THE BETTS COVE OPHIOLITE AND RELATED ROCKS OF THE SNOOKS ARM GROUP, NEWFOUNDLAND

H. D. UPADHYAY

THE BETTS COVE OPHIOLITE AND RELATED ROCKS OF THE SNOOKS ARM GROUP, NEWFOUNDLAND

by

H. D. Upadhyay

A Thesis
submitted in partial fulfilment of the
requirements for the degree of
DOCTOR OF PHILOSOPHY

MEMORIAL UNIVERSITY OF NEWFOUNDLAND

1973

FRONTISPIECE

A panoramic view of the Betts Cove ophiolite; Kitty Pond is to the left and Betts Cove to the far right. Ultramafic rocks to the west (left) are interpreted as upper mantle material which grade into intercalated gabbro and pyroxenite representing Moho. This is transitionally overlain by the dyke-bearing gabbro and sheeted dykes (layer 3), and by pillow lavas and sedimentary/pyroclastic rocks that are correlated with layers 2 and 1 respectively of the oceanic crust. Note the swing in the trend of dykes at the base of layer 3.

ABSTRACT

The Snooks Arm Group consists of a Lower Ordovician ophiolite suite at the base which is conformably overlain by two sedimentary/pyroclastic and two pillow lava formations. The ophiolite suite, best developed at Betts Cove, comprises a basal ultramafic member followed in order by a gabbroic, a sheeted dyke, and a pillow lava member, all of which have transitional contacts with each other. The ultramafic member consists of interlayered peridotite, pyroxenite, serpentinized dunite, and minor rodingitized gabbro. The gabbroic member consists of layered quartz gabbro, minor clinopyroxenite and quantitatively negligible diorite. The sheeted dyke member possesses over 90 percent diabase and ultramafic dykes that are subvertical and parallel to one another. The pillow lava member contains basaltic (spilitic) and ultramafic pillows and sills. The sedimentary/pyroclastic formations consist of andesitic agglomerate and flysch, deep-sea sediments and diabase sills. The two pillow lava formations of the upper part of the Snooks Arm Group comprise tholeiitic basalt and diabase dykes/sills that were derived from a different magma group.

The Betts Cove ophiolite is interpreted as part of a Lower Ordovician oceanic crust and mantle that was developed through sea-floor spreading and was subsequently thrust (obducted) on to the Fleur de Lys continental mass. Field and chemical data indicate that the ophiolite suite was produced through fractional crystallization of a single parent magma with a hiatus between the formation of the ultramafic-gabbro sequence and that of the sheeted dyke-pillow lava assemblage. The upper part of the Snooks Arm Group shows affinity to an island arc type

environment and might represent crust of a basin marginal to a major Lower Paleozoic ("Proto-Atlantic") ocean.

The Snooks Arm Group was subjected to low-grade burial metamorphism during Early Ordovician and to greenschist facies regional metamorphism during Devonian (Acadian) time.

The ophiolite suite is similar in many respects to the modern oceanic crust and to ophiolites elsewhere. The origin and tectonic evolution of the Snooks Arm Group is described in terms of a plate tectonic model related to the opening and closing of an ancient ocean.

Copper mineralization in Betts Cove area is concentrated at the contact of sheeted dykes and pillow lavas and is interpreted as co-magmatic with the mafic host rocks. Models are proposed for the magmatism and metallogeny of Betts Cove ophiolite that may help understanding of some of the processes operative on modern ocean floors.

ACKNOWLEDGMENTS

I am grateful to Professor E. R. W. Neale for introducing me to the geology of eastern Burlington Peninsula, for supervising this work, and for critically reading the thesis. This study of the Snooks Arm Group was initiated at the suggestion of Dr. M. J. Kennedy who shared with me his extensive knowledge of the structural evolution of ophiolitic and related rocks of the Burlington Peninsula and who critically evaluated a part of this thesis. Dr. D. F. Strong provided helpful discussions concerning the interpretation of chemical data and kindly corrected an earlier draft of this thesis.

Thanks are also due to Professor Harold Williams for informative discussions on the ophiolites of western Newfoundland and to Professor J. F. Dewey of Albany for originally calling my attention to some significant features of ophiolites in orogenic belts and also for useful suggestions concerning the mapping of Betts Cove - Tilt Cove area.

Professor V. S. Papezik, Dr. C. J. Hughes and Mr. R. K. Stevens provided helpful advice concerning the mineralogy and petrography of a variety of rock samples. My colleagues in the graduate program at Memorial University, notably W. R. Smyth, J. G. Malpas, B. E. Marten, and K. V. Rao provided stimulating discussion on many of the topics covered herein.

It is also a pleasure to thank the following persons: (i) Mrs. G. Andrews for chemical analyses; (ii) Foster Thornhill and Lloyd Warford for preparing thin sections; (iii) Wilfred Marsh for photographic services; (iv) B. Osbourne, W. Foote, C. Austin, and B. Ryan for capable and cheerful assistance in the field; (v) Alan Foote and family of Snooks Arm for hospitality and help; (vi) R. D. G. Prasad

for assistance in drafting work; (vii) Mrs. G. Bennett and Mrs. G. Woodland for typing services.

This work was supported by grants from the National Research Council of Canada (A-5540) and the Geological Survey of Canada (24-68) held by Professor E. R. W. Neale.

CONTENTS

Page

Page

CHAPTER VI

ECONOMIC GEOLOGY

LIST OF FIGURES

LIST OF TABLES

CHAPTER I

INTRODUCTION

A. LOCATION AND ACCESS

The thesis area is located on the southeastern margin of the Burlington Peninsula in northeastern Newfoundland. The Snooks Arm Group is exposed in a crescent-shaped area that stretches approximately from Betts Cove in the southwest to Tilt Cove in the northeast. This area, bounded on the east by the waters of Green Bay, is about 15 kilometers long and extends inland up to a maximum of 4 kilometers in the central region. Rocks of the adjacent Cape St. John and Nippers Harbour Groups are exposed to the north and west respectively.

The inhabited localities in the area are, from west to east, Nippers Harbour, Snooks Arm, Round Harbour, and Tilt Cove, all lying on the coast. Brents Cove and Harbour Round, just a few kilometers from Snooks Arm, are located in Confusion Bay to the north. All these localities are connected to each other by gravel roads and are mainly fishing communities. Snooks Arm, roughly in the centre of the thesis area, is about 50 kilometers from Baie Verte, which, in turn is connected to the Trans-Canada Highway by a 60-kilometer paved road.

Betts Cove, where the basal part of the Snooks Arm Group is best preserved and exposed, is only 7 kilometers from Nippers Harbour. Betts Cove is uninhabited and is connected with Nippers Harbour by an old trail which is now mostly obscured. The most convenient access is by boat from Nippers Harbour.

B. PHYSIOGRAPHY

B.1 Topography

The Betts Cove - Tilt Cove area is an undulating dissected plateau with an average elevation of 182 meters (600 feet). The plateau is marked by low parallel ridges with a few isolated hills that range up to 243 meters (800 feet) elevation. The ridges run northeast parallel to the strike of the rock units; intrusive and volcanic rocks generally stand out as ridges whereas serpentinite and sedimentary rocks occur in valleys. Ponds and lakes trend roughly northeast. Distinctly elongated ponds occur within the serpentinite belt. Some of the largest ponds occur in the basal pillow lava unit of the Snooks Arm Group.

The shoreline between Nippers Harbour and Beaver Cove is bounded by cliffs in a number of places. It is marked by various coves and fiord-like indentations, the most conspicuous among which is the three-kilometer long Snooks Arm. All these large coves have a northwest elongation. The fiords occur both along shear zones and along the strike of some sedimentary units.

B.2 Drainage

The drainage system is not well integrated. Streams that connect the ponds with the sea pass through steep slopes and cliffs and thereby form waterfalls. This occurs at Betts Cove, Snooks Arm, Venams Bight, and Beaver Cove.

B.3 Glaciation

Glacial erosion in the area has produced smooth and rounded rock surfaces (roche moutonnées), glacial striations, glacial drift,

erratics and a dearth of soil. U-shaped valleys and glacial cirques (e.g. Tilt Cove) occur in a few places. Although local exceptions exist, a predominantly southeastward movement of ice is indicated by glacial striations and other data.

B.4 Vegetation

The area under discussion is moderately to heavily wooded. Volcanic and shaly rocks support maximum vegetation. The valley slopes contain a thick growth of spruce, birch, fir, etc. The ultramafic rocks are conspicuously free of vegetation.

C. HISTORY OF GEOLOGICAL WORK

Geological interest in this area was shown as early as 1857 when Smith McKay discovered copper ore at Tilt Cove, and the first systematic geological study of the area was undertaken in 1864 by Alexander Murray of the Newfoundland Geological Survey. In 1864, Murray surveyed the Tilt Cove property and in 1867 he made a reconnaissance survey of the Betts Cove locality and mines. In 1872, E. Marret described the Tilt Cove mines which were then being exploited. J. P. Howley recorded some geological observations on the Betts Cove mine and a couple of prospects towards the southwest in 1878.

M. E. Wadsworth (1884) published some observations on the rocks and ores of Notre Dame Bay. Copper and nickel deposits of Tilt Cove were again described by J. Garland (1888). L. deLaunay (1894) compiled the observations of earlier workers on Tilt Cove mine and compared it with some Norwegian deposits.

In 1915, 1916, and 1919, Princeton University expeditions to Notre Dame Bay carried out studies of regional geology and ore deposits; Edward Sampson (1923), in particular, investigated the occurrence of ferruginous chert associated with pillow lavas and tuffs.

A. K. Snelgrove (1928) summarized all available data on the geology and ore deposits of Notre Dame Bay. He later published a paper dealing with the geology and ore deposits of Betts Cove - Tilt Cove area based on original field work that formed a part of his doctoral thesis (Snelgrove, 1931).

Douglas, Williams, Rove, and others (1940) compiled all information on the copper prospects of Newfoundland and described various base metal deposits of the Tilt Cove - Betts Cove area.

D. M. Baird, then with the Geological Survey of Newfoundland, did a reconnaissance mapping of the Burlington Peninsula during 1944-45 and published a map at 1 inch = 1 mile and a report on the geology of that region.

E. R. W. Neale, then with the Geological Survey of Canada, mapped the Snooks Arm Group and adjacent rocks and presented evidence for an "ambiguous" intrusive relationship of the Betts Cove - Tilt Cove serpentinite belt (Neale, 1957). Neale later compiled his and D. M. Baird's observations and prepared a detailed map of eastern Burlington Peninsula (Neale, 1958) and later a map of the entire Peninsula (Neale and Nash, 1963).

Neale and Kennedy (1967) described the relationship of the Fleur de Lys Group to younger Groups of the Burlington Peninsula. Church (1969) described the metamorphic rocks of the Burlington

Peninsula and reported the occurrence of ultramafic detritus and clasts of Fleur de Lys rocks in the Snooks Arm graywackes. Kennedy and Phillips (1971) discussed the ultramafic rocks of the Burlington Peninsula and, on the basis of their structural histories, divided them into two (Ordovician and Pre-Ordovician) age-groups.

The recent development of the concept of "Plate Tectonics" and more specifically Wilson's (1966) suggestion that the Atlantic Ocean once closed and then re-opened, has focused the attention of earth scientists on Newfoundland. Dewey (1969) was the first to discuss the evolution of the Appalachian - Caledonian orogen in terms of Plate Tectonics and sea-floor spreading. Later, on the basis of spatial and chronologic relationships between bulk stratigraphic units and tectonic events, Bird and Dewey (1970) suggested a lithosphere plate model for the evolution of the Appalachian orogen.

In Betts Cove an ophiolite suite consisting of ultramafic rocks, gabbro, sheeted dykes, and pillow lavas was independently recognised by Upadhyay *et al.*, (1971) and by Church and Stevens (1971). Dewey and Bird (1971), in a comprehensive discussion of the origin and emplacement of ophiolites, described all the Newfoundland ophiolites, including the Betts Cove - Tilt Cove occurrences.

D. THE PRESENT STUDY

This thesis deals with the stratigraphy, petrochemistry, metallogeny, and origin of the Snooks Arm Group, with special emphasis on its basal ophiolite suite. This group is interpreted as marginal to an ancient ("Proto-Atlantic") oceanic crust and a comparison with

models of the present-day ocean floors is made in the light of Plate Tectonics. A brief account of those parts of the Cape St. John, Beaver Cove, and Nippers Harbour Groups, that lie in contact with the Snooks Arm Group, is also presented.

The author spent three field seasons mapping and collecting the field data. Air photos at 4 inches = 1 mile scale were used as base maps. Much of the mapping was done from coastal exposures.

CHAPTER II

REGIONAL GEOLOGY

A. GENERAL STATEMENT

The area lies within the Newfoundland Appalachians at the northeastern extremity of the Appalachian Mountain Belt of North America. The Newfoundland Appalachians are broadly divided into three northeasterly trending tectonic zones (Williams, 1964a) as shown in Figure 1, inset: (i) the Western Platform comprising Precambrian basement rocks overlain by Cambro-Ordovician carbonates and other continental shelf facies, (ii) the Central Mobile Belt underlain by "eugeosynclinal" volcanic, pyroclastic, and sedimentary rocks of Ordovician age, (iii) the Avalon (Eastern) Platform consisting of Precambrian to Ordovician sediments and silicic volcanic rocks. The Central Mobile Belt and a part of the Avalon Platform are intruded by granitic rocks of Devonian age. The thesis area is part of the Burlington Peninsula of northeast Newfoundland which lies within Williams' (1964a) Central Mobile Belt.

The Burlington Peninsula is predominantly underlain by two types of rocks ("terrains"): (i) in the west mostly silicic, polyphase-deformed "continental" rocks and (ii) in the east dominantly volcanic - pyroclastic "oceanic" rocks that show similarities with those elsewhere around Notre Dame Bay. The age of these rocks ranges from Late Precambrian to Devonian (Neale and Kennedy, 1967).

A summary of the geology of Burlington Peninsula is in order

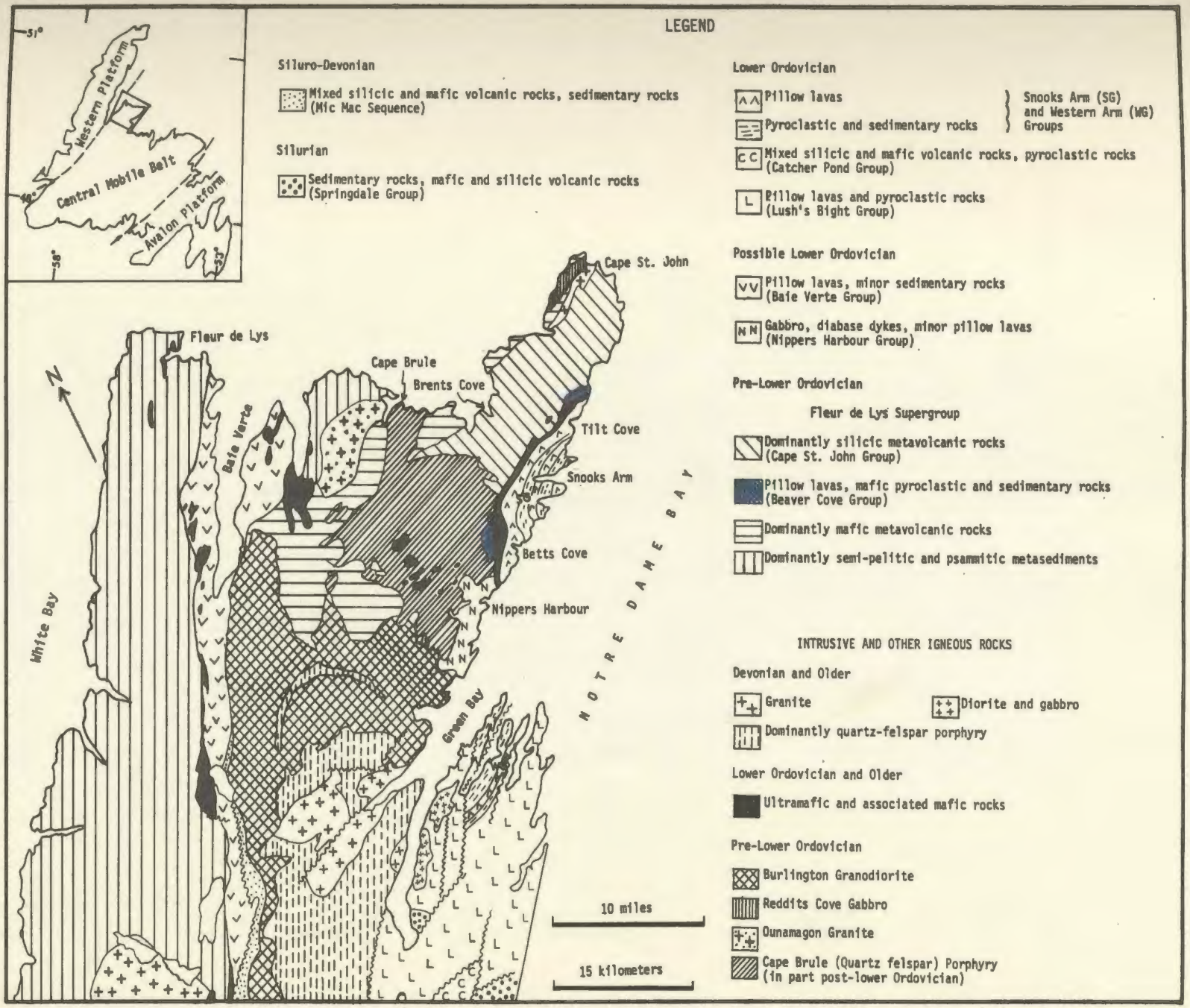


Figure 1: Generalized geological map of the Burlington Peninsula, Newfoundland

because many of the problems of the thesis area are closely associated with those of the adjacent region to the west.

