 5
Fusion	41.25	0.02799	9.532	17.59	81.80	9.26	464 ± 5
TOTAL				100.00			472 ± 15
Plateau (without 600, 875 and 925°C)				92.87			464 ± 5

continued .....

Table AIV-3: *(continued)*

Release Temp. (°C)	$(^{40}Ar/^{39}Ar)^m$	$(^{36}Ar/^{39}Ar)^m$	$(^{37}Ar/^{39}Ar)^c$	^{39}Ar % of Total	Radiogenic %	% $^{36}Ar_{Ca}$	Age (Ma)$^+$
Sample HGC-42-77 (100/150), J = 0.008026:							
600	53.92	0.09112	1.922	5.03	50.34	0.57	365 ± 15
875	50.00	0.07139	16.095	7.27	60.39	6.13	395 ± 5
925	43.94	0.04703	13.494	9.84	70.82	7.81	405 ± 5
975	36.15	0.01962	12.754	20.20	86.79	17.69	408 ± 5
1025	33.94	0.01111	10.326	22.80	92.76	25.29	409 ± 5
1075	34.75	0.01371	10.317	22.23	90.71	20.46	409 ± 5
Fusion	43.28	0.04528	13.533	12.62	71.58	8.13	404 ± 5
TOTAL				100.00			404 ± 10
Plateau (without 600°C)				94.97			406 ± 5
Sample DA-BP6-79, J = 0.008841:							
550	29.08	0.01064	0.572	3.49	89.33	1.46	373 ± 12
600	33.28	0.02581	1.193	3.23	77.36	1.26	370 ± 12
650	28.76	0.01089	3.035	3.01	89.64	7.58	371 ± 11
700	35.60	0.02282	1.318	9.58	81.34	1.57	411 ± 10
750	30.78	0.01220	9.783	6.74	90.81	21.81	404 ± 10
800	30.55	0.00730	10.023	24.65	95.56	37.35	417 ± 8
850	30.46	0.00652	9.759	27.00	96.23	40.73	418 ± 8
900	33.82	0.01673	11.328	21.20	88.06	18.42	421 ± 9
Fusion	60.37	0.09389	10.790	1.10	55.47	3.13	471 ± 42
TOTAL	32.03	0.01268	8.562	100.00	91.04	26.08	413 ± 10
Total without 550, 600, 650°C and fusion				89.17			417 ± 8

m - measured
c - corrected for ^{37}Ar decay
+ - calculated using correction factors of Dalrymple and Lanphere (1971); two sigma error estimates