B. EOCAMBRIAN - CAMBRIAN

B.1 Fleur de Lys Supergroup

The Fleur de Lys Supergroup forms the most extensive rock assemblage in the Burlington Peninsula. It has been described in recent years by Neale and Kennedy (1967), Church (1969), Kennedy (1971), and Kennedy and Phillips (1971). It is exposed in two main belts that are separated by the northeast-trending Baie Verte lineament. The western belt consists of psammitic and semi-pelitic schists with minor amphibolites, conglomerates, and marble. The eastern one contains mainly silicic flows and porphyries with minor mafic volcanic rocks. The western belt of rocks overlies gneissic basement of presumed Grenvillian age (M. DeWit, personal communication, 1972).

In the area to the west of Baie Verte lineament (Western belt) the Fleur de Lys Supergroup is divided into White Bay and Rattling Brook Groups, divisions first used by Betz (1948) and Watson (1947) respectively and retained by Church (1969). Within Watson's Rattling Brook Group, in the type Fleur de Lys area, Kennedy (1971) has recognised three main structural units, White Bay Sequence, Harbour Sequence, and Eastern Sequence that are separated from each other by tectonic slides. These are intruded(?) by mafic (now amphibolitic) and ultramafic rocks that have a structural and metamorphic history similar to their host rocks (Kennedy and Phillips, 1971). In the Flatwater Pond - Westport area, Church (1969) has divided the Fleur

de Lys Supergroup into seven formations, the lithologies of some of which can be correlated with Kennedy's (1971) "Sequences" in the north.

In this western belt, folding has caused repetitions of various units. The succession is shown to be generally westward facing (Kennedy, 1971) but at one locality eastward-facing units are known (Church, 1969).

In the eastern belt, east of the Baie Verte lineament, the Fleur de Lys Supergroup consists of four groups: Mings Bight, Pacquet Harbour, Beaver Cove, and Grand Cove/Cape St. John. The Mings Bight Group is exposed between Mings Bight and Pacquet Harbour and consists of massive psammite, banded quartzite, conglomerate and mafic schist. The Pacquet Harbour Group, which crops out to the southwest of Pacquet Harbour village and along the coast between Confusion Bay and Cape St. John (Kennedy and Phillips, 1971), comprises metavolcanic rocks including pillow lava, siliceous and mafic volcanic fragmental rocks, tuffaceous and minor graywacke-like rocks. At Pacquet Harbour village, this group is reported to conformably overlie the Mings Bight Group (Church, 1969).

The Beaver Cove Group consists of undeformed to moderately deformed pillow lava, andesitic flysch and pyroclastic rocks, chert, argillite and sandstone. At the basal part it has a fault contact with the Snooks Arm Group. Towards the top it is transitionally overlain by the silicic flows and agglomerates of the Cape St. John Group. The latter also contains variable amounts of amygdaloidal mafic flows and quartz-felspar welded tuffs. Parts of these two groups were mapped by the author and a more detailed account is presented later in

this chapter.

The Fleur de Lys Supergroup underwent three main phases of deformation but the effect was different in different types of rocks (Kennedy, 1971; Williams *et al.*, 1972). The structures were best developed in metasediments, amphibolites and the margins of some of the intrusive rocks. The first deformation, which was preceded by tectonic slides, produced minor tight, isoclinal folds (F_1) and associated schistosity (S_1). The F_1 folds are now almost obscured except in some minor F_2 fold hinges. This, after minor crenulations of S_1, was followed by the second (D_2) deformation that produced northward-facing tight major and minor folds (F_2) with S_2 schistosity in the western belt and southward-facing recumbent folds in the eastern belt. This deformation was perhaps the most penetrative and produced the present major structures of the area. The nearly opposing facing-directions of F_2 folds in west and east indicate that they were either developed on opposite limbs of an even larger (F_1) fold or that the F_2 folds form part of a nappe complex and face outward, away from its symmetry axis. The third deformation (D_3) was semi-penetrative and produced strain-slip fabrics and minor folds. Large folds related to this deformation are rare except at Pacquet Harbour and near the shores of Baie Verte where it refolds some of the major F_2 folds. D_3 was post-dated by faults, including the Baie Verte fault or lineament, kink bands, and crenulations. Church (1972, Fig. 1) however contends that there are two anticlinoria and two intervening synclinoria in the Burlington Peninsula with northeasterly trending axes that continue through the whole length of the Appalachians.

Metamorphism in the Fleur de Lys Supergroup is generally in the amphibolite facies. Garnet and plagioclase grew between the first and second deformations; some garnet and plagioclase grew syntectonically with D_2. Post-D_2 kyanite and staurolite occur in the Mings Bight Group. Micas and amphiboles show a prolonged history of growth. The presence of polyphase deformed Fleur de Lys rocks as detritus in the Lower Ordovician Snooks Arm Group dates the metamorphism and deformation as Pre-Ordovician (Church, 1969).

C. ORDOVICIAN AND POSSIBLE ORDOVICIAN

C.1 Snooks Arm Group

The graptolite-bearing Lower Ordovician Snooks Arm Group (Snelgrove, 1931; Neale, 1957) is exposed approximately between Tilt Cove and Betts Cove. This group, with the Betts Cove ophiolite suite at its base (Upadhyay *et al.*, 1971), consists predominantly of pillow lava with minor sedimentary and pyroclastic rocks. The Snooks Arm Group is the subject of this thesis and its various aspects will be discussed in the following chapters.

C2. Western Arm Group

In Notre Dame Bay, rocks similar to those of the Snooks Arm Group are exposed between Green Bay and Halls Bay where they are called the Western Arm Group (MacLean, 1947; Marten, 1971). This group consists of pillow lava, tuff, agglomerate and sedimentary rocks; a single specimen of the brachiopod *Discotreta* sp. indicating an Early Ordovician (Arenig) age was found by McLean (1947). The Western Arm Group is underlain, with a transitional contact, by the

Lush's Bight Group which contains mostly pillow lavas with minor pyroclastic rocks. The Lush's Bight Group has been interpreted as layer-2 of a "Proto Atlantic Ocean" by Bird and Dewey (1970) and, on structural grounds, these authors have assigned a Pre-Ordovician age to this group. More recently Strong (1972) with the discovery of "sheeted" dykes, and Smitheringale (1972) with the establishment of the low-K nature of the tholeiites, have strengthened the suggestion that the Lush's Bight Group, at least in part, does represent a remnant of an ancient ocean floor.

C.3 Baie Verte Group

Pillow lavas, pyroclastic, and sedimentary rocks that underlie a linear belt extending south-southwestward from Baie Verte, have been referred to as the Baie Verte Group (Watson, 1947; Neale and Nash, 1963) and assigned an Early Ordovician age on the basis of lithological similarity to the dated Snooks Arm Group. From a structural study, Kennedy and Phillips (1971) have suggested that the ultramafic belt at the base of this group may be of Pre-Ordovician age.

The Baie Verte Group is faulted against the Fleur de Lys rocks in the west; to the east, the Siluro-Devonian Mic Mac Sequence is interpreted to conformably underlie it and form the eastern limb of a syncline, the axis of which was later modified by a fault (Neale and Kennedy, 1967). The Baie Verte rocks are steeply dipping to the east. At Mings Bight this group contains "sheeted" dykes and the whole assemblage is interpreted as an ophiolite suite representing a fragment of an ancient ocean floor (R. Norman, personal communication, 1973).

C.4 Nippers Harbour Group

A predominantly gabbroic complex with associated dykes and minor pillow lavas, exposed along the coast southwest of Nippers Harbour village, was named "Nippers Harbour Group" and assigned a probable Ordovician age (Baird, 1951; Neale, 1958). It is similar to a part of the adjacent Betts Cove ophiolite. A more detailed description of this group is presented later in this thesis.

C.5 Deformation

The Snooks Arm and Western Arm Groups have been subjected to one major episode of deformation and have one penetrative cleavage. Metamorphism is within the lower greenschist facies. The ultramafic rocks of the Baie Verte belt, according to Kennedy and Phillips (1971), have undergone polyphase deformation.

D. SILURIAN

On the Burlington Peninsula, rocks mapped as Silurian occur near the head of Southwest Arm in Green Bay. These include red conglomerate, arenite, siltstone, and minor amounts of wackes, shales, limestone, mafic and silicic lavas that are referred to as the Springdale Group (Espenshade, 1937; MacLean, 1947; Kalliokoski, 1953). Rarity of pillow structures and the presence of terrestrial sediments are in strong contrast to the predominantly pillow lava-chert-graywacke assemblages of the Ordovician of Newfoundland.

E. SILURO-DEVONIAN

The Mic Mac Sequence, a linear belt of sediments and mixed silicic and mafic volcanic rocks has yielded a 393 million year Rb/Sr

whole rock isochron age and is probably of late Silurian or Early Devonian age. Lithologically, rocks of the Mic Mac Sequence strongly resemble those of the Springdale Group although locally they are more intensely deformed than the Springdale rocks (Neale and Nash, 1963). On the eastern side, the Mic Mac Sequence unconformably overlies the Burlington granodiorite and on the west it has a faulted contact with the Baie Verte Group (Neale and Kennedy, 1967). The Mic Mac sediments face westward and have a strong penetrative cleavage and a second, locally developed, strain slip fabric.

F. ROCKS OF MAGMATIC ORIGIN

F.1 Ultramafic and Associated Mafic Rocks

These rocks occur mainly in three places: (i) as a linear discontinuous belt running from Baie Verte towards the southwest, (ii) at the west and south shores of Mings Bight, and (iii) as an arcuate belt between Betts Cove and Tilt Cove. Some of these occurrences are now known to contain "sheeted" dykes and are interpreted as remnants of an old ocean floor. Pods of ultramafic rocks also occur west of Fleur de Lys village and within quartz-felspar porphyry northwest of Betts Cove. Although these ultramafic rocks are highly altered, primary layering can still be seen in many places.

Nearly all the ultramafic rocks were considered to be of Ordovician age by earlier workers but recently Kennedy and Phillips (1971), on the basis of structural studies, have suggested that there are two distinct ages of ultramafic rocks and that some are pre-Ordovician. This problem will be further discussed elsewhere in this thesis.

F.2 Gabbro

A polyphase deformed gabbro (Reddits Cove Gabbro) is exposed between Reddits Cove and Cape St. John, east of La Scie. From structural correlation, this is equivalent to the Fleur de Lys rocks, i.e. Pre-Ordovocian in age.

F.3 Granodiorite

A large granodiorite intrusion (Burlington Granodiorite) is exposed to the northwest of Green Bay. The occurrence of granodiorite clasts in the conglomerates of the (?Ordovician) Baie Verte Group and an unconformable contact with the overlying (Siluro-Devonian) Mic Mac Sequence shows that the Burlington Grandiorite is pre-Baie Verte (?pre-Ordovician) in age. In the northeast it is known to intrude the Pacquet Harbour Group of the Fleur de Lys Supergroup before D_1 deformation, further suggesting a pre-Ordovician age of the pluton. Compositionally, the Burlington Granodiorite ranges from granite to quartz diorite, with granodiorite as the predominant phase.

F.4 Silicic Rocks

Quartz-felspar porphyry occurs in two main outcrop-areas. The first one extends from Pacquet Harbour in the north to Nippers Harbour in the south. It includes pink to grey porphyry with quartz and K-felspar phenocrysts, silicic tuffs, agglomerates and minor rhyolites. Equivalents of this rock are also associated with the Cape St. John Group in the east. This porphyry appears to be a composite pluton of two different ages. In the north it has the same structural history as the Fleur de Lys Supergroup (Coates, 1970) suggesting a pre-Ordovician age. In the south, west of Kitty Pond, the Lower Ordovician

ultramafic rocks of the Betts Cove ophiolite are intruded by the porphyry, showing that some of it is post-Lower Ordovician (?Silurian/Devonian) in age. The position and nature of contact between the two "types" of porphyries is not known.

The other quartz-felspar porphyry is exposed on, and inland from, the shore of Green Bay. An arcuate dyke, convex northwards, occurs within the Burlington Granodiorite. The porphyry also includes silicic flows and pyroclastic rocks. In its north-central part syenite and quartz syenite are exposed. This porphyry and its associated volcanic rocks are interpreted as Silurian and possibly related to volcanism in the Mic Mac Sequence and Springdale Group (Neale and Nash, 1963).

The Dunamagon Granite, exposed between Mings Bight and Pacquet Harbour, is reported to have undergone the same stages of deformation as rocks of the surrounding Fleur de Lys Supergroup (Williams, *et al.*, 1972).

G. ROCKS ADJACENT TO THOSE OF THE SNOOKS ARM GROUP

The author mapped those parts of Beaver Cove Group, Cape St. John/Grand Cove Group, Nippers Harbour Group and Cape Brulé Porphyry that are exposed next to the Snooks Arm Group. A brief account of this study follows.

G.1 Beaver Cove Group

G.1.1 Field Relationships

This term was introduced, but not adequately defined, by Dewey and Bird (1971). The Beaver Cove Group is here defined as an

assemblage of undeformed to moderately deformed pillow lava, andesitic flysch and pyroclastic rocks, chert, argillite, and sandstone that are exposed in the area between Beaver Cove, Great Caplin Cove and Goss Pond (Fig. 2); the type-section occurs along the coast between Beaver Cove and Great Caplin Cove. This group is faulted against a thick diabase sill of the Snooks Arm Group in Beaver Cove: elsewhere the contact is not exposed. The top of this group is marked by a transitional and conformable contact with the overlying Cape St. John rhyolites and tuffs; the transition is represented by alternating mafic volcanic rocks and calcareous sandstone units, best seen on the cliff face along the north shore of Beaver Cove Pond. The pillow lava units gradually decrease in thickness and number stratigraphically upward towards the Cape St. John Group.

An assemblage of gabbro, andesitic flysch, pyroclastic rocks, and isolated pillow lava, exposed to the west of Betts Big Pond and Kitty Pond (Fig. 2), is identical to that occurring in Beaver Cove. It has a fault contact with the Betts Cove ultramafic rocks. The author incorporates them within the Beaver Cove Group and suggests that this group may be a part of the Nippers Harbour Group, the two probably formed an ophiolite suite prior to their displacement by faults and younger intrusions. To the northwest, all these rocks are adjoined by rhyolites and silicic porphyries, the latter at one place intrude the ultramafic rocks.

The rocks of the Beaver Cove Group dip moderately towards the north. Their northward younging directions are indicated by bun-shaped pillows, current-bedded sandstones, and some graded sediments

in Beaver Cove. Some of the units exposed on the coast cannot be traced inland to the west of Beaver Cove Pond.

G.1.2 Stratigraphy and Rock Types

The Beaver Cove Group consists of pillow lava, chert, argillite and andesitic flysch and pyroclastic rocks that show strong similarity to the Snooks Arm lithologies. In the upper part it consists of massive mafic flows, conglomerate and current-bedded sandstone that mark the transition towards the overlying Cape St. John Group. The pillows are light grey to greyish-green in colour and contain interstitial chert. On higher stratigraphic levels, where associated with sandstone, they are deformed and slightly flattened.

The massive mafic volcanic rocks are maroon coloured and amygdaloidal. They consist of highly saussuritized plagioclase in an epidote-rich turbid groundmass with abundant carbonate and hematite, the latter causing the maroon colour of the rock. Some less altered plagioclase crystals have the composition of An_{35}.

The andesitic pyroclastic rocks and flysch constitute a significant part of the Beaver Cove Group. Bedding in these rocks is poor or absent except where marked by some thin chert and argillite units. The pyroclastic rocks consist of pale-green lumps in a greyish-green groundmass containing pyroxene or, less commonly, amphibole crystals up to 1 centimeter. The pale-green lumps contain highly saussuritized, rarely zoned, plagioclase crystals in an epidote-rich turbid groundmass; the greyish-green groundmass has very little epidote. Highly irregular shaped inclusions of chert and argillite are also common. Where the pyroxene crystals are small and even-grained, the

rock resembles diabase in hand specimen. The flysch is similar to the one described from an island-arc in New Hebrides (Mitchell, 1970).

The sandstones are light to purple grey, thinly bedded and in places show current bedding. They consist of even-grained quartz clasts in a calcareous and sericitic matrix. Chromite grains in these rocks were reported by Neale (1957) and also observed by the author. The conglomerates contain boulders, average size 5 centimeters, predominantly of quartz, chert, and pink rhyolitic material. The boulders are highly flattened parallel to bedding. Chromite grains were observed in one of them.

G.1.3 Deformation

The Beaver Cove Group has one penetrative cleavage, best developed in the calcareous sandstone. This cleavage dips northward and is subparallel to the bedding. A second, less pervasive, strain-slip cleavage is present only in the sandstones and some lapilli tuffs that occur near the contact zone with the Cape St. John Group. This cleavage is subvertical in the vicinity of Beaver Cove Pond. Green schist facies metamorphism is suggested by epidote, chlorite and actinolitic amphibole.

The Beaver Cove Group can be correlated with the Pacquet Harbour Group in the north, which contains deformed pillow lava, tuffs, volcanic fragmental rocks, and graywacke-like rocks (Neale and Kennedy, 1967; Church, 1969) metamorphosed to amphibolite facies. The difference between the two groups is in the degree of deformation and grade of metamorphism, both gradually decrease southwards.

G.2 Cape St. John Group

G.2.1 Field Relationships

This name was first used by Baird (1951) to include a sequence of lava flows with interbedded sedimentary and pyroclastic rocks exposed along the coast between Beaver Cove and Cape St. John and towards the southwest to Sister Ponds. A part of this sequence had been referred to as Goss Pond Volcanics and Red Cliff Pond Volcanics by Snelgrove (1931). Baird (1951, p. 36) considered this group to overlie the fossil-bearing Lower Ordovician Snooks Arm Group and to be "lithologically more similar to the Ordovician rocks of Notre Dame Bay than to any others in northern Newfoundland," and hence he assigned them an Ordovician age. Later workers, e.g. Neale (1958) and Williams (1967), on the basis of lithological similarity, correlated them with some Silurian rocks of Newfoundland.

Church (1969) termed some polyphase deformed silicic rocks exposed in the vicinity of Confusion Bay, as Grand Cove Group and included them in the Fleur de Lys Supergroup. He used the term Cape St. John Group for the rocks exposed to the north of Tilt Cove and assigned them a Siluro-Devonian age but did not clearly indicate their geological relations with the Grand Cove Group (Church, 1969, Fig. 2). The present author studied the rocks in the Great Caplin Cove-Beaver Cove area and the road-side exposures between Red Cliff Pond and Brents Cove and concluded that the entire area has undergone polyphase deformation, although the structures are not developed in all rocks in all places and the deformation, in general, becomes less pervasive towards the south. It is therefore considered as misleading to use

the two names Grand Cove Group and Cape St. John Group for essentially the same sequence of rocks. The author retains the term Cape St. John Group because: (a) This name has been in use for over two decades and, although the structural studies in recent years suggest a pre-Ordovician age, the geographical distribution of this group still remains practically unchanged and (b) the name Grand Cove Group was introduced by Church (1969, p. 220 and Fig. 2) apparently to separate out a more obviously deformed part of this sequence which he interpreted as older than the rest of the Cape St. John Group. As his evidence is not considered valid, it seems preferable to drop the term Grand Cove Group.

The Cape St. John Group is exposed in a northeast-trending nearly 7 X 24 kilometers area, bounded in the north by the Pacquet Harbour Group (Kennedy and Phillips, 1971), in the west by the Cape Brulé Porphyry, in the south by the Snooks Arm and Beaver Cove Groups, and in the east by the water of Notre Dame Bay (Fig. 1). The major stratigraphic and structural trends are northeasterly, concordant with those of the Pacquet Harbour and Snooks Arm Groups. The contact with the Cape Brulé Porphyry in the west is nearly at right angles to this trend.

The contact between Cape St. John and Snooks Arm Groups, in the area between West Pond and the eastern end of Red Cliff Pond, is marked by a narrow lineament, and some of the quartz-felspar porphyry outcrops on the Cape St. John side are also abruptly truncated at this lineament. This contact is interpreted as a fault by the author. A somewhat similar contact exists between the western end of Long Pond and Supply

Pond. The Cape St. John Group has conformable contacts with the Pacquet Harbour Group in the north (Kennedy and Phillips, 1971; Church, 1969) and Beaver Cove Group in the south.

The age relationship between Cape St. John Group and Burtons Pond/Cape Brulé Porphyry is at least partly intrusive. This is suggested by the following features:

(a) At Kitty Pond, the Porphyry intrudes the Betts Cove ultramafic rocks of post-Cape St. John age and, as mentioned above, there may be two distinct ages of this porphyry.

(b) Tongues and pods of the porphyry occur within the Cape St. John Group that are best seen to the north and west of Red Cliff Pond (Neale, 1958).

(c) Some of the conglomerate-like diatremic rocks that occur within the Cape St. John Group (as intrusions?) also form a part of this porphyry to the west of Kitty Pond.

G.2.2 Stratigraphy and Rock Types

Most units of the Cape St. John Group die out along strike and therefore different section-lines show somewhat different rock-successions. The structural base of the Cape St. John Group at Red Cliff Pond is formed by a poorly sorted breccia with fragments of red cherty argillite and mafic igneous rocks that are cemented by a calcareous material. The rock has one penetrative cleavage and lacks any definite bedding. This breccia shows a transitional contact with the overlying purple-green volcanic rocks. Alternating amygdaloidal and non-amygdaloidal horizons within these volcanic rocks suggest a dip of about 35° towards the north. Thinly laminated lenses of pink

mudstone within these flows also indicate a similar dip, but elsewhere in this area dips as steep as 50° have been observed. The volcanic rocks consist of highly saussuritized plagioclase, carbonate, and a hematitic opaque material. Pillow structures are rare or absent in them. These are overlain successively by rhyolitic tuffs and a conglomerate-like rock, probably of intrusive origin. From here to Brents Cove the rocks are predominantly silicic flows and agglomerates with minor mafic volcanic rocks.

No definite stratigraphic tops are known in the area between Red Cliff Pond and Brents Cove. From the conformable contact with the underlying Beaver Cove Group, which faces and dips northwards, it is inferred that the Cape St. John Group, at least in the southern part, also faces northwards.

G.2.3 Deformation

The Cape St. John Group carries evidence of two main phases of deformation. The schistosity related to the first deformation is folded by a second deformation that gave rise to a series of recumbent F_2 folds, best seen at Brents Cove. The F_2 folds are not so well developed in the area adjacent to the Snooks Arm Group, although a second, steep, north-dipping cleavage is seen in some silicic tuffs. Most Cape St. John lithologies, however, remain unchanged all the way from Brents Cove to the base of the Snooks Arm Group.

The ages of the two deformations in the Cape St. John Group are significant factors in interpreting the tectonic evolution of the area. Church (1969) reported clasts of metamorphosed and polydeformed rocks of the Fleur de Lys Supergroup in the Snooks Arm sediments,

implying that the deformations took place in pre-Ordovician times. The author found clasts of undeformed Cape St. John rhyolites in the Snooks Arm conglomerates but no clasts with definite polyphase structures were noticed. Furthermore it is noticed in the Tilt Cove - Beaver Cove area that the main cleavage in the Snooks Arm Group dips moderately to steeply northward and is vaguely concordant to that of the Beaver Cove Group and perhaps with the second cleavage of the Cape St. John silicic tuffs. The possibility that the second deformation of the Cape St. John Group and the main deformation of the Snooks Arm Group are related to the same post-Ordovician (?Acadian) episode, can not be entirely ruled out. A comparable situation prevails to the northwest (R. Norman, personal communication, 1973) where one deformation in the Fleur de Lys Supergroup is reported to grade into the (?Ordovician) Baie Verte ophiolite.

G.3 Nippers Harbour Group

G.3.1 Field Relationships

The Nippers Harbour Group (Baird, 1951; Neale, 1958) is exposed along the coast between Pittman Bight in the northeast and Northwest Arm and beyond in the southwest. This group, a part of which was studied by the author (Fig. 2), consists predominantly of gabbro, diabase dykes (locally "sheeted dykes")* with minor isolated ultramafic rocks and pillow lavas.

---

*The term "sheeted dykes", probably first introduced by Gass (1968) but long used as a stratigraphic term in Cyprus, is used here to denote an assemblage consisting of over 90 percent parallel dykes with very little or no country rocks.

The Nippers Harbour Group is intruded by the Burtons Pond Porphyry and it is faulted against the Betts Cove ophiolite suite. Between Pittman Bight and Kitty Pond, this fault is marked by a deep vegetated gully. It is best exposed in the Nippers Harbour Islands where it consists of a 15 meter wide, highly sheared, almost chloritic zone of basic rocks. Here, the Betts Cove ophiolite is represented by dykes with brown-weathering ultramafic screens, and the Nippers Harbour Group to the south of the fault contains gabbro and subordinate dykes.

G.3.2 Stratigraphy and Rock Types

The Nippers Harbour Group is herein interpreted as part of an ophiolite suite, probably the zone where gabbro grades into overlying sheeted dykes.

Approximately 70 percent of the outcrop-area is underlain by a leucocratic to mesocratic gabbro with massive diabasic phases. It is locally layered and also contains pods of pyroxenite. Pegmatitic gabbro is not uncommon. It consists of actinolite, saussuritized plagioclase, chlorite with minor epidote, quartz and magnetite. Less altered plagioclase crystals have a composition in the oligoclase-andesine range. Actinolite occurs as fibrous interlocking grains with diffuse outlines, most probably secondary after pyroxene. In some cases it is partly or wholly altered to chlorite. Gabbro from the contact zone (with the Betts Cove ophiolite) has amphibole and quartz grains that are bent and show undulatory extinction.

The dykes are highly variable in concentration as well as orientation. Where concentration is great, they locally form a sheeted complex within the surrounding gabbro. Horizons consisting

exclusively of sheeted dykes, such as at Betts Cove or Troodos, do not exist in the Nippers Harbour ophiolite. Compositionally, most of the dykes are diabasic but red-weathering uralitized pyroxenites also occur.

Pillow lavas occur as small isolated outcrops northwest and north of Burtons Pond. They are intruded by, and engulfed in, the Burtons Pond Porphyry. Constituent minerals include chlorite, actinolite-tremolite, saussuritized plagioclase, carbonates and minor pyroxene, quartz and opaques.

G.3.3 Deformation

The intensity of deformation and the resulting structures in the Nippers Harbour Group vary from place to place. In some parts, especially those adjacent to the Betts Cove ophiolite, a strong deformation is indicated by quartz-rich and amphibole/chlorite-rich segregation bands, bent and stretched amphibole crystals and flattened quartz grains. In a number of places, on the other hand, the rocks carry no evidence of having undergone any deformation.

Dewey and Bird (1971) have assigned a pre-Ordovician age to the Nippers Harbour Group because the latter is intruded by the Burtons Pond Porphyry which, according to them, is a part of the Cape St. John Group (Fleur de Lys Supergroup). The author has observed that the Burtons Pond Porphyry also intrudes the Lower Ordovician Betts Cove ultramafic rocks and hence the porphyry can not be a part of the pre-Ordovician Cape St. John Group. However, a pre-Ordovician age for the Nippers Harbour Group is inferred from the following: The gabbro, pillow lavas, and the andesitic flysch of Beaver Cove Group, exposed

to the west of Betts Big Pond and Kitty Pond, may be a part of the mainly gabbroic Nippers Harbour Group. The former assemblage was probably isolated from the Nippers Harbour Group during the emplacement of the Burtons Pond Porphyry as demonstrated by some pillow lava outcrops now completely engulfed in the porphyry. Since the flysch-pillow lava assemblage of the Beaver Cove Group is conformably overlain by the Cape St. John Group in Beaver Cove, the Nippers Harbour Group can, therefore, be tentatively assigned a pre-Cape St. John age.

From the structural and lithological similarity with the nearby Betts Cove ophiolite, on the other hand, a possible Lower Ordovician age for the Nippers Harbour Group can not be completely ruled out. It is not unlikely that it represents a faulted part of the Betts Cove ophiolite although it is too thick to be correlated with the anomalously thin gabbro member of the Betts Cove ophiolite. The author however assumes this group to be of pre-Ordovician age but wishes to emphasize the tentative nature of this assumption.

G.4 Cape Brulé and Burtons Pond Porphyries

G.4.1 Field Relationships

A large area, to the northwest of Betts Cove, extending from Cape Brulé in the north to Rogues Harbour in the south is underlain by quartz-felspar porphyry and its "contaminated facies" (Neale, 1958). The eastern boundary of this porphyry runs nearly at right angles to the eastward trending stratigraphic and structural units of the Cape St. John Group. To the south, it intrudes the Nippers Harbour Group on all scales ranging from hand specimen size to mappable apophyses. Its contact with the ultramafic rocks of the Betts Cove ophiolite, in

most places, is marked by a deep vegetated gully that stands out as a well-defined lineament on air photos. Near the northern shore of Kitty Pond, however, a tongue of the porphyry intrudes the Betts Cove ultramafic belt, giving rise to an arcuate ultramafic xenolith. This relationship shows a post-Lower Ordovician (Devonian?) age for at least some of the porphyry.

In the Cape Brulé area, on the other hand, the porphyry carries evidence of having undergone polyphase deformation suggesting a pre-Ordovician age (Coates, 1970). It is therefore conceivable that the quartz-felspar porphyry exposed between Cape Brulé and Rogues Harbour is a composite pluton belonging to two different ages. The author tentatively maintains this view and refers to the older as "Cape Brulé Porphyry" and the younger as "Burtons Pond Porphyry"; the latter term was first used by Snelgrove (1931, p. 24) to describe the "granite porphyries" of the Betts Cove - Tilt Cove area.

Demarcation of the boundary between possibly two separate and distinct porphyries needs further investigations. A possible criterion could be the presence or absence of mafic-ultramafic xenoliths. The porphyry in its southern exposures contains small but mappable pods of pillow lava, mafic and ultramafic rocks. Based on the field relationship in outcrops near the ophiolite boundary, these pods are interpreted as xenoliths of pre-existing ophiolites. The porphyry in the northern parts of its outcrop area is not known to contain such xenoliths. Judging from Neale's map (1958), the boundary between the xenolith-bearing and xenolith-free porphyries is a nearly east-trending line that passes through Peridot Lake, about 11 kilometers northwest of

Nippers Harbour.

A part of the Burtons Pond Porphyry studied by the author consists of the following two rock types.

G.4.2 Quartz-Felspar Porphyry

It is a medium-grained rock with individual crystals averaging 3 millimeters in size. Some coarse varieties contain quartz/felspar grains up to 1 centimeter across. The rock is generally creamy-pink, greyish-pink or greenish-buff. Inclusions of foreign rocks are common and, where heavily concentrated, give rise to a conglomerate-like rock. Quartz and felspar are present in approximately equal amounts. Both have resorbed edges and are commonly embayed by the matrix material. Felspar is predominantly orthoclase with minor microcline and albite, all partially sericitized. Orthoclase shows Baveno and Carlsbad twins. The matrix consists of fine-grained quartz, sericitized felspars with subordinate carbonates and opaques.

G.4.3 Quasi-Conglomerate

Rocks originally interpreted as Cape St. John conglomerates (Neale, 1957) outcrop to the west of Kitty Pond - Betts Cove area. This rock occurs as pods within the quartz-felspar porphyry and shows transitional contacts with it. These pods do not seem to have any definite pattern or shape. A gradual increase in the amount of such pod-like "inclusions" results in a rock which consists of over 70 percent quasi-conglomerate or breccia. At the west shore of Neale's Pond (Fig. 2) they show excellent layering, individual layers being 30 to 60 centimeters thick. Elongate fragments are aligned parallel to the bands.

The fragments are angular to well rounded varying from 1 centimeter to 30 centimeters in size. They are made up dominantly of rhyolites, and also include andesitic volcanic rocks, amphibolite, pyroxenite, serpentinite, gabbro, silicic porphyry etc. These fragments occur in a somewhat similar quasi-clastic matrix.

Quasi-conglomerate and silicic agglomerate, showing similar gradational contact with the quartz-felspar porphyry, also occur in the Red Cliff Pond - Lond Pond area where they are spatially associated with the Cape St. John Group. Best exposures are on the west shore of Goat Pond where ellipsoidal units of jasper and quartz-felspar porphyry, up to 0.5 meters long, occur within a quartz-felspathic matrix. Unlike those of the Neale's Pond area, the variety of rock fragments is very restricted here; 80 percent of the inclusions are of jasper.

The following features suggest a magmatic origin for these quasi-conglomerates:

(1) The occurrence within and transitional contacts with the quartz-felspar porphyry of undisputed magmatic origin.

(2) Close spatial association with rhyolites.

(3) Tongues of quasi-conglomerate intruding pillow lavas.

From the patchy outcrop pattern it seems that the magmatic material and the associated xenoliths were brought up through somewhat narrow vents. These vents probably also acted as feeders to the nearby rhyolites. The layering observed in some places may have developed in response to the sloping flanks of such explosive vents. The perfect roundness in some inclusions can be explained by a mutual attrition in a hot environment that would ultimately result in the smoothing of

corners and edges (cf. Reynolds, 1954).

The mafic/ultramafic fragments in the conglomerate are here interpreted as xenoliths of Nippers Harbour and Betts Cove ophiolites. The rhyolitic fragments could be genetically related to these diatremes or tuffisite pipes.

G.4.4 Deformation

The Burtons Pond Porphyry has one cleavage. This cleavage in some cases augens around the phenocrysts and gives the impression of being a composite fabric. Cleavage is, however, not developed in all parts of the porphyry. Thus coarse-grained varieties and some of the quasi-conglomerates or tuffisites, e.g. those near Neale's Pond, show slight or no evidence of having undergone deformation.

CHAPTER III

THE SNOOKS ARM GROUP

A. INTRODUCTION

The Snooks Arm Group is exposed in a crescent-shaped area in southeastern Burlington Peninsula. Snelgrove (1931) referred to it as "Snooks Arm Series" in which he included three volcanic and two sedimentary/pyroclastic units of Lower Ordovician age. Baird (1951) and Neale (1957) have used the term "Snooks Arm Group" for the same sequence of rocks; this term has been retained by all the subsequent workers. An arcuate ultramafic belt runs parallel to, and along the base of, this volcano-sedimentary assemblage and was interpreted as an intrusive rock by earlier investigators. Recent work (Upadhyay *et al.*, 1971) however shows that at Betts Cove this ultramafic belt is successively overlain by gabbro, sheeted dykes, and pillow lava. The contacts between these rock units are highly transitional and together they form a single suite of rocks--an ophiolite suite. Since the pillow lava portion of this suite forms the basal unit of the "Snooks Arm Group" of earlier workers, it follows that the Snooks Arm Group be redefined to include the ophiolite suite.

The present work shows that the Snooks Arm Group extends from Pittman Bight in the southwest to Beaver Cove in the northeast, a distance of about 20 kilometers. This area has a maximum width of about 4.5 kilometers in its central part. The base of the redefined Snooks Arm Group is marked by a fault in most places. In the area

between Red Cliff Pond and the western end of Long Pond the nature of this base can not be ascertained due to lack of outcrops.

The Snooks Arm Group is about 5.7 kilometers thick. It has a southeasterly dip that generally varies from 35° to 85°; in the Red Cliff Pond - West Pond area the beds are overturned and dip to the northwest. The author divides the Snooks Arm Group into five formations. The basal (Betts Cove Ophiolite) formation is further divided into four members. Names, lithologies, and thicknesses of various formations appear in Table 1.

B. THE BETTS COVE OPHIOLITE

B.1 General Statement

This formation includes, from the base upwards, an Ultramafic Member, a Gabbroic Member, a Sheeted Dyke Member and a Pillow Lava Member. The highly transitional contacts between any two members warrant their grouping as a single igneous formation.