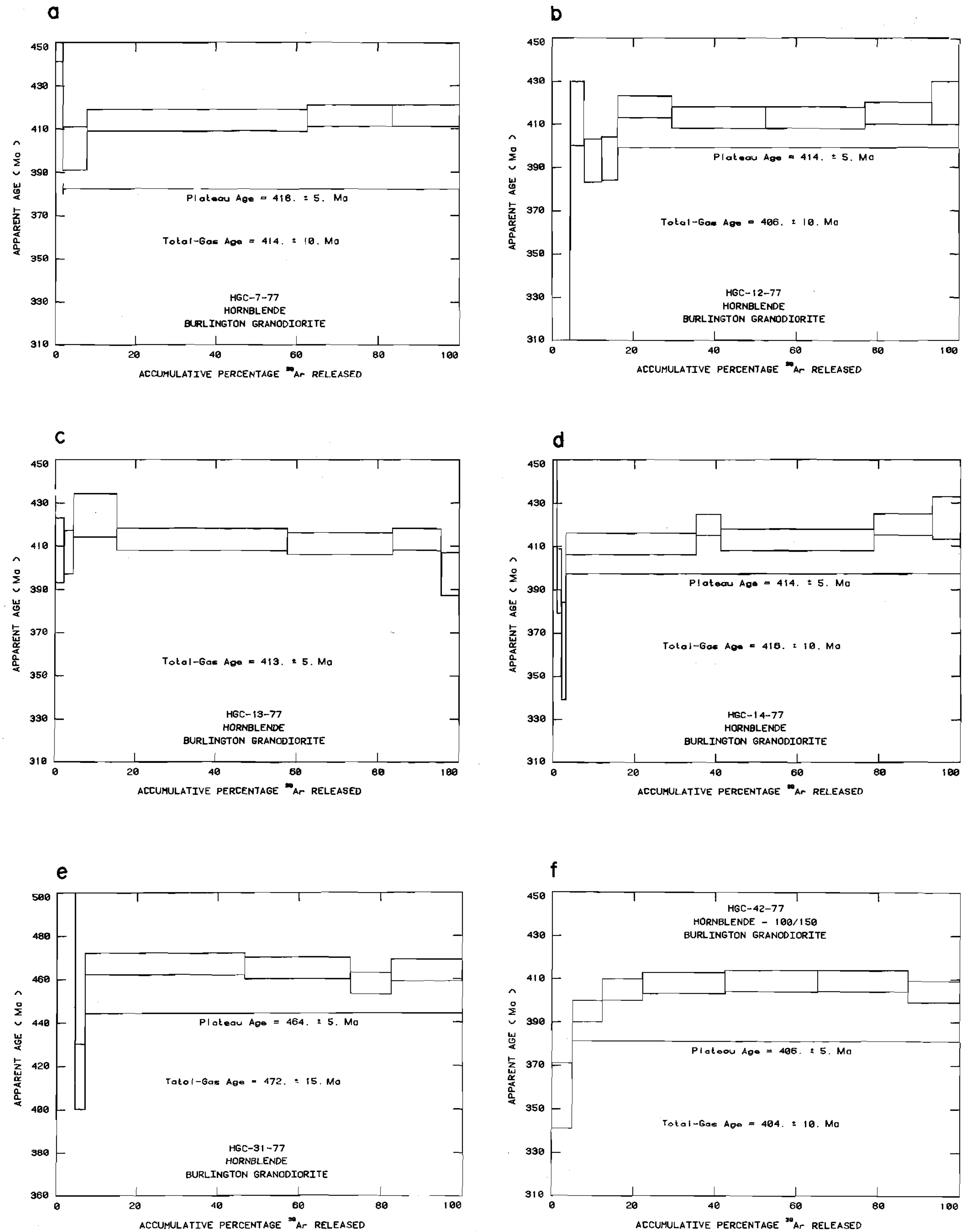

Figure AIV-3a-g: *$^{40}Ar/^{39}Ar$ age spectra for hornblende samples from the Burlington Granodiorite.*

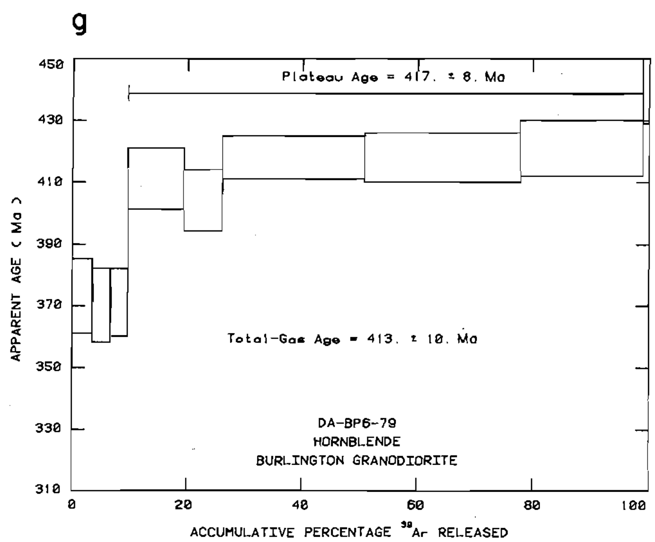

Figure AIV-3g

Table AIV-4: *$^{40}Ar/^{39}Ar$ analytical data for incremental heating experiments on biotite from the Burlington Granodiorite and the Dunamagon Granite.*

Release Temp. (°C)	$(^{40}Ar/^{39}Ar)^m$	$(^{36}Ar/^{39}Ar)^m$	^{39}Ar % of Total	Radiogenic %	Age (Ma)$^+$
Sample HGC-7-77, J = 0.008347:					
475	25.39	0.03951	1.15	54.03	195 ± 20
525	34.84	0.02655	3.23	77.46	367 ± 15
600	34.83	0.01309	3.35	88.88	415 ± 10
800	30.69	0.00109	30.53	98.93	407 ± 5
875	31.98	0.00215	7.10	97.99	419 ± 5
950	32.51	0.00283	7.95	97.41	423 ± 5
1025	32.18	0.00276	19.13	97.45	419 ± 5
Fusion	31.01	0.00138	27.55	98.67	410 ± 5
TOTAL			100.00		409 ± 10
Plateau (without 475 and 525°C)			95.62		414 ± 10
Sample HGC-12-77, J = 0.008448:					
475	17.18	0.02110	2.18	63.69	160 ± 20
525	31.19	0.00999	4.09	90.52	386 ± 15
600	31.66	0.00387	8.40	96.37	414 ± 5
700	31.35	0.00315	9.96	97.02	413 ± 5
800	31.35	0.00402	10.99	96.20	409 ± 5
875	32.20	0.00359	6.57	96.69	421 ± 5
925	32.08	0.00257	9.48	97.61	423 ± 5
950	31.14	0.00139	13.87	98.67	416 ± 5
1000	30.42	0.00078	26.45	99.22	410 ± 5
Fusion	30.63	0.00057	8.01	99.43	413 ± 5
TOTAL			100.00		407 ± 10
Plateau (without 475 and 525°C)			93.73		412 ± 10
Sample HGC-42-77, J = 0.008360:					
475	7.13	0.01158	6.71	51.90	55 ± 25
525	29.97	0.00377	10.43	96.26	390 ± 10
700	32.79	0.00319	12.47	97.10	426 ± 5
800	36.38	0.01604	1.66	86.95	423 ± 15
925	36.34	0.00817	6.74	93.34	450 ± 10
1025	36.13	0.00155	24.78	98.72	471 ± 5
Fusion	31.87	0.00109	37.20	98.97	422 ± 5
TOTAL			100.00		409 ± 10

continued.....

Table AIV-4: *(continued)*

Release Temp. (°C)	$(^{40}Ar/^{39}Ar)^m$	$(^{36}Ar/^{39}Ar)^m$	^{39}Ar % of Total	Radiogenic %	Age (Ma)$^+$
Sample DA-BP4-79, J = 0.008130:					
475	30.26	0.04134	0.53	59.61	247 ± 12
500	26.23	0.00437	7.21	95.06	333 ± 8
550	25.96	0.00045	22.59	99.46	344 ± 6
600	26.23	0.00077	7.62	99.11	346 ± 6
650	26.68	0.00199	4.56	97.78	347 ± 7
675	27.20	0.00366	3.08	96.01	347 ± 8
700	27.19	0.00312	2.74	96.59	349 ± 8
725	26.43	0.00079	5.61	99.09	348 ± 7
750	26.24	0.00063	11.43	99.27	347 ± 6
800	26.04	0.00039	16.54	99.53	345 ± 7
850	26.14	0.00070	11.51	99.19	345 ± 6
Fusion	28.00	0.00668	6.59	92.93	346 ± 6
TOTAL	26.35	0.00169	100.00	98.15	344 ± 6
Total without 475°C			99.47		345 ± 5
Sample DA-BP5-79, J = 0.008320:					
475	29.82	0.03789	0.53	62.43	260 ± 7
500	25.32	0.00199	12.20	97.66	338 ± 6
550	25.34	0.00052	18.22	99.37	343 ± 6
600	25.47	0.00073	8.21	99.13	344 ± 6
650	25.75	0.00135	6.58	98.43	345 ± 7
700	25.89	0.00170	4.75	98.04	346 ± 8
725	25.52	0.00058	9.02	99.30	345 ± 6
750	25.27	0.00024	15.48	99.70	343 ± 6
775	25.22	0.00060	11.36	99.27	341 ± 6
800	25.43	0.00080	5.88	99.04	343 ± 6
Fusion	26.94	0.00616	7.77	93.23	342 ± 6
TOTAL	25.55	0.00145	100.00	98.36	342 ± 6
Total without 475°C			99.47		343 ± 5
Sample HGC-44-77, J = 0.008379:					
475	20.66	0.03819	1.07	45.35	136 ± 20
525	20.86	0.00510	4.74	92.74	271 ± 10
700	24.97	0.00075	34.61	99.09	340 ± 5
800	25.73	0.00157	13.58	98.17	346 ± 5
925	25.34	0.00058	30.08	99.30	345 ± 5
1025	25.36	0.00101	13.47	98.80	344 ± 5
Fusion	28.93	0.01233	2.44	87.38	346 ± 5
TOTAL			100.00		337 ± 10
Plateau (without 475 and 525°C)			94.19		344 ± 5

m - measured
\+ - calculated using correction factors of Dalrymple and Lanphere (1971); 2 sigma error estimates (rounded to 5 Ma); $^{37}Ar/^{39}Ar$ corrected ratio less than 0.020 in all analyses.