All members of the Betts Cove Ophiolite are exposed in the area enclosed by Pittman Bight, Betts Cove, and Betts Big Pond (Fig. 2); this area is hereafter referred to as the "Betts Cove area". The type-section lies along the Kitty Pond - Betts Cove - Betts Island join. The Ultramafic Member is layered for the most part, the layers having a moderate dip to the east. This is transitionally overlain by an irregular and poorly developed Gabbroic Member which is locally layered. The Sheeted Dyke Member (sheeted complex) consists almost entirely of parallel to subparallel dykes, with individual dykes having an average thickness of about 0.5 meter. In general, the dykes are vertical to

TABLE 1

LITHOSTRATIGRAPHIC CLASSIFICATION OF SNOOKS ARM GROUP

Formation Name	Lithology	Thickness (meters)
Round Harbour Basalt	Pillow lavas, diabase dykes/sills	1000
Balsam Bud Cove Formation	Sedimentary rocks, subordinate pyroclastic rocks, diabase sills, minor pillow lavas	750
Venams Bight Basalt	Pillow lavas, diabase dykes/sills	500
Bobby Cove Formation	Sedimentary and pyroclastic rocks, diabase sills, minor pillow lavas	500
Betts Cove Ophiolite	Pillow Lava Member Sheeted Dyke Member Gabbroic Member Ultramafic Member	3000

subvertical in attitude and perpendicular to the trend of layers in the underlying Ultramafic and Gabbroic Members. At its base the sheeted complex locally grades into the underlying members. This complex is overlain by the Pillow Lava Member with a highly transitional contact. The Pillow Lava Member, which dips about 45° to the east, strikes nearly perpendicular to the trend of dykes of the underlying sheeted complex. The Pillow Lava Member is conformably overlain by the Bobby Cove Formation.

Between Betts Cove and the central part of Long Pond in the northeast, the Ultramafic Member is faulted against the Pillow Lava Member, which has resulted in the omission of gabbro and sheeted dykes. Ultramafic rocks are highly altered in this region and remnants of layering, dipping southeastwards, are seen only in the area immediately to the north of Betts Big Pond. The Ultramafic Member is missing, presumably due to faulting between Red Cliff Pond and the western end of Long Pond. Towards the northeastern shore of Red Cliff Pond a tabular outcrop of highly altered ultramafic rock, surrounded on all of its exposed sides by the Cape St. John Group, shows some evidence of having undergone a similar structural history to the surrounding pre-Ordovician Cape St. John Group. For this reason it is not interpreted as part of the ophiolite suite under discussion even though the two appear to be linked in outcrop pattern.

All members of the Betts Cove Ophiolite re-appear in the northeast in the area between the central part of Long Pond and Beaver Cove, referred to as the "Tilt Cove area" in this thesis. However, this part of the ophiolite suite differs from that of the Betts Cove

area in the following respects: (1) Excessive thinning of the whole ophiolite suite to about 750 meters as compared to 3000 meters in Betts Cove, (2) presence of dyke breccia instead of sheeted dykes, and (3) outcrop patterns highly modified by faults. Layering in the ultramafic rocks and gabbro is scarce or nearly absent, and the pillow lavas are sheared and chloritized. A near-vertical attitude for the ophiolite suite is inferred from the subvertical to vertical sedimentary units that conformably overlie the Pillow Lava Member. The contacts are generally concordant except that the Gabbroic and Pillow Lava Members wedge out into each other near Winser Lake. Dykes, both within the poorly developed sheeted complex and the Gabbroic Member, are oblique to the trends of the ophiolite stratigraphy. Outcrop patterns in the Tilt Cove area are greatly modified by faults. An undulatory northwest-trending fault between Tilt Cove and Beaver Cove, for instance, truncates the Pillow Lava Member. Another set of faults causes a similar truncation of a thick diabase sill exposed at, and to the west of, Beaver Cove.

The ophiolite suite is better developed in the Betts Cove area and also the rocks are less altered than in the Tilt Cove area. Description and discussion of this formation are, therefore, concentrated on exposures in the Betts Cove area.

B.2 The Ultramafic Member

B.2.1 Field Relationships

In the Betts Cove area the Ultramafic Member includes about 25 percent of the area underlain by entire ophiolite suite. The maximum thickness, around Kitty Pond, is about 0.75 kilometer. The Betts Cove

area is the only region between Pittman Bight and Beaver Cove where the Ultramafic Member shows primary layering, primary textures, and minimum alteration.

Between Betts Big Pond and Red Cliff Pond, the Ultramafic Member has a consistent width of about 150 meters. 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 meters in the Tilt Cove area and then gradually thins out towards Beaver Cove Pond. Tongues of ultramafic rock that are in continuity with the Ultramafic Member, run into the older Beaver Cove and Cape St. John Groups. At Supply Pond and to the west of Beaver Cove Pond (Fig. 2), this is probably due to the remobilization and re-intrusion of the Ultramafic Member. Similar remobilization has been reported from the Troodos ophiolite in Cyprus (Moores and Vine, 1971, p. 446). In the Tilt Cove area, especially along the road between Winser Lake and Beaver Cove Pond, the basal contact of the Ultramafic Member with the Beaver Cove Group is marked by a narrow zone of "blackwall" alteration; this alteration of the mafic pillows of this group has produced a greenish-black, highly chloritic rock. Similar alteration is associated with the processes of serpentinization and rodingitization in ultramafic terrains (Cady *et al.*, 1963).

The ultramafic rocks between Betts Big 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. The talc-carbonate rock weathers to a rusty-brown colour. Carbonates occur as patches and reticulating veins. Pods and veins of quartz occur generally in association with talc-carbonate rock. On weathered surfaces, due to resistance to weathering, quartz stands out in relief. Specular hematite occurs as tiny flakes.

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 Betts Cove area the ultramafic rocks occur in four forms: (i) Intrusive breccia, (ii) layered sequence (iii) serpentinites and (iv) pegmatitic veins and dykes.

B.2.2 Intrusive Breccia

(a) Field Relationships and Petrography

The intrusive ultramafic breccia is exposed to the west of Kitty Pond, which is the basal part of the Ultramafic Member (Fig. 2). The breccia crops out as irregular isolated pods up to 150 meters across. It 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 (Fig. 3). The fragments, which are mostly angular to subangular, range in size from a few millimeters to over half a meter; sorting is very poor.

Figure 3: Outcrop of ultramafic breccia showing transitional contact between rubbly breccia (right) and smooth-weathering serpentinite. West of Kitty Pond.

The coarser types are 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. It consists almost exclusively of serpentine with magnetite, carbonates, and minor chromite. The carbonates are generally confined to the interspaces among large patches of serpentinites. Clusters of fresh clinopyroxene, although rare, occur in some samples.

(b) Origin

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. It should be noted that gabbro breccia, dyke breccia and pillow breccia also occur as isolated pods and are restricted to the Gabbroic, Sheeted Dyke and Pillow Lava Members respectively. This, coupled with the absence of any "foreign" rock fragments, indicates a very local origin of this material within various members of the ophiolite suite. It is suggested that the breccia was brought up through isolated small vents by some kind of gaseous activity that fluidized and brecciated the surrounding ultramafic material to give rise to the breccia.

Somewhat similar but much more extensive breccia has been

described from Musa Valley, Eastern Papua (Green, 1961) and from Vourinos ophiolite complex (Moores, 1969, p. 36). In the former the breccia also includes chalcedony, quartz, andradite, garnierite, magnesite, and chrysotile. Green (1961), on the basis of detailed mineralogical and chemical evidence, suggests that in Musa Valley the ultramafic breccia was produced by the penetration, brecciation and local fluidization of peridotitic country rock by volcanic gases.

B.2.3 The Layered Sequence

(a) Field Relationships

The layered ultramafic rocks are exposed in the vicinity of Kitty Pond and to the south of Axe Pond. In a gross way the layered sequence forms the upper part of the Ultramafic Member. The thickness of layers varies from about one millimeter to several meters (Figs. 4, 5). 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 of peridotite occur to the south of Kitty Pond; pyroxene crystals get gradually smaller within a vertical distance of 0.5 meter or less (Fig. 6). Reverse gradation, i.e. coarse crystals on top and fine at

Figure 4: Thinly laminated clinopyroxenite (high relief) and serpentinized dunite. North shore of Kitty Pond.

Figure 5: Layered clinopyroxenite (grey; bluish tint due to film) and peridotite (coarse, brown-weathering). Note the faint size-gradation in the peridotite layers. Near the outlet of Kitty Pond, Atlantic Ocean in background.

Figure 6: Size-gradation in wehrlite layer. The clinopyroxene crystals gradually decrease in proportion and the rock grades into pyroxene-bearing dunite towards top (upper right).

Figure 7: Cross-laminated ultramafic rock. The light material is serpentinized dunite, dark is pyroxene-bearing dunite. North shore of Kitty Pond.

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 (Fig. 7). Slump and primary pinch-and-swell structures are also seen in the layered sequence (Fig. 8). In these, the individual layers pinch and swell or abruptly merge into another rock type along strike.

The trend of the ultramafic layers is extremely variable. Abrupt sharp swings within a short distance along the strike is a common feature. These varying attitudes of the ultramafic layers do not occur in other members of the ophiolite suite. The dip of the ultramafic layers varies between 20° and 75°. Abrupt changes in dip are commonly encountered.

The thickness, composition, and other properties of individual layers were recorded by the author in two parts of the layered sequence, one on the eastern side of Kitty Pond and the other to the west of Joey's Pond (Fig. 9). The total thicknesses of these two sections are about 200 and 240 meters respectively. The top of both sections is close to or only slightly below the base of the Gabbroic Member. In Section A of Figure 9 the thickness and position of each layer has been precisely recorded whereas in Section B, some parts of which consist of closely spaced thin layers, only the proportion of various rock types is presented without locating the position of individual thin layers within that part of the sequence. In view of extensive alteration, the identification of these rocks in the field is largely based on texture and weathering habits.

Figure 8: Primary pinch-and-swell structure in the layered ultramafic rocks. South of Axe Pond.

From Sections A and B, the following observations can be made.

(1) Both sections show rhythmic layering (cf. Poldervaart and Taubeneck, 1960). Recurrence of various rock types at different levels is a common feature.

(2) Gabbro is nearly absent in Section A and only appears, rather abruptly, at the base of the Gabbroic Member (not shown in the section). In Section B, on the other hand, it constitutes about 7 percent of the whole section and first appears nearly at its base. The contact between Ultramafic and Gabbroic Members is more transitional in B than in A.

(3) In general, layers are thicker towards upper, pyroxenite-rich parts of the sections.

(b) Mineralogy and Rock Classification

Pervasive alteration, particularly of olivine and orthopyroxene, poses a serious problem in the classification of Betts Cove ultramafic rocks. Pseudomorphs of different minerals, however, carry certain distinctive characters that are helpful in classifying the altered rocks. Primary textures are still preserved in most of them. These remnant textures and the extent of alteration result in different types of weathered surfaces that characterize various ultramafic rocks. Dunite, for instance, weathers to featureless smooth surfaces. Peridotites are distinguished by rough surfaces due to the greater susceptibility of serpentinized olivine to weathering than the more durable pyroxenes. Pyroxenites have moderately smooth surfaces with clearly visible crystals. Fresh pyroxenites weather to a greenish-grey colour, reddish hues appear where the extent of alteration is greater. In places

extremely sharp contacts exist between greenish-grey and reddish-brown pyroxenite layers. The greenish-grey are the resistant clinopyroxenites and the reddish-brown are interpreted as more susceptible orthopyroxenites. Remnants of unaltered orthopyroxene are in some cases recognizable in the latter.

Microscopic study bears out the fact that orthopyroxenes are more intensely altered than clinopyroxenes. Sharp contacts between less altered clinopyroxene and talcose orthopyroxene is a common observation. Distinction between the pseudomorphs after olivine and orthopyroxene can be made provided the rock is not heavily serpentinized. Olivine pseudomorphs are equant with round shape and irregular fractures (Fig. 10). Orthopyroxene, on the other hand, has somewhat elongate subhedral crystals and the pseudomorphs show some suggestion of remnant cleavage without any irregular fractures (Fig. 11). The common alteration products of orthopyroxene are tremolite-actinolite, bastite, and talc.

Rocks consisting of olivine, orthopyroxene, and clinopyroxene generally show cumulate textures (Fig. 10). Olivine occurs as cumulate, and orthopyroxene/clinopyroxene as intercumulate. Orthopyroxene, where associated with only clinopyroxene occurs as cumulate. This relationship, once established in unaltered rocks, can also be applied to altered cumulates.

Magnetite content generally increases with increasing serpentinization. Chromite has a maximum concentration in olivine-rich rocks and as the proportion of pyroxenes increases, the chromite content declines. Pyroxenites, for example, contain little or no chromite.

Figure 10: Photomicrograph showing cumulate serpentinized olivine (with black magnetite-bearing veins) and talcose orthopyroxene (grey) surrounded by intercumulate clinopyroxene (black). Plane polarized light (X40).

Figure 11: Photomicrograph showing bastite pseudomorphs after orthopyroxene. The pseudomorphs are altered to talc along the margins. Note the remnant cleavage in bastite. Plane polarized light (X40).

Chromite occurs as isolated crystals, in contrast with magnetite which forms clusters or trails of tiny granules, especially in the more serpentinized varieties. In some samples it constitutes up to 1 percent of the rock, but distinct chromitite bands such as those in the Bay of Islands pluton of western Newfoundland (Smith, 1958), are not known in Betts Cove. Under transmitted light it looks reddish brown against magnetite which is generally brownish black. A number of chromite grains are divided by irregular fractures with matching borders on both sides. Growth of serpentine minerals along the fractures and replacement of chromite is not uncommon, leaving euhedral pseudomorphs where such replacement is extensive.

The classification of the various ultramafic rocks of Betts Cove is shown diagrammatically in Figure 12. Description of these rock types follows.

(c) Dunite

Dunite weathers to saffron or reddish brown smooth surfaces. Its layers are characteristically thin, rarely exceeding a thickness of 60 centimeters. In Sections A and B (Fig. 9) dunite has total thicknesses of about 18 and 11 meters which are, respectively, about 9 and 5 percent of the respective sequences.

The rock is an olivine-pyroxene cumulate with over 90 percent olivine. It is completely altered to fine-grained serpentine which, in some specimens, consists of pseudomorphs after olivine (Fig. 10). The average size of olivine crystals is about 1.5 millimeters. Olivine pseudomorphs contain irregular veinlets that are filled with magnetite granules. Pyroxenes occur as an intercumulate and show complete to

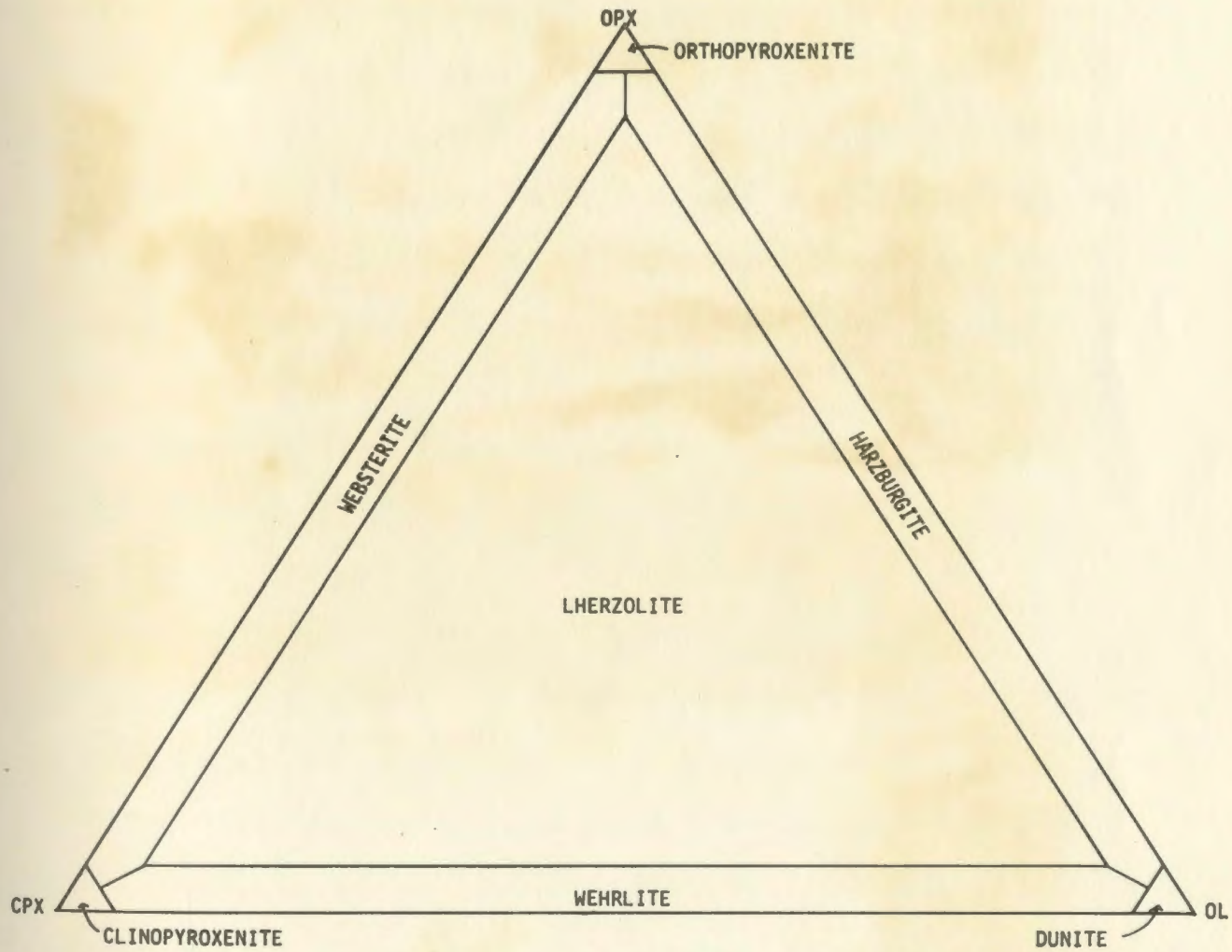


Figure 12: Triangular diagram showing the mineral composition of various ultramafic rocks described from Betts Cove. (cf. Coleman, 1971b).