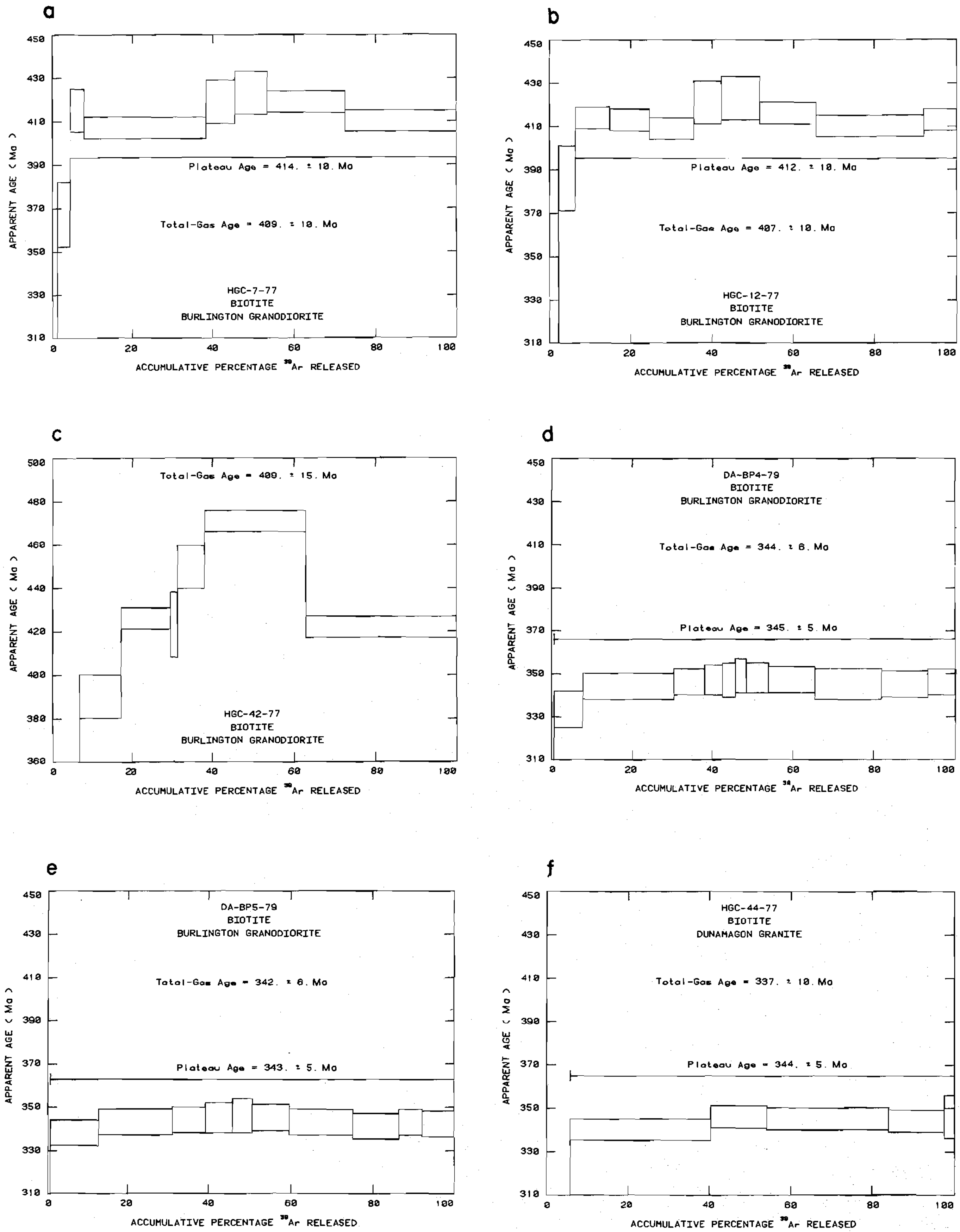

Figure AIV-4a-f: *$^{40}Ar/^{39}Ar$ age spectra for biotite samples from the Burlington Granodiorite.*

Table AIV-5: *$^{40}Ar/^{39}Ar$ analytical data for incremental heating experiments on hornblende from the Pacquet Harbour and Ming's Bight Groups.*

Release Temp. (°C)	$(^{40}Ar/^{39}Ar)^{m}$	$(^{36}Ar/^{39}Ar)^{m}$	$(^{37}Ar/^{39}Ar)^{c}$	^{39}Ar % of Total	Radiogenic %	% $^{36}Ar_{Ca}$	Age (Ma)+
			Pacquet Harbour Group				
Sample HGC-27-77, J = 0.008070:							
600	90.47	0.22638	4.320	1.97	26.43	0.52	319 ± 20
800	212.21	0.63318	6.095	2.64	12.06	0.26	340 ± 10
900	62.47	0.12550	10.991	3.21	42.04	2.38	349 ± 10
950	57.02	0.11016	21.935	3.59	45.99	5.42	351 ± 10
1000	34.15	0.03366	18.510	14.53	75.22	14.96	344 ± 5
1050	32.72	0.02715	19.062	13.09	80.14	19.10	350 ± 5
1100	46.25	0.07204	19.842	5.19	57.14	7.49	354 ± 5
Fusion	29.47	0.01594	19.968	45.78	89.44	34.07	352 ± 5
TOTAL				100.00			349 ± 5
Plateau (without 600°C)				98.03			350 ± 5
Sample HGC-36-77, J = 0.008632:							
600	1237.13	4.01040	22.052	0.14	4.35	0.15	696 ± 40
875	96.58	0.24137	19.381	2.03	27.75	2.18	380 ± 15
950	27.50	0.01493	19.487	56.12	89.63	35.51	352 ± 5
Fusion	35.35	0.03868	15.327	41.71	72.67	10.78	358 ± 5
TOTAL				100.00			358 ± 5
Plateau (without 600°C)				99.86			356 ± 5
Sample HGC-28-77, J = 0.008429:							
550	83.27	0.20472	5.417	4.55	27.87	0.72	323 ± 15
700	33.18	0.03078	13.247	12.94	75.77	11.70	349 ± 5
800	67.58	0.15001	15.398	2.72	36.24	2.79	342 ± 10
900	79.01	0.18107	14.930	2.27	33.79	2.24	369 ± 10
1000	31.06	0.02653	17.675	15.90	79.32	18.12	344 ± 5
1050	31.40	0.02440	17.867	19.65	81.59	19.92	356 ± 5
1075	51.80	0.09247	17.126	2.61	49.90	5.04	359 ± 10
1100	61.69	0.12870	17.696	2.43	40.65	3.74	349 ± 10
1150	31.98	0.02542	18.405	28.08	81.12	19.70	361 ± 5
Fusion	35.88	0.03881	18.779	8.87	72.23	13.16	360 ± 5
TOTAL				100.00			353 ± 5
Plateau (without 550°C)				95.45			355 ± 5
			Ming's Bight Group				
Sample HGC-21-77, J = 0.008632:							
600	59.36	0.11057	1.205	2.03	45.11	0.30	375 ± 15
600	36.33	0.04034	9.057	6.63	69.17	6.11	356 ± 10
900	48.80	0.08614	8.973	2.88	49.30	2.83	342 ± 15
Fusion	28.42	0.01464	11.335	88.45	87.96	21.06	355 ± 5
TOTAL				100.00			356 ± 5
Sample HGC-22-77, J = 0.007799:							
600	262.73	0.77024	15.738	1.44	13.85	0.56	455 ± 20
875	141.89	0.38238	3.121	3.74	20.54	0.22	370 ± 15
925	33.25	0.02336	17.281	38.64	83.40	20.12	357 ± 5
1025	39.06	0.04388	17.259	4.33	70.34	10.70	354 ± 5
1075	33.41	0.02528	17.976	11.11	81.95	19.34	353 ± 5
1125	32.47	0.02198	18.294	20.77	84.51	22.64	354 ± 5
Fusion	49.08	0.07558	18.061	19.91	57.44	6.50	362 ± 5
TOTAL				100.00			359 ± 5
Plateau (without 600°C)				98.56			357 ± 5

m - measured
c - corrected for ^{37}Ar decay
+ - calculated using correction factors of Dalrymple and Lanphere (1971); 2 sigma error estimates (rounded to 5 Ma).