OPX Orthopyroxene
CPX Clinopyroxene
OL Olivine

partial alteration.

(d) Pyroxenite

Pyroxenite constitutes the major part of the layered ultramafic sequence. In Sections A and B, it has total thicknesses of 130 meters (64 percent) and 140 meters (58 percent) respectively. It is predominant in the upper part of both sections. Where alteration is pervasive, reliable distinction can not be made between orthopyroxenite and clinopyroxenite. Single layers of pyroxenite are as much as 12 meters thick; further subtle layers also exist within such thick layers.

Pyroxenite commonly contains pegmatitic phases of similar composition. Grain gradation is not as common in pyroxenite as it is in peridotite. The longer axes of pyroxene crystals are generally aligned parallel to the layering.

Orthopyroxenite layers are intimately associated with those of clinopyroxenite and websterite, and occur at almost every level of the layered ultramafic sequence in Betts Cove. Orthopyroxenite forms reddish brown grooves against greenish grey unaltered clinopyroxenite that stands out as narrow ridges. With increasing clinopyroxene content the rock locally grades into websterite. Streaks of clinopyroxenite, as thin as 1 millimeter, are commonly associated with orthopyroxenite.

The orthopyroxene crystals are seen in thin sections to be completely altered to tremolite, bastite, serpentine, and talc. Remnants of original orthopyroxene within the pseudomorphs can be seen in some specimens. Optical characters suggest an enstatitic composition for these remnant orthopyroxenes. They occur as cumulates, with

clinopyroxene (up to 10 percent) associated as an inter-cumulus phase. The clinopyroxenes generally stand out in relief and exhibit sharp contacts against the altered orthopyroxenes. Where clinopyroxene also has low optical relief due to alteration, the cumulus texture is best seen under crossed nicols. The clinopyroxene alters to a low-relief actinolitic amphibole and less commonly to a nearly isotropic chlorite. The orthopyroxenes are euhedral with sharp outlines except in highly altered samples where the crystals are subrounded and show diffuse boundaries (Fig. 11).

Magnetite occurs as isolated crystals and in some cases as clusters around the margins of altered orthopyroxenes.

Clinopyroxenite forms layers that vary in thickness from less than a millimeter to more than a couple of meters. In general the rock is even-grained (average grain size 2-3 millimeters), although crystals ranging in size from a fraction of a millimeter to over a centimeter have been noted. The clinopyroxenite grades into wehrlite and websterite in some places, with increasing amounts of olivine and orthopyroxene.

The clinopyroxenite, unlike other ultramafic rocks, lacks any cumulate textures. The rock is granular, with somewhat diffuse crystal-outlines caused by alteration of crystal margins. Unaltered clinopyroxenes are colourless and contain prominent partings. Twinning, in some cases polysynthetic, is common. A number of crystals show undulatory extinction.

The clinopyroxenes show partial to complete alteration to tremolitic amphibole, serpentine, talc, and less commonly chlorite.

Remnants of unaltered minerals indicate that the alteration took place in different stages: first to amphibole, then serpentine and finally talc. The unaltered clinopyroxene remnants have a distinctive higher optical relief, birefringence and extinction angle than the surrounding altered material. All the minerals altered from clinopyroxene show gradational contacts with one another.

Websterite consists of altered, reddish weathering orthopyroxene and greenish grey fresh clinopyroxene. Cumulate textures between the two pyroxenes are commonly seen in thin sections which are not easily identifiable in handspecimens. Orthopyroxene pseudomorphs are euhedral and occur as cumulates. The intercumulate clinopyroxene forms a network pattern, marked by high optical relief.

(e) Peridotite

The term peridotite is used here to include those ultramafic rocks that consist predominantly of olivine and lesser amounts of clinopyroxene and/or orthopyroxene. In Sections A and B, it has the total thicknesses of about 56 meters (28 percent) and 75 meters (30 percent) respectively. Peridotite also occurs beyond the layered sequence in the Betts Cove area, as inferred from the textures even though the diagnostic minerals are not recognizable in these highly altered rocks. Varieties of peridotite can not always be distinguished from one another in the field, but they can be easily distinguished from dunite and pyroxenite by their cumulate textures and generally coarse, rough-weathering surfaces. In Betts Cove, the best cumulate structures and textures are developed in peridotite. The three varieties of peridotite--harzburgite, wehrlite, and lherzolite--are

described below.

Harzburgite is an olivine-orthopyroxene cumulate. Orthopyroxene crystals are up to 2 centimeters in dimension. In hand specimen the orthopyroxenes or their pseudomorphs show shining cleavage planes in which the dull-looking serpentinized olivine cumulates are poikilitically enclosed. Olivine is more deeply weathered than the pyroxene.

The proportions of olivine and orthopyroxene are generally equal in most samples, although there are significant departures from this. Clinopyroxene is scarce or virtually absent. Olivine occurs as elliptical crystals, poikilitically enclosed within orthopyroxene. Orthopyroxene is altered to bastite, tremolitic amphibole, talc and less commonly mesh-structure serpentine.

Wehrlite consists primarily of olivine and clinopyroxene with very little or no orthopyroxene. The proportions of the first two minerals may vary even within the same layer and thus names such as "pyroxene dunite" and "olivine pyroxenite" may be appropriate for varieties that locally contain exceedingly high proportions of olivine and pyroxene respectively.

Average crystal size is 5 millimeters in the wehrlites although there may be wide variations in grain size. The rock generally has a granular appearance with the resistant clinopyroxene crystals standing in relief against a deeply weathered olivine-rich (serpentinized) groundmass. In some varieties spongy clinopyroxene crystals up to 2 centimeters across contain tiny elliptical blobs of serpentinized olivine.

Olivine and clinopyroxene, as seen in thin sections, occur as cumulate and intercumulate respectively. The granular appearance, seen in handspecimen, is caused by evenly distributed, equant, high-relief clinopyroxene areas surrounded by numerous tiny olivine blobs.

Lherzolite consists of olivine, orthopyroxene, and clinopyroxene. Having similar texture and weathering surfaces as harzburgite and wehrlite, it can not be easily distinguished from these two in handspecimen. Gradations into harzburgite and wehrlite occur in several places.

Lherzolite is essentially an olivine-orthopyroxene cumulate. In some of the typical thin sections olivine, orthopyroxene and clinopyroxene constitute 50-60 percent, 25-30 percent, and 15-20 percent respectively. Olivine has round shapes, which is altered to a mesh-structure serpentine (Fig. 13). Orthopyroxene, which is almost completely altered to talc, has a tabular shape and the pseudomorphs carry remnant cleavage. Clinopyroxene always occurs as intercumulate with some alteration to tremolitic amphibole. This amphibole, which is mostly colourless, grades into a pleochroic brown amphibole similar to the one reported by Melson *et al.*, (1967) from the Mid-Atlantic Ridge (Fig. 14).

B.2.4. Pegmatitic Veins and Dykes

Veins and dykes of pegmatitic pyroxenite, although constituting a quantitatively negligible part, occur at nearly every level of the Ultramafic Member. The width of the veins varies from about 1 centimeter to over 60 centimeters. Within the layered sequence the veins generally run oblique to the layers. Bifurcation of single veins or

Figure 13: Photomicrograph showing mesh-structure serpentine cut by a younger vein that carries the fibrous "cross-slip" variety of serpentine. Note the sheaf-like flakes oriented at right angles to each other in the mesh-structure serpentine. Crossed nicols (X80).

Figure 14: Photomicrograph of a lherzolite sample showing intercumulate brown amphibole (secondary after clinopyroxene) grading into colourless amphibole. Plane polarized light (X40).

cross-cutting between different veins is not uncommon. The size of individual crystals in the veins ranges from less than 5 millimeters to 30 centimeters. The most conspicuous exposure of the pegmatitic veins occurs at Pittman Bight. Here the light grey weathering veins and dykes, over 60 centimeters wide, cut massive serpentinites (Fig. 15). Several thick cross-cutting veins may locally constitute a major part of the outcrop, so that the serpentinite country rock looks like inclusions in it. The pyroxene crystals at this particular locality are up to 30 centimeters long and show partial alteration to a tremolitic amphibole. An X-ray diffraction pattern shows that it is diopside.

Where the veins are thin, the pyroxene is almost completely altered to actinolitic light-green amphibole. The crystals and their alteration products are generally oriented perpendicular to the walls of the veins.

The coarse-grained texture, nearly monomineralic composition, and the random orientation of the veins indicate that the latter were probably emplaced during late stages of the consolidation of the surrounding ultramafic rocks. The clinopyroxene, which is most resistant to weathering, still remains very little altered whereas the surrounding rocks are completely serpentinized. Bowen and Tuttle (1949) explain the origin of such dykes as due to SiO_2 - saturated water vapour streaming through cracks in peridotite and converting the peridotite to pyroxenite.

B.2.5 Serpentinite

Those ultramafic rocks in which the primary textures and

Figure 15: Pegmatitic clinopyroxenite dyke (top of photo) cutting serpentinite. Pittman Bight.

structures are completely obliterated due to extensive alteration, are grouped here as "serpentinite". These are exposed in two main outcrop-areas, one to the north of Burtons Pond and the other to the south of Kitty Pond (Fig. 2). Small isolated outcrops of serpentinite also occur on the southwestern and northern shores of Betts Big Pond. Serpentinite, along with talc-carbonate rock, also constitutes the bulk of the Ultramafic Member between Betts Big Pond and Tilt Cove. It generally weathers to creamy grey and various shades of brown. Fresh surfaces are dark green to greenish black. In the vicinity of fault zones the serpentinite is highly sheared. Massive serpentinites in some cases show remnant cumulate textures.

In thin sections the rock consists of serpentine and some talc. Magnetite is intimately associated with the serpentinites. It occurs as trails and aggregates of tiny grains. Talcose facies contain very little or no magnetite; it is so heavily concentrated in serpentinite that it renders some grains completely opaque.

A more detailed account of the serpentine minerals and related alteration is presented below.

B.2.6 Alteration of the Ultramafic Member

(a) Serpentinization

The serpentinization of the Betts Cove Ultramafic Member has resulted in the development of mainly three types of serpentine minerals: (i) bastite (ii) flaky mesh-structure serpentine, and (iii) fibrous "cross-slip" serpentine occurring in veins. Bastite occurs as light green, weakly pleochroic crystals mostly pseudomorphing pre-existing pyroxenes (Fig. 11). It shows long and parallel fibers

which probably represent the prismatic cleavage or (010) parting of the pyroxene. It has nearly parallel extinction and first-order grey/yellow polarization colours. Due to a variable degree of alteration, patches of different birefringence are commonly seen within the same crystal. Patches and needles of tremolitic amphibole within bastite show that the latter was formed by the alteration of the amphibole which, in turn, was derived from some pyroxene. The amphibole shows a strikingly higher optical relief and birefringence than the enclosing bastite. Bastite alters to talc with very little magnetite associated with it.

The flaky mesh-structure serpentine occurs in highly altered rocks. Unlike bastite, very little or no suggestion of any remnant cleavage/parting exists in the pseudomorphs. It consists of an aggregate or sheaf-like rossettes or flakes that are oriented grossly at right angles to one another, giving rise to a mesh-structure (Fig. 13). These flakes have very low birefringence. Textures suggest two modes of formation of the flaky serpentine: (i) by further alteration of bastite in which the serpentine patches are free from veins and contain virtually no magnetite and (ii) by alteration of olivine in which the flakes contain innumerable irregular veins full of magnetite grains.

The fibrous "cross-slip" variety of serpentine occurs along veins that cut across nearly every mineral including bastite, mesh-structure serpentine, fresh pyroxenes, fractured chromite,etc. (Fig. 13). This indicates a younger age for this serpentine. The width of the veins varies from almost submicroscopic to several millimeters. A number of these veins are faulted on a microscopic scale and new veins

are emplaced along such fault planes. This suggests that the formation of these veins and that of the serpentine they contain, took place in more than one phase. The colour of the serpentine ranges from pale white, light green to yellow. The fibers run at right angles to the length of the veins. The latter in places contain one or more trails of tiny magnetite granules that show a remarkable parallelism with the walls of the veins, indicating that the origin of magnetite was closely related to serpentinization both in space and time.

Optical characters and form of the fibrous serpentine suggest that it could be the variety chrysotile (Deer *et al.*, 1962; Aumento and Loubat, 1971). Such minor cross-cutting veins have been interpreted as a non-stress product of a late alteration after pervasive serpentinization of ultramafic rocks (Coleman, 1971b, p. 907).

(b) Steatitization and Related Alteration

Talc and carbonate are the end-products of the alteration of ultramafic rocks in the Betts Cove - Tilt Cove area. In the Betts Cove occurrences talc is intimately associated with serpentine and can be seen only in thin sections. Between Betts Big Pond and Red Cliff Pond, the Ultramafic Member consists almost entirely of talc and Mg-Fe carbonates. Despite such alteration, the remnant banding and cumulate textures are still preserved in some places.

Talc occurs as submicroscopic flakes, the aggregates of which look somewhat turbid. Larger flakes grow within bastite crystals. The growth of talc generally initiates from the margins of the serpentine patches. Some rocks, perhaps originally orthopyroxenites, consist of over 90 percent talc pseudomorphs. Unlike serpentinization, talc

alteration does not involve the extensive growth of magnetite. Carbonates, although not common in Betts Cove area, show a spatial association with talc.

(c) Discussion

Experiments of Bowen and Tuttle (1949) and Kitahara *et al.*, (1966) on the system $MgO-SiO_2-H_2O$ show that serpentinites can not be formed at temperatures above 500°C. Bowen and Tuttle did not observe any liquid phase in this system at temperatures up to 1000°C and pressure of 15,000 lbs/in^2 and consequently ruled out the possibility of the existence of a serpentine magma. Serpentinization, therefore, is a solid-state process in which olivine and pyroxene are altered to hydrous magnesian silicates through addition of water.

Serpentine can be formed by the reaction between forsterite and water below 400°C or by addition of silica to forsterite (Deer *et al.*, 1962, p. 185):

$$2Mg_2SiO_4 + 3H_2O \rightarrow Mg_3Si_2O_5(OH)_4 + Mg(OH)_2$$

forsterite serpentine brucite . . . . . (I)

$$3Mg_2SiO_4 + SiO_2 + 2H_2O \rightarrow 2Mg_3Si_2O_5(OH)_4 \quad \text{. . . . .(II)}$$

There is no concrete evidence for any potential source of silica in Betts Cove in order to meet the requirement of reaction (II). Reaction (I) may therefore represent the serpentinization of olivine-rich ultramafic rocks of Betts Cove ophiolite. The presence of brucite, demanded by reaction (I), was not observed by the author. This apparent lack of brucite might be explained by Coleman's (1971b, p. 908) observation that brucite is uncommon in those lherzolites in

which MgO/SiO_2 ratio is characteristically low;* some of the analysed serpentinized ultramafic samples (Table 2) from Betts Cove do have a relatively low (0.59 to 0.80) MgO/SiO_2 ratio.

Talc, which is intimately associated with serpentine, is stable between 400°C and 800°C (Bowen and Tuttle, 1949). In Betts Cove - Tilt Cove area talc is formed from orthopyroxene, tremolitic amphibole and serpentine. These can be shown by the following reactions (cf. Deer *et al.*, 1962).

$$\underset{\text{enstatite}}{6MgSiO_3} + 3H_2O \rightarrow \underset{\text{serpentine}}{Mg_3Si_2O_5(OH)_4} + \underset{\text{talc}}{Mg_3Si_4O_{10}(OH)_2} \quad \ldots\ldots \text{(III)}$$

$$\underset{\text{tremolite}}{Ca_2Mg_5Si_8O_{22}(OH)_2} + CO_2 \rightarrow \underset{\text{dolomite}}{2CaMg(CO_3)_2}$$

$$+ \underset{\text{talc}}{Mg_3Si_4O_{10}(OH)_2} + \underset{\text{(removed in solution)}}{4SiO_2} \quad \ldots\ldots \text{(IV)}$$

$$\underset{\text{serpentine}}{2Mg_3Si_2O_5(OH)_4} + 3CO_2 \rightarrow \underset{\text{talc}}{Mg_3Si_4O_{10}(OH)_2} + \underset{\text{magnesite}}{3MgCO_3} + 3H_2O$$

$$\ldots\ldots \text{(V)}$$

Reaction (III) appropriately explains the origin of talc and bastite (serpentine) pseudomorphs after orthopyroxene in Betts Cove. According to Deer *et al.*, (1962), such reaction may take place during cooling cycles when the peridotites are reheated by extraneous solutions. The type of reactions involved in the genesis of the voluminous talc-carbonate-silica rock constituting the relatively

---

*In brucite-bearing massive serpentinites, the MgO/SiO_2 ratio is about 1.23 (Thompson, 1968) whereas the ratio for the brucite-free sheared and monomineralic serpentinites varies from 0.99 to 1.00 (Coleman, 1971b).

thin ultramafic belt between Betts Pig Pond and Tilt Cove, is shown in reactions (IV) and (V). The remnant patches and needles of tremolite within talc are accounted for by reaction (IV); the expelled SiO_2 is perhaps represented by the pods and irregular veins of quartz within talc-carbonate rock. Reaction (IV) shows the chemical adjustments involved in the alteration of serpentine to talc, as seen in a number of serpentine-talc pseudomorphs in Betts Cove samples.

Textural relationships in Betts Cove ultramafic rocks show that the degree of serpentinization is temporally and spatially related to the growth of magnetite. Page (1966) has shown that as a result of recrystallization, magnetite increases during the later stages of serpentinization. The following could represent an unbalanced general reaction (Coleman, 1971b) involving the growth of magnetite in Fe-bearing dunites:

$$\underset{\text{olivine (dunite)}}{(Mg, Fe)_2 SiO_4} + H_2O + O_2 \rightarrow \underset{\text{brucite}}{(Mg, Fe^{+2})(OH)_2} + \underset{\text{chrysotile}}{(Mg, Fe^{+2})_3Si_2O_5(OH)_4}$$

$$+ \underset{\text{lizardite}}{(Mg, Fe^{+3}, Fe^{+2})_3 Si_2O_5 (OH)_4} \pm \underset{\text{magnetite}}{Fe_3O_4} \pm \underset{\text{awaruite}}{FeNi}$$

The sources of water for serpentinization are still being debated. Bowen and Tuttle (1949) showed that ultramafic melts can not contain sufficient water to allow autometamorphism to give rise to serpentinite. Three main sources of water have been suggested: (i) Tectonic invasion of peridotites into wet geosynclinal sediments (Coleman, 1966). (ii) A view once held by Hess (1962) that the serpentinized peridotites within the third layer of the oceanic crust

derived their water from the underlying mantle. (iii) Recent work of Wenner and Taylor (1969) on O^{18}/O^{16} ratios in serpentinites has indicated that, except for antigorite, low-temperature meteoric water may be responsible for serpentinization. Their investigations have also suggested that the serpentinization of oceanic-ridge ultramafic rocks was accomplished by ocean water rather than mantle water as suggested by Hess (1962).

B.2.7 Gabbro and Rodingite within the Ultramafic Member

Thin layers of gabbro occur sporadically within the Ultramafic Member in Betts Cove area. It is most commonly associated with pyroxenite, best seen to the west of Joey's Pond (Fig. 9, Section B), but lenses and layers of gabbro also occur within the serpentinized ultramafic rocks in Pittman Bight.

The gabbro is leucocratic with a range from medium-grained to almost pegmatitic texture. It consists of uralitized clinopyroxene, clinozoisite, chlorite, idocrase, and highly altered plagioclase. The plagioclase in some cases alters to a turbid white isotropic matte possibly hydrogarnet. Garnet pseudomorphs after bytownite in this type of environment have also been reported by Thayer (1966). The clinopyroxene shows partial alteration to a tremolitic amphibole. Clinozoisite occurs as prismatic crystals, generally in veins and patches. Idocrase, not as common as clinozoisite, and chlorite occurs in veins and patches as aggregates of tiny crystals (Fig. 16).

The assemblage of zoisite, clinozoisite, idocrase, tremolitic amphibole (altered from pyroxene), chlorite, and hydrogarnet(?) was noted in samples from two localities in the Betts Cove area (Fig. 2).

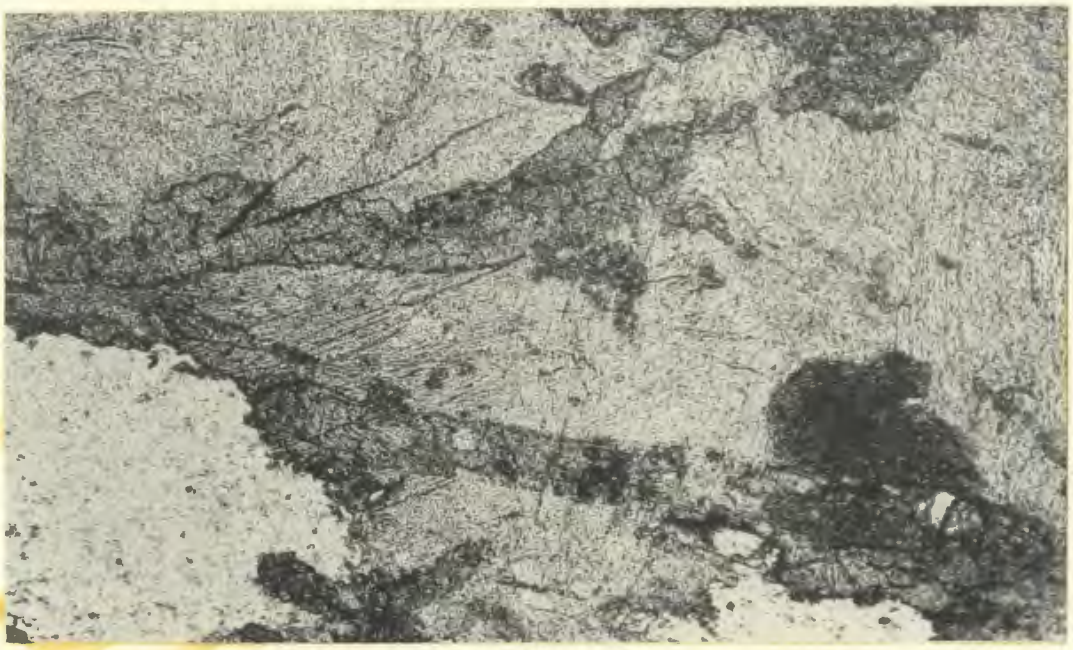


Figure 16: Photomicrograph of rodingite sample showing veins of idocrase cutting through uralitized clinopyroxenes. Sample from Pittman Bight. Plane polarized light (X40).

A rock that contains such an assemblage, invariably associated with serpentinites, has been called "rodingite". It was first identified by Marshall in New Zealand (Bell *et al.*, 1911).

Rodingite consists of a distinctive suite of Ca-rich minerals (Thayer, 1966) namely prehnite, hyrdogarnet,zoisite, clinozoisite, idocrase, diopside, edenite, xonotlite, and chlorite (the principal magnesium mineral). Since its recognition in 1911, rodingite has been subsequently described by others, e.g. Grange (1927), Bilgrami and Howie (1960), and Reinhardt (1969). More recently rodingite has been dredged from the Mid-Atlantic Ridge (Aumento and Loubat, 1971).

Rodingites are generally known to occur as selvages and as border zones of country rocks against serpentinite. Rodingitization of gabbroic and dioritic dykes which have chilled margins against serpentinite, and the development of hornblendic and pyroxenic metamorphic rocks around ultramafic plutons (MacKenzie, 1960; Smith and MacGregor, 1960) show that rodingite is not of magmatic or contact metamorphic origin. Common association of rodingites with serpentinites has led to the belief that rodingitization of country rocks is a low-temperature (500°C) metasomatic process which involves the expulsion of MgO and CaO from and introduction of SiO_2 into the ultramafic rocks that are undergoing serpentinization. In the presence of mafic country rocks Na_2O and K_2O are also lost to the ultramafic rocks (Bilgrami and Howie, 1960). The large-scale expulsion of CaO during serpentinization is also accounted for by the work of Barnes and O'Neil (1969) who found calcium hydroxide waters issuing from partly serpentinized peridotite of the Pacific coast.

B.2.8 Discussion

Ultramafic-mafic complexes are divided into three main types, in terms of origin and field relationships: (i) Stratified or layered intrusions such as Stillwater, Bushveld, Muskox that occur in stable continental platforms. (ii) Alpine types that generally occur in orogenic belts and have complex structural histories. (iii) Zoned ultramafic plutons that are characterized by a dunite-peridotite core surrounded by nearly concentric rims that become progressively less mafic towards the margins. Detailed discussions of these types are provided by Thayer (1960), Wyllie (1967), and McTaggart (1971). The following features suggest that the Betts Cove - Tilt Cove complex belongs to the Alpine-type assemblage (cf. Thayer, 1960, 1967): (i) association of sheeted dykes and pillow lavas, (ii) diabase dykes cutting the ultramfic-gabbroic portion, and (iii) the complex structural relations between Ultramafic and Gabbroic Members (described later).

Size-graded layers and the abundance of cumulate textures in the Betts Cove ultramafic rocks show that gravitational accumulation was a dominant process during its crystallization. Rhythmic layering, such as in Betts Cove, is explained chiefly by the following four processes (cf. Poldervaart and Taubeneck, 1960; Wager and Brown, 1968). (1) Emplacement of periodic pulses of undifferentiated magma. (2) Regular variation in water pressure, especially applicable to granitic and dioritic rocks. (3) Undercooling of magma combined with turbulent or convective currents, perhaps most appropriate to account for the crystals orientated at right angles to layering. (4) Combination of crystal settling and convection currents.

Presence of current bedding and graded bedding in the Betts Cove ultramafic rocks suggest that there was significant movement and flow within the magma chamber and therefore a combination of crystal settling and convection currents was probably the most effective agent to produce the layering. The flow currents might have varied from slow-and-discontinuous to fast-and-sporadic, perhaps giving rise to uniform and size-graded layers respectively (cf. Wager and Brown, 1968).

Geochemical and petrogenetic aspects of the Ultramafic Member are discussed in Chapters IV and V.

B.3 The Transitional Zone and the Gabbroic Member

B.3.1 Field Relationships

The Gabbroic Member has a transitional contact with the underlying Ultramafic Member. In some parts, such as to the east of Burtons Pond, the transitional zone and/or the Gabbroic Member are cut out by a fault; a similar relationship occurs in part of the Tilt Cove area. The combined thickness of the transitional zone and the Gabbroic Member varies from about a maximum of 330 meters (west of Joey's Pond) to complete absence in several places.

The transitional zone is marked by an intimately associated grey clinopyroxenite and gabbro. The two are either interlayered or, less commonly, occur as diffuse pods of one into the other (Fig. 17). 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 meters) in the vicinity of

Figure 17: A typical outcrop of transitional zone between Ultramafic and Gabbroic Members showing interlayered clinopyroxenite (dark) and gabbro. West of Joey's Pond.

Figure 18: Possible ultramafic xenoliths (dark patches) in a gabbro-clinopyroxenite groundmass. The xenoliths show cumulate texture and have relatively sharp contacts with the groundmass material. Hammer in centre for scale. Near southern end of Betts Big Pond.

Kitty Pond than to the west of Joey's Pond. Layering in the transitional zone is parallel to that in the underlying Ultramafic Member.

Isolated, generally lense-shaped, brown-weathering ultramafic pods occur in the transitional zone. They range in size from less than a meter to tens of meters. The contacts 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 preconsolidated 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. Those with cumulate textures, which exhibit sharp contacts with the surrounding granular gabbroic rocks (Fig. 18) may be xenoliths, as cumulate textures are not common in the pyroxenite of the transitional zone. Those parallel to, and lying within, the ultramafic layers may have been produced by the second process. The main difficulty of the first explanation, which interprets some of these pods as xenoliths, is that in most cases the pods occur within the layered gabbro-ultramafic sequence and in the immediate vicinity of this sequence there are practically no rocks that intrude it which could have brought in these pods as xenoliths.

Proceeding upwards from the transitional zone, the proportion of clinopyroxenite gradually decreases and finally the thin Gabbroic Member appears. The top of this member has a gradational contact with the Sheeted Dike Member, marked by huge lenticular inclusions of gabbro within the dykes.

B.3.2 Gabbro

Gabbro ranges from pegmatitic to fine-grained, with an average grain-size of 1.5 to 2.0 millimeters. Milky white plagioclase laths show a distinct contrast against the greenish grey background of uralitized pyroxenes. The gabbro is generally very light coloured due to presence of white plagioclase.

Thin section study shows that the gabbro consists of plagioclase (40-50 percent) and pyroxenes plus amphiboles (40-50 percent) with variable amounts of quartz (in some samples up to 10 percent); pyroxenes and amphiboles are partially altered to chlorite. In two samples minor potash felspar was noticed. Magnetite, pyrite, and carbonates occur as accessories. The proportion of uralitized pyroxene and plagioclase varies even within small areas. The rock shows a hypidiomorphic granular texture. It is classified as a quartz gabbro.

Plagioclase shows two kinds of alteration: (i) A fine, granular, light brown aggregate of low birefringence which is optically indeterminate and, less commonly, (ii) aggregates of tiny sericite flakes which form transparent pseudomorphs. The brownish material is found particularly in those samples that are close to serpentinized ultramafic rocks, suggesting a combined effect of rodingitization and saussuritization. Less altered crystals show a composition in the range of oligoclase-andesine.

Clinopyroxene occurs as euhedral to subhedral crystals that are commonly twinned. It shows partial to complete alteration to actinolitic amphibole and chlorite. Alteration begins from cores or margins of the crystals. The amphibole occurs in prismatic or fibrous forms, some showing polysynthetic twinning, and exhibits a moderate pleochroism from pale green to green. Chlorite is massive, light green and in some cases shows anomalous blue polarization colours. Amphibole and chlorite from sheared zones are wispy-looking and form augen around fresh pyroxene crystals. In highly altered rocks amphibole and chlorite together constitute as much as 40 percent of the rock.

Quartz occurs as large patches or subhedral grains up to 2 centimeters in size. Its boundaries are highly resorbed and corroded where veined and embayed by a turbid saussurite-like material. Quartz and felspar show granophyric texture (Fig. 19).

B.3.3 Gabbro Breccia

Gabbro breccia differs from the gabbro in having rubbly and fractured surfaces. Distinct rock fragments are not evident. It is a monolithologic rock in which non-gabbroic material is extemely rare. Texturally and compositionally it is identical to the gabbro. Its mode of origin is believed to be quite similar to that of the dyke breccia (described later).

B.3.4 Diorite

Near Joey's Pond a thin unit of diorite occurs interlayered with gabbro (Fig. 2). It is fine-grained and has a greyish buff colour. It consists of sodic plagioclase with minor amphibole, epidote, and potash felspar. The rock has a granular texture with average crystal

Figure 19: Photomicrograph of a gabbro sample showing the granophyric intergrowth between quartz and altered felspar. Crossed nicols (X80).

size of 0.5 millimeter. The plagioclase is relatively fresh and has a composition in the oligoclase-andesine range. Patches of diorite also occur within the Gabbroic Member in the Tilt Cove area. Along the Supply Pond - Lond Pond road, for instance, light grey diorite is exposed which probably grades into the surrounding gabbro. It consists of over 90 percent sodic plagioclase and minor potash felspar, chlorite, quartz, epidote, and carbonate.

B.3.5 Discussion

The highly gradational contact between the Ultramafic and the Gabbroic Members, particularly the perfect parallelism in layering, shows that the gabbro was formed by the differentiation of the same magma that gave rise to the ultramafic rocks. The higher quartz content in the gabbro of the Gabbroic Member than in that of the Ultramafic Member suggests that the differentiation was accompanied by a progressive enrichment in silica.

In most ophiolites the ultramafic-gabbro contact is marked by a zone consisting of clinopyroxenite and gabbro. In some of these, e.g. Troodos (Moores and Vine, 1971) and Betts Cove, the gabbro grades downwards into the ultramafic part whereas in others, such as Oman (Reinhardt, 1969) and Liguria (Bezzi and Piccardo, 1971) it intrudes into the ultramafic rocks; in Vourinos (Moores, 1969) both gradational as well as intrusive contacts exist.

B.4 The Sheeted Dyke Member

B.4.1 Field Relationships

This member consists almost entirely of dykes with minor ultramafic and gabbroic screens. The exposed maximum width of the

sheeted complex, measured across the trend of the dykes, is about 4 kilometers. Its thickness within the ophiolite 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 Member are best seen to the south of Kitty Pond and Burtons Pond. These dykes locally constitute mappable sheeted complexes, e.g. at Axe Pond and to the south of Burtons Pond. They have the same composition and thickness as those of the Sheeted Dyke Member above.

In the Burtons Pond - Joey's Pond area, the sheeted complex has a fault contact against the Ultramafic/Gabbroic Members. Elsewhere a transitional contact exists which is marked by a gradually decreasing number of dykes towards gabbro. 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 (Fig. 20). A similar contact zone is also exposed in the Mount Misery area near Betts Cove.

In the Betts Cove - Green Pond area, the sheeted dykes grade into the pillow lava without any change in their orientation. To the north of Foote Pond, on the other hard, the dykes become highly irregular in orientation as they approach the pillow lavas and finally assume the attitude of sills by becoming parallel to the dip and strike of the Pillow Lava Member. The sill-like attitude off these dykes becomes consistent throughout the Pillow Lava Member between Betts Cove and Tilt Cove. Another important change the dykes undergo after entering the pillow lavas is excessive thickening of individual dykes; the average thickness becomes about 2 to 3 meters, against 50 centimeters in the Sheeted Dyke Member.

A fault passing through Kitty Pond and Betts Cove divides the area into two parts that have different trends of dykes. To the north of this fault the dykes have an easterly trend and to the south the trend is nearly east-southeasterly, a discordance of about 30° (Fig. 2). In places the dykes show an abrupt swing of 90° or more in their trend within a distance of 50 meters. One such sharp swing occurs near the eastern outlet of Kitty Pond (see Frontispiece). It is not known whether these swings are primary features or were caused by folding. The following observations suggest a primary (magmatic) origin, although a tectonic origin still remains a possibility.

(i) The swings in dyke trend are not accompanied by swings of comparable "wavelength" in the adjacent members.

(ii) No axial plane cleavage is developed, even in the highly acute swings.

Figure 20: Pillow lavas cut by dykes, marking the transition zone between Sheeted Dyke and Pillow Lava Members. Betts Head. Length of outcrop about 20 meters.

(iii) The dykes have irregular orientations near the contact zone with the overlying pillow lavas and finally become sills in the higher levels. This change in attitude from dyke to sill is undoubtedly a primary feature which may be comparable to the type of swings under discussion.

(iv) Similar swings in the dyke trends of the Troodos ophiolite, on the basis of paleomagnetic study, have been recognized as primary features (Moores and Vine, 1971).

On weathered surfaces the sheeted dykes appear as grey and reddish, nearly vertical stripes (Fig. 21). The weathering colours reflect the composition of the dykes. The diabases, for instance, weathers to medium grey and with increasing pyroxene/amphibole content it becomes reddish. Considerable shearing has taken place along the contacts between adjacent dykes and due to preferential weathering along these contact-shears the gross aspect is that of a strong joint set in a homogeneous rock face (Fig. 22).