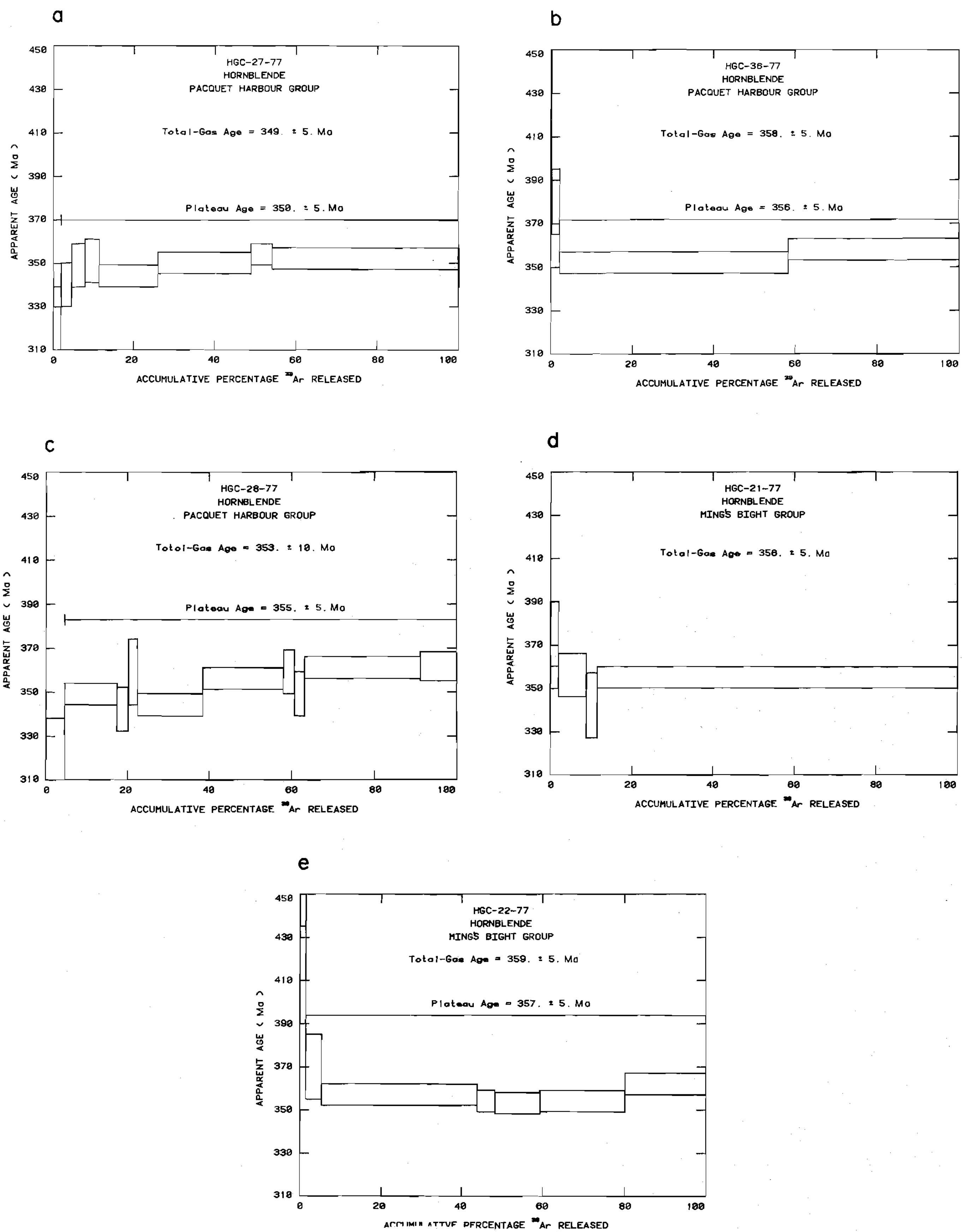

Figure AIV-5a-e: *$^{40}Ar/^{39}Ar$ age spectra for hornblende samples from the Pacquet Harbour and Ming's Bight Groups.*

Table AIV-6: $^{40}Ar/^{39}Ar$ *analytical data for incremental heating experiments on biotite from the Pacquet Harbour and Ming's Bight Groups.*