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; these margins are marked by a finer-grained texture and a lighter weathering-colour than the central part of the dyke. Some dykes show one-sided chilled margins. Field evidence shows that in such cases a single older dyke is split into two by a younger dyke, so that each of the two counterparts on both sides of the younger dyke show chilling on only one side (Fig. 23). These features are taken as evidence for a tensional environment during

Figure 21: Outcrop of sheeted dykes. Grey dykes are diabase, reddish dykes have low felsic mineral content and approach ultramafic composition. Betts Cove mine trail.

Figure 22: Panoramic view of sheeted dykes. Looking northwest from the beach at Betts Cove. Note the vertical joint-like fracture pattern caused by preferential weathering along sheared contact zones between adjacent dykes.

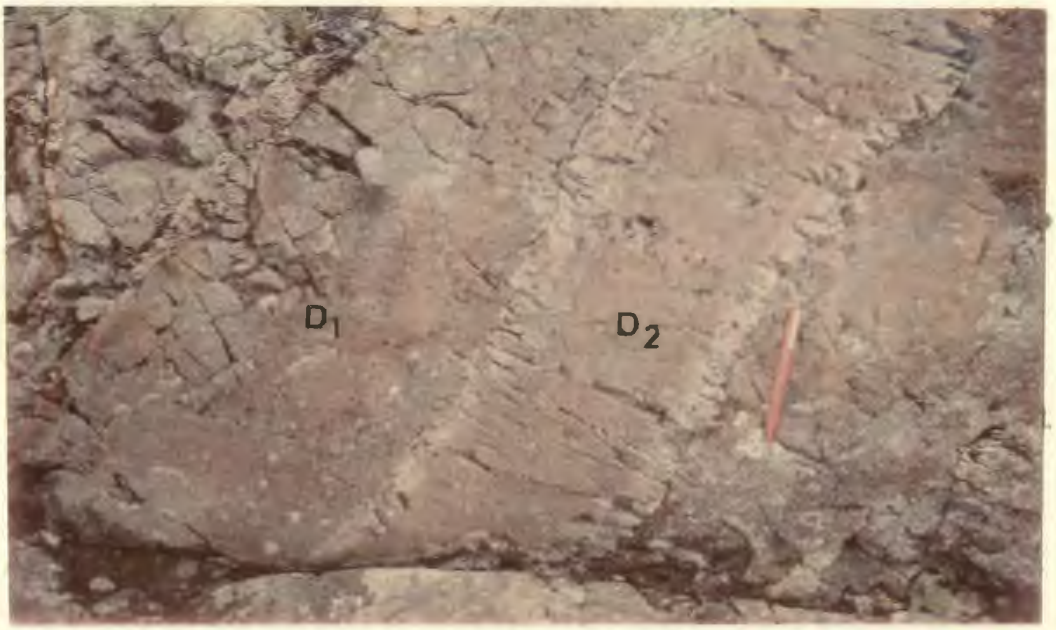


Figure 23: Dyke (D_1) showing one-sided chilled margin. A single dyke was split into two parts (D_1 and the one beneath pen) by a younger dyke (D_2). East of the southern end of Betts Big Pond.

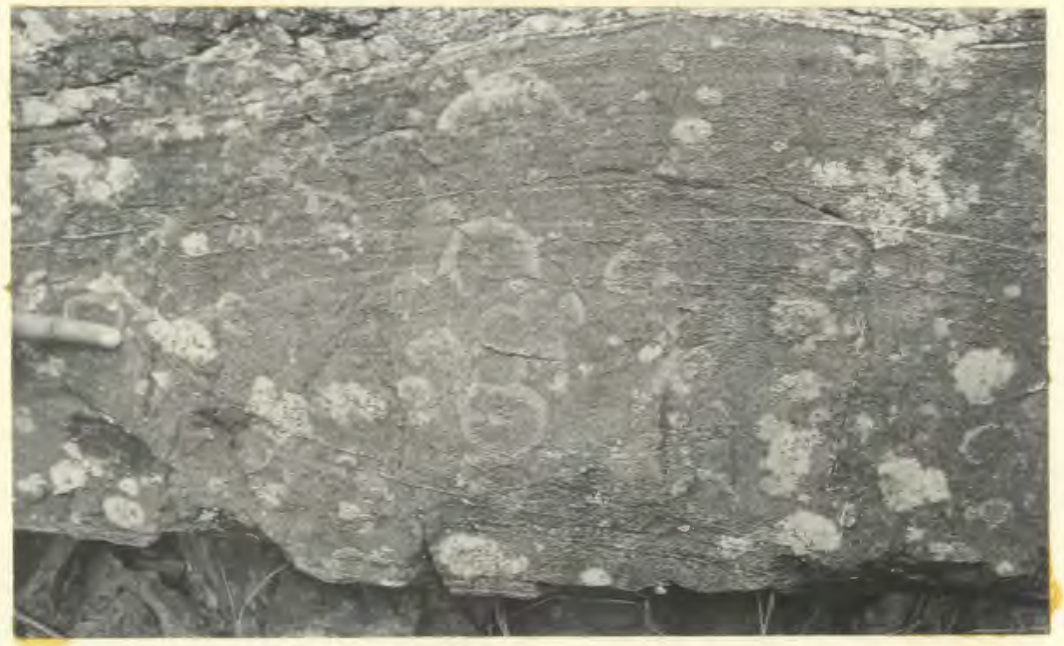


Figure 24: Close-up of a diabase dyke within the sheeted dykes showing fine layers defined by mafic and felsic minerals. Note the fold hinge indicated by the finger.

the emplacement of sheeted dykes.

Very fine layers, defined by mafic and felsic minerals, occur within individual dykes. The layering is parallel to the margins of the dykes and some layers show small-scale folds (Fig. 24). Pyroxene/amphibole crystals are aligned parallel to the margin of the ultramafic dykes. This alignment (foliation) is manifested by closely-spaced planes along which the dyke samples tend to split apart. Dykes containing lesser amounts of pyroxene/amphibole, such as diabase, lack this foliation. The foliation is bent along with the dykes, i.e. it remains parallel to the margins even where the change in trend of the dykes is extremely acute, showing that the alignment of crystals is a primary feature. This is best illustrated by the dyke-swing near the eastern outlet of Kitty Pond.

The dykes are generally, but not necessarily, parallel to one another. Cross-cutting between any two dykes at very low angles and bifurcation of single dykes is observed in places. An extremely fine-grained variety of diabase, 5 to 20 centimeters thick, cuts across the thicker dykes in an irregular zig-zag manner.

A detailed study and documentation of the thickness, composition, and chilled margins of sheeted dykes was conducted along various traverse lines between Betts Cove and east of Kitty Pond. The results are shown in Figure 25, from which the following observations can be made.

(i) The basal part of the sheeted dykes (section A) contains many more gabbroic screens than the upper part.

(ii) As one proceeds stratigraphically upwards, i.e. from Kitty Pond

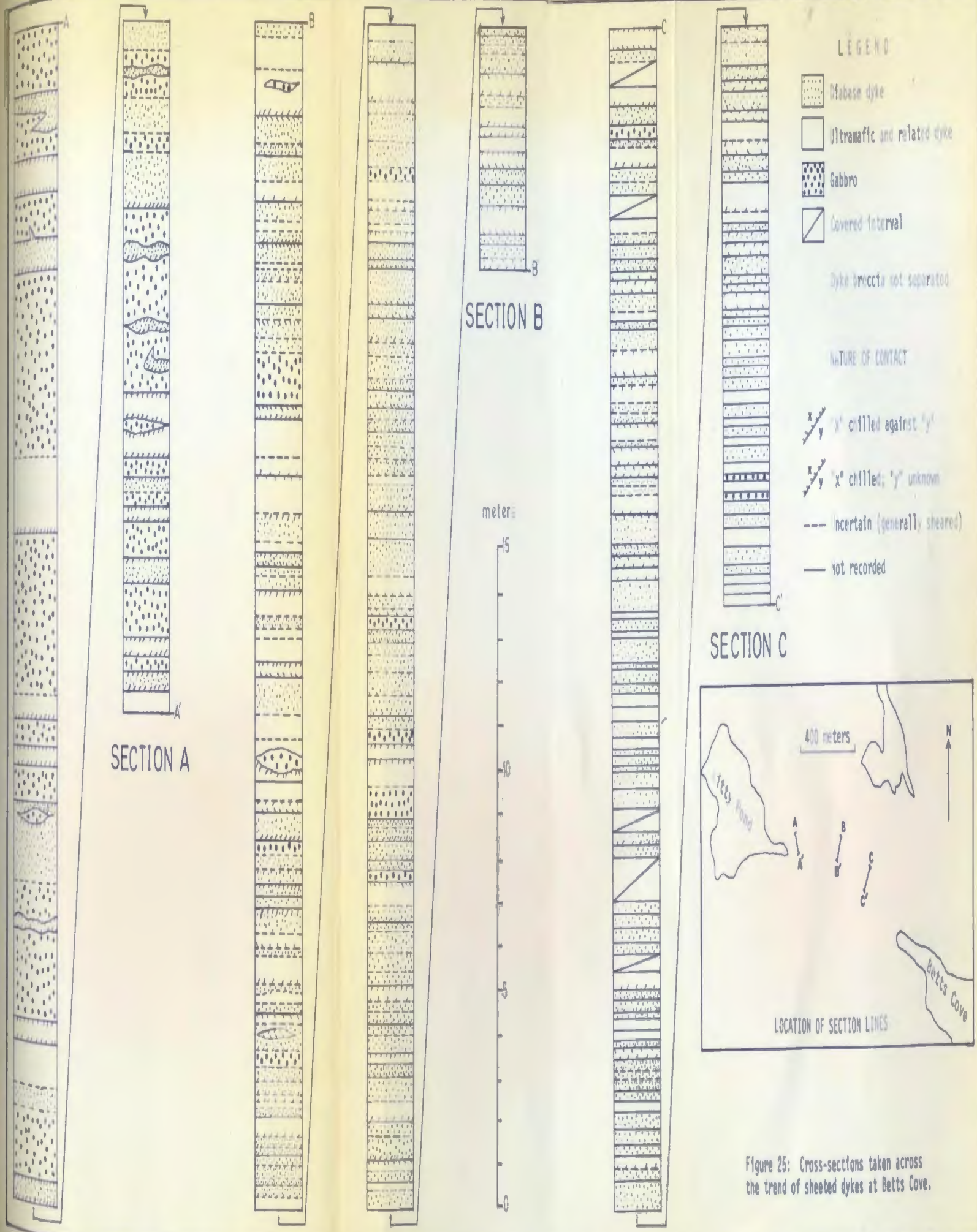


Figure 25: Cross-sections taken across the trend of sheeted dykes at Betts Cove.

towards Betts Cove, the dykes generally get thinner.

(iii) There is no central "line" that divides dykes having predominantly left-sided chilled margins on one side from right-sided chilled margins on the other. Instead, the chilling sides are random, indicating that the younger dykes were intruded along different but nearly parallel planes.

Apart from the exceedingly high proportion of dykes, the Sheeted Dyke Member also includes minor amounts of dyke breccia and screens of ultramafite, gabbro, and granodiorite. A more detailed account of these appears in the following section.

B.4.2 Diabase Dykes

Diabase dykes constitute the majority of the sheeted complex. They weather to overall greys or reddish-greys in which tiny laths of white plagioclase are discernible. Fresh surfaces are generally greenish-grey. The rock varies in grain size from almost aphanitic to the one with crystals up to 1 millimeter across. The chilled margins appear as dark grey zones, 1 to 3 centimeters wide, in which no textural or mineralogical details can be observed.

The diabase dykes consist of pyroxene/amphibole (up to 70 percent), plagioclase (up to 25 percent), quartz (about 5-10 percent) with subordinate epidote, carbonate, chlorite, magnetite, and chromite. The rock shows an imperfectly developed subophitic-hypidiomorphic texture. Quartz and some plagioclase occur as interstitial material. The amphibole is secondary after clinopyroxene, although remanants of the latter are rarely seen. The amphibole, as suggested by optical characters, belongs to the tremolite-actinolite series. Long and

tapering crystals of amphibole in places protrude across the grain boundaries of plagioclase and quartz. Plagioclase (oligoclase-andesine) is altered to patches of epidote and carbonate. Epidote constitutes up to 5 percent of the rock.

B.4.3 Ultramafic and Related Dykes

Dykes belonging to this category constitute about 25 percent of the area traversed and documented in detail by the author (Fig. 25). A reconnaissance survey of other parts of the Sheeted Dyke Member suggests that this ratio of ultramafic and related dykes may be representative of the whole member. These dykes weather to light brick-red surfaces; less altered varieties have greenish colours. The crystals occur as tiny needles up to 2 millimeters in length. Cumulate textures are absent. Alignment of crystals parallel to the dyke walls imparts a foliation which contrasts with the adjacent, generally massive diabase dykes. These dykes also contain minor amounts of felsic minerals. On the basis of felsic material present, they can be classified as two types: (i) Perknite and (ii) felsic perknite which are referred to as "ultramafic dykes" and "related dykes" respectively. Perknite, consisting mainly of pyroxene and amphibole (Peterson, 1960), includes those that contain up to five percent felsic material whereas in the felsic perknite this material ranges from 5 to about 15 percent. Chlorite and chromite occur as accessories in both of them. Most of the perknite samples studied by the author have over 95 percent pyroxene/amphibole.

The amphibole, which constitutes about 75 percent of the perknite samples, was formed by the alteration of clinopyroxene;

remnants of the latter occur in the cores of some amphibole crystals (Fig. 26). Pyroxene grains can be easily distinguished from the surrounding amphiboles by their higher refractive indices. The amphibole is of tremolite-actinolite series and occurs as laths and fibers. Some of these show porphyritic texture in which pyroxene phenocrysts are altered to serpentine in the core and tremolitic amphibole on the margin (Fig. 27). The felsic material is chiefly quartz with subordinate felspar. It generally occurs as aggregates of several grains that form circular or elliptical areas up to 3 millimeters across. These areas, resistant to weathering, project as tiny blobs in handspecimen and closely resemble the variolitic textures that are common in the pillow lavas in the Betts Cove area. Quartz and felspar also fill the interspaces between pyroxene/amphibole crystals. Similar rocks having the composition of felsic perknite have also been reported from the Precambrian Onverwacht Group in South Africa (Viljoen and Viljoen, 1969a).

B.4.4 Silicic Dykes

Silicic dykes consist mainly of albite aplite. They constitute a quantitatively negligible part of the sheeted complex and were observed only at a few places in the Betts Cove area. In nearly all cases the aplite grades into grey-weathering diabase, within a few meters along the length of the dyke. The aplite is a medium-grained, creamy white weathering rock that shows saccharoidal texture in handspecimen. Sporadic patches of mafic minerals are sometimes present.

The aplite consists of quartz and sodic plagioclase with minor potash felspar, clinopyroxene, amphibole, epidote, and carbonates.

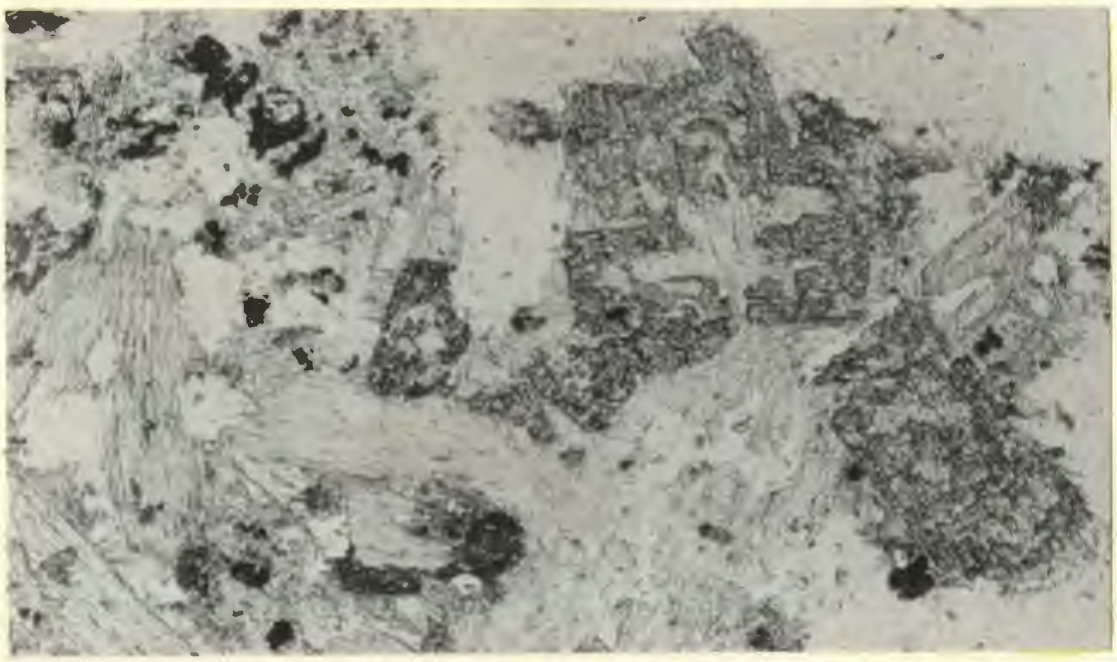


Figure 26: Photomicrograph of an ultramafic (perknite) dyke showing remnant clinopyroxene grains (dark grey) that are altered to the surrounding uralitic amphibole. Plane polarized light (X60).

Figure 27: Photomicrograph of a porphyritic ultramafic (perknite) dyke in which pyroxene phenocrysts are altered to serpentine in the cores and to amphibole on the margins. The groundmass consists of partly uralitized pyroxene crystals. The dyke cuts layered ultramafic rocks to the south of Kitty Pond. Crossed nicols (X40).

It has a hypidiomorphic granular texture. The average grain-size is 1 millimeter, although at the gradational contact with the diabase it becomes somewhat coarser (Fig. 28). Such quartz grains have quasi-graphic intergrowth with saussuritized plagioclase.