Release Temp. (°C)	$(^{40}Ar/^{39}Ar)^m$	$(^{36}Ar/^{39}Ar)^m$	^{39}Ar % of Total	Radiogenic %	Age (Ma)$^+$
Sample DA-BP2-79, J = 0.008315:					
475	25.57	0.03111	0.58	64.02	230 ± 15
500	25.29	0.00176	8.49	97.92	338 ± 6
550	25.16	0.00063	16.06	99.24	340 ± 6
650	25.27	0.00066	10.71	99.20	342 ± 7
775	25.20	0.00032	42.11	99.60	342 ± 6
Fusion	25.32	0.00128	22.07	98.48	340 ± 6
TOTAL	25.24	0.00092	100.00	98.91	340 ± 6
Sample DA-BP3-79, J = 0.008492:					
475	32.17	0.06412	0.30	41.08	192 ± 18
500	23.32	0.00354	7.35	95.48	312 ± 9
550	25.63	0.00184	8.02	97.86	348 ± 6
675	25.46	0.00105	17.25	98.76	349 ± 6
775	25.16	0.00050	31.99	99.39	347 ± 6
Fusion	25.03	0.00087	35.09	98.95	344 ± 6
TOTAL	25.09	0.00125	100.00	98.54	344 ± 8
Total without 475 and 500°C			92.35		347 ± 5
Sample DA-BP1-79, J = 0.008135:					
475	15.76	0.01273	1.78	76.10	168 ± 4
500	24.97	0.00174	11.14	97.92	327 ± 7
550	25.35	0.00052	21.65	99.37	336 ± 6
650	25.81	0.00095	9.31	98.89	340 ± 7
775	25.48	0.00044	36.10	99.47	338 ± 6
Fusion	25.82	0.00169	20.03	98.04	338 ± 6
TOTAL	25.32	0.00112	100.00	98.52	334 ± 6
Total without 475°C			98.22		337 ± 4

m - measured
\+ - calculated using correction factors of Dalrymple and Lanphere (1971); two sigma error estimates; $^{37}Ar/^{39}Ar$ corrected ratio less than 0.020 in all analyses.

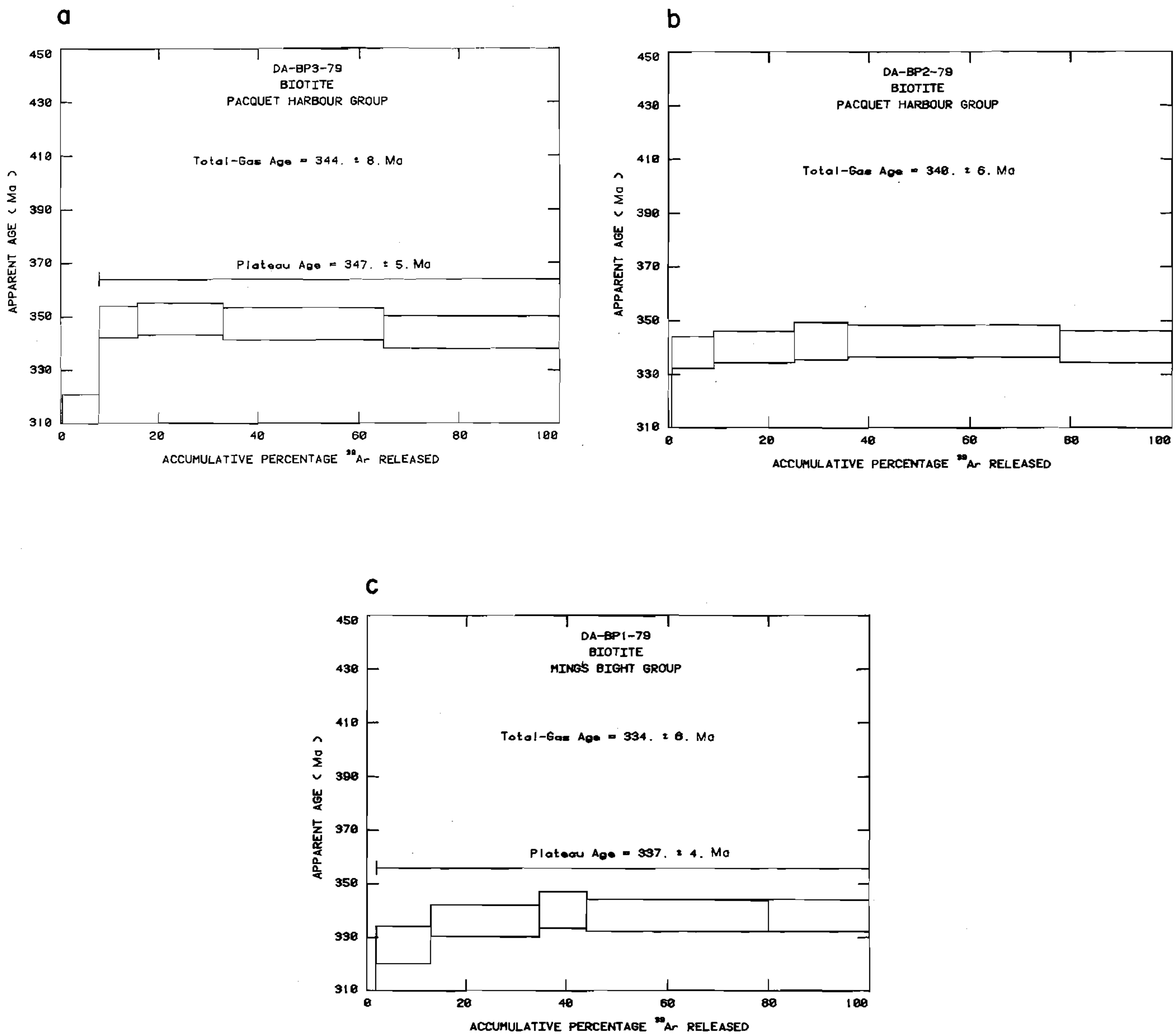

Figure AIV-6a-c: *$^{40}Ar/^{39}Ar$ age spectra for biotite samples from the Pacquet Harbour and Ming's Bight Groups.*

Table AIV-7: *Recalculated age dates used in this study.*

STRATIGRAPHIC UNIT	REFERENCE	REPORTED AGE (Ma)	RECALCULATED AGE (Ma)
Rb/Sr Ages:			recalculated using λ = 1.42
Dunamagon Granite	Pringle (1978)	413 ± 10	425 ± 10
		356 ± 10	366 ± 10
		358 ± 10	368 ± 10
		334 ± 10	343 ± 10
Burlington Granodiorite	Pringle (1978)	422 ± 10	434 ± 10
Micmac Lake Group	Pringle (1978),	375 ± 15	386 ± 15
	Neale & Kennedy (1967)	393 ± 24	404 ± 24
Cape St. John Group	Pringle (1978)	343 ± 15	353 ± 15
		429 ± 50	441 ± 50
Cape Brulé porphyry	Pringle (1978),	393 ± 25	404 ± 25
	Bell & Blenkinsop (1977)	325 ± 14	334 ± 14
Seal Island Bight Syenite	Bell & Blenkinsop (1977)	315 ± 25	324 ± 25
K/Ar Ages:			recalculated according to Mankinen and Dalrymple (1979)
Rattling Brook Group	Lowdon (1961)	355	362
Wild Cove Pond Igneous Suite	Lowdon (1961),	358	365
	Wanless et al. (1972)	384 ± 16	392 ± 16
Partridge Point Granite	Wanless et al. (1972)	361 ± 16	368 ± 16
Burlington Granodiorite	Lowdon et al. (1963)	373	380
Dunamagon Granite	Wanless et al. (1972)	348 ± 15	355 ± 15
$^{40}Ar/^{39}Ar$ Ages:			recalculated according to Mankinen and Dalrymple (1979)
East Pond Metamorphic Suite	Dallmeyer (1977)	386 ± 5	394 ± 5
Old House Cove Group	Dallmeyer (1977)	376 ± 5	383 ± 5
		380 ± 5	388 ± 5
		381 ± 5	388 ± 5
		392 ± 5	400 ± 5
		408 ± 5	416 ± 5
		413 ± 5	421 ± 5
		411 ± 5	419 ± 5
		420 ± 10	429 ± 10
Rattling Brook Group	Dallmeyer (1977)	366 ± 5	373 ± 5
		368 ± 5	375 ± 5
		386 ± 5	394 ± 5
		390 ± 5	398 ± 5

16:1

Figure 9-1: *Possible model for the geological evolution of the Baie Verte Peninsula.*

NFld/1497