B.4.5. Dyke Breccia

The dyke breccia consists of angular to subrounded fragments of either diabase, gabbro, or ultramafic rocks; the fragments have the same lithologies as those constituting the sheeted dykes. The breccia occurs in irregular patches and veins that have no planar structures. It lies both along and across the trend of the dykes, the former generally in the contact zone between adjacent dykes. It is present at every level of the Sheeted Dyke Member. In Tilt Cove area the dyke breccia forms practically all of the Sheeted Dyke Member.

The breccia has a rubbly appearance (Fig. 29). The diameter of the fragments ranges from 1 millimeter to over 30 centimeters. The most typical size is in the order of 1 to 5 centimeters. Sorting is very poor. The fragments are set in a finer but compositionally similar matrix. The breccia commonly includes a variety of rock fragments but monolithologic breccia also occurs. The latter consists of a single rock type and superficially conveys the appearance of a highly fractured outcrop. The fragments are generally equidimensional with varying degrees of roundness, both suggesting abrasion during their formation. The existence of well-developed dyke breccia in the regions away from shear/fault zones and the sharp contacts between nonbrecciated dykes and the dyke breccia indicate that the breccia could not have been formed by tectonic deformation.

Figure 28: Photomicrograph showing the contact between diabase (lower left) and the coarse phase of an aplite dyke. In the outcrop the two grade into each other along strike. Near the outlet of Kitty Pond. Crossed nicols (X40).

Figure 29: Rubbly-looking diabase dyke breccia intertonguing with relatively smooth, finer ultramafic dyke breccia (overlain by hammer). Near southern shore of Winser Lake, Tilt Cove.

In thin sections, the fragments show well-rounded corners. The contact between fragments and the matrix, at least in some cases, is somewhat gradational, even though the two have contrasting mineralogy and texture. This is interpreted as due to abrasion and welding of the fragments in a "hot" magmatic environment.

Field and laboratory evidence indicates that the dyke breccia is genetically related to the associated dykes. The matrix of typical breccia is similar to that of fragmented igneous rocks developed through fluidization or gas action (cf. Reynolds, 1954), and it is likely that the dyke breccia of Betts Cove - Tilt Cove region was formed in this way. The volatiles escaping under a high pressure-gradient (Szekely, 1971) could have caused the brecciation of the dyke material. Dyke breccia has also been described from the Bay of Islands ophiolite in west Newfoundland by Williams and Malpas(1972) who give a similar explanation of its origin, and it appears that it may be a common association of ophiolite suites, although hitherto not widely recognized.

B.4.6 Screens within Sheeted Dykes

(a) Field Relationships.

The screens differ from the enclosing sheeted dykes in two respects: (i) lack of chilled margins and (ii) generally lenticular shape. Any dyke that occurs adjacent to a screen invariably shows a chilled margin against the screen, whereas the screen rocks have a consistently uniform grain size from centre towards margins, without any suggestion of chilling (Fig. 30). The screens are thickest in the centre and they taper and finally disappear along the trend of the

Figure 30: Lenticular gabbro screens (G) enclosed by diabase dykes (light coloured). Note that the grain-size of gabbro remains unchanged from core to margins whereas dykes (e.g. the one underlain by penknife) have chilled margins. Betts Bight.

dykes. The longest dimension of these screens varies from a few centimeters to as much as 150 meters. Detailed work in Betts Cove (Fig. 25, Section A) shows that the screens generally become smaller and less abundant as one proceeds from the base towards the top of the Sheeted Dyke Member. Exceptions to this, however, are fairly common, e.g. the occurrence of huge gabbro units near the contact with the pillow lavas around Betts Cove mine. Individual screens may or may not be cut by the dykes. In Figure 2 only the large mappable screens containing less than 30 percent dykes are shown. The rock types that form the screens are gabbro, ultamafites, and granodiorite.

(b) Gabbro Screens.

Gabbro screens are most abundant. They consist of a leucocratic medium-grained rock identical to the quartz gabbro of the Gabbroic Member. This is distinguished from the diabase of associated dykes by its coarser texture, lighter colour and, most important, by the absence of chilled margins. Layering in the gabbro screens is scarce. Pods of grey pyroxenite are commonly associated with the gabbroic screen rocks.

(c) Ultramafic Screens.

These are generally reddish weathering and best studied in the coastal exposures, such as Nippers Harbour Islands and Bluff Point-Green Point areas. These screen rocks are heavily serpentinized, and cumulate textures are not recognizable. Some ultramafic screens are also formed by pyroxene-amphibole rock (perknite) that locally grade into gabbro.

(d) Granodiorite Screens.

A granodiorite screen, over 30 meters in length, is exposed to

the east of Burtons Pond (Fig. 2). It is a medium-grained, light grey rock in which quartz grains stand out distinctly against the weathered ground mass. It is associated with and perhaps grades into gabbro. The chief constituents of the granodiorite are quartz and felspar, both in nearly equal amounts. Chlorite, epidote, and carbonate occur as accessories. Some of the quartz grains show idiomorphic crystal outlines. Felspars include plagioclase and lesser orthoclase. Most of the epidote and carbonate have resulted from the alteration of felspars.

B.4.7 Discussion

Apart from Betts Cove - Tilt Cove area, sheeted dykes are also known from Troodos, Cyprus (Gass, 1968; Moores and Vine, 1971), Oman (Reinhardt, 1969), Kizil Dagh Massif, Turkey (Vuagnat and Cogulu, 1968), Macquarie Island (Varne *et al.*, 1969), central Newfoundland (Strong, 1972), Mings Bight, northeast Newfoundland (R. Norman, personal communication, 1972), and Bay of Islands (Williams and Malpas, 1972). The most spectacular feature of these dykes is the near-absence of country rock within them. In the places mentioned, sheeted dykes form an integral part of ophiolite suites and where known, always occupy a particular stratigraphic position within them. In the light of the information available from modern ocean-floors, ophiolite suites are now interpreted as ancient oceanic crust and mantle (Dietz, 1963; Hess, 1964; Gass, 1968; Moores and Vine, 1971; Dewey and Bird, 1971; Upadhyay *et al.*, 1971). The ultramafites, gabbro plus sheeted dykes, and pillow lavas of the ophiolites are considered to be equivalent to the upper mantle, layer-3, and layer-2 respectively of the oceanic crust.

Generation of the oceanic crust through ocean-floor spreading (Vine, 1966) demands a tensional environment at the site of the spreading (mid-oceanic ridges). The sheeted dykes, particularly those with one-sided chilled margins (Fig. 23), in the ophiolites are taken as evidence to indicate a strongly tensional environment prevalent during their formation. Any new dyke within the complex, in order to make room for itself, would tend to push the walls away from one another thus resulting in the lateral movement (spreading) of the whole ophiolite suite (oceanic crust). Sheeted dykes, therefore, seem to be the best physical criteria for the recognition of ophiolite suites that were generated through spreading at the mid-oceanic ridges.

All the ophiolites however do not contain sheeted dykes. The origin of such dykes perhaps depends on the rate of spreading. Moores and Vine (1971) have suggested that the ophiolites without sheeted dykes, such as the Italian types, were formed at ridges that had a fast rate of spreading. In these the material injected in a narrow axial ridge would move without further addition of material and would crystallise as it moved. The upper part of the magma would crystallise immediately and display extrusive features (pillow lavas), and the part immediately under-lying it would crystallise more slowly and show gradation downward into the coarser-grained intrusive material (gabbro).

The screens within the sheeted dykes, as stated earlier, lack chilled margins and always occur as discontinuous, generally lenticular units in Betts Cove. This, in conjunction with the younger age of dykes that cut the Ultramafic-Gabbroic Members, shows that the screens were picked up by the dykes as consolidated or partly-consolidated

"cold" rocks from the Ultramafic-Gabbroic Members. This interpretation is also substantiated by the petrological similarity between the gabbro samples of the screens and those of the Gabbroic Member.

It is important to note that the leucocratic, medium-grained quartz gabbro that occurs as screens in the sheeted dykes, is not found in the overlying Pillow Lava Member, further emphasizing that the screens could not have been intrusive into the sheeted dykes, otherwise they should also be found in the pillow lavas. Further support for this view comes from the observation that the screens do not run oblique to the trend of the sheeted dykes, as would be expected if they were intrusions.

The author proposes that the chief factor to force the "cold" solid screens upwards along with the "hot" molten dykes was a high pressure-gradient, such as that suggested by Szekely (1971). As the dykes reached the surface (i.e. the dykes became flows) the gradient would cease to exist and consequently the screens would not be carried into the pillowed volcanic rocks.

B.5 The Pillow Lava Member

B.5.1 Field Relationships

The Pillow Lava Member stretches from Betts Cove to Tilt Cove and, in these two areas, it is underlain by the sheeted complex. The nature of contact between the two has been discussed in earlier parts of this thesis. The Pillow Lava Member has a maximum thickness of 1500 meters in the Fly Pond (Betts Cove) area. It shows a slight-thinning towards east and is abruptly reduced to a thickness of merely 40 meters between Long Pond and Tilt Cove (Fig. 2). Farther east, in

the vicinity of Tilt Cove mines, it again attains a thickness of about 300 meters.

The Pillow Lava Member includes a sedimentary unit about 60 meters thick, consisting chiefly of thinly bedded argillite. This unit, very consistent in thickness, can be traced from Fly Pond in the west to the western side of Long Pond, in the east. The dip of the Pillow Lava Member, as indicated by this sedimentary unit and some bun-shaped pillows, gradually decreases at stratigraphically higher levels. This member comprises the following lithologies: (i) pillow breccia, (ii) sill complex, and (iii) pillow lava.

B.5.2 Pillow Lava.

The pillows are close-packed, generally bun-shaped or ellipsoidal, and moderately flattened. Fresh exposures are greenish grey. They weather to rusty brown or reddish grey surfaces, the former being rich in mafic minerals. The pillows vary in size from about 30 to 150 centimeters. Tiny pillows, as small as 5 centimeters across, occur between larger pillows in some places. Dark grey glassy chilled margins up to 2 centimeters wide, are common. The outer zones of the pillows are marked by a concentration of varioles, up to 1 centimeter across, that tend to stand out in relief. The interspaces among pillows are generally filled with chert. Both mafic and ultramafic pillow lavas occur.

(a) Mafic Pillows.

Over 95 percent of pillow lavas in the Betts Cove ophiolite are of mafic composition. In handspecimen these lavas differ from the ultramafic lavas in their lighter grey or reddish grey weathering

colours.

Microscopic study shows that the pillows consist chiefly of amphibole, chlorite, plagioclase, and variable amounts of clinopyroxene, with minor epidote, quartz, calcite, and opaques. The mafic pillows can be divided into three textural and mineralogical types: (1) Those consisting of plagioclase and pyroxene in nearly equal proportions (about 40 percent each). The pyroxenes are stout and the plagioclase crystals are lath-shaped. Some samples show porphyritic texture with pyroxene and plagioclase as phenocrysts. (2) Those showing an intergrowth of long, slender crystals of plagioclase, and associated clinopyroxene. The plagioclase crystals are needle-shaped and in places bent. Pyroxenes also have similar shapes and are elongated parallel to the length of the plagioclase crystals. Such intergrown crystals may radiate away from common centres, giving rise to partial or complete rosette-like features (Fig. 31). The plagioclase crystals show partial to complete alteration to an epidote-rich semi-opaque turbid material. (3) Those types in which prismatic amphiboles occur in a matrix of quartz, sodic plagioclase, and chlorite. The amphiboles commonly constitute over 60 percent of the rock, and pillow lavas with as much as 80 percent amphibole have been observed, the latter mark the transition towards ultramafic pillows.

The mafic pillows are highly altered. Clinopyroxenes are altered to an amphibole of the tremolite-actinolite series. The latter, in turn, has altered to chlorite. Rocks consisting solely of chlorite and saussuritized plagioclase also occur. Chlorite varies from almost isotropic to a variety with deep blue and brown interference colours.

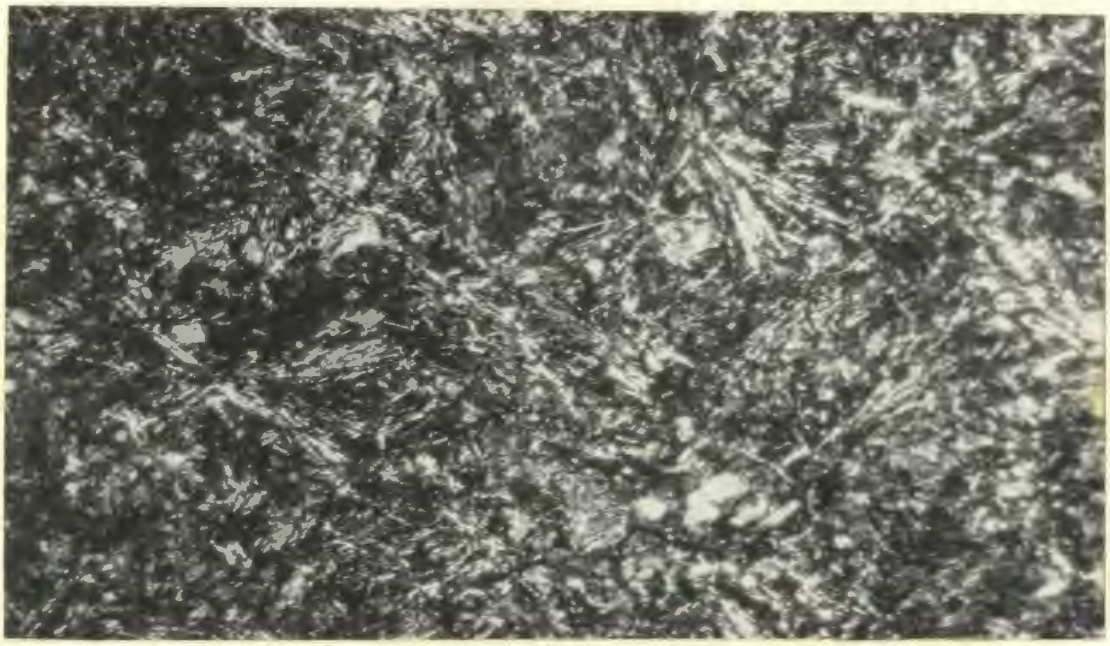


Figure 31: Photomicrograph of Betts Cove pillow lava showing the rosette-like features caused by the radial arrangement of intergrown plagioclase and secondary amphibole crystals. Crossed nicols (X20).

Figure 32: Photomicrograph of a pillow lava sample from near Betts Cove mine showing randomly oriented clinozoisite laths in a plagioclase-uralite groundmass. Plane polarized light (X15).

Chloritization generally begins either from the margins of, or the fractures within, the pyroxene/amphibole crystals. In some cases massive pale green chlorite is associated with patches of quartz.

Reconnaissance measurements of extinction angles (Michel-Lévy method) in plagioclase crystals indicate a composition in the oligoclase-andesine range. In porphyritic samples, plagioclase phenocrysts are commonly embayed by the chlorite of the groundmass. Epidote occurs in cavities and veins, commonly associated with massive chlorite. It constitutes up to 10 percent of the rock. Zoisite and clinozoisite are absent in some and abundant in others. In one sample from the Betts Cove mine area, randomly oriented slender crystals of clinozoisite produce a curious network pattern (Fig. 32); aggregates of several clinozoisite crystals form shell-shaped growth patterns. Quartz fills the interstices between plagioclase and pyroxene/amphibole crystals. It also occurs in cavities and veins.

Chromite occurs as reddish brown grains in the amphibole-rich (type 3) pillows. Magnetite is scarce and, where present, occurs as euhedral grains or as tiny granules. On the north shore of Betts Cove some of the maroon pillows contain up to 15 percent magnetite. Native iron has been reported from the Betts Cove pillow lavas by Deutsch and Rao (1972).

The varioles in the pillows contain a high concentration of epidote in the centre and needle-shaped plagioclase on the margins. The latter are nearly perpendicular to the periphery of the varioles and form a semi-radiating pattern.

(b) Ultramafic and Related Pillows.

These pillow lavas occur as isolated pods in close association with the mafic lavas (Upadhyay, 1973). The main occurrence, on which the present study is based, lies in the lenticular pillow lava outcrop at Partridge Pond. Here the ultramafic pillows form the upper parts of thick ultramafic (perknitic) sills. At Partridge Pond 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 (type 3) mafic pillows. Along strike the ultramafic pillow unit also shows a gradation into the massive reddish brown ultramafic sill.

The degree of the development of pillow structures is variable. A deep-brown weathering ultramafic rock at Partridge Pond, for instance, shows ellipsoidal structures that strongly resemble pillows of the volcanic rocks of the area. But no definite chilled margins or cuspate "V" shaped bottoms are identifiable in them. Extensive shearing and fracturing adds to the difficulty of establishing any pillow structures. They occur in close spatial association with more definite ultramafic pillows. They consist of serpentine, talc, and carbonates. Some of the less altered samples show remnants of pyroxene grains. Suggestions of serpentine pseudomorphs after olivine also exist.

On the other hand, a buff-weathering variety shows definite pillow structures that have ovoid shapes and faint indications of chilled margins. The presence of "V" shaped cuspate bottoms, formed by the curved edges of adjacent pillows, is perhaps the best evidence to describe them as pillow structures. Differential weathering of the

pillow lava outcrops generally produces a knobbly surface in which round pillows stand out in relief. Where seen in cross-sections the central parts of the pillows are more deeply eroded than the margins. 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. This gradation corresponds to the similar variation from perknite to felsic perknite dykes described from the Sheeted Dyke Member. These very dykes after coming to the surface have become flows and produced the pillow lavas of mafic and ultramafic compositions.

The chemical composition of some of the ultramafic and related pillows is given in Table 2 and discussed in Chapter IV. The creamy white ultramafic pillows have remarkably low Al_2O_3 content. The ultramafic pillows differ from the mafic types in their consistently higher MgO and lower Na_2O contents. The silica content in the ultramafic pillows is over 45 percent, showing that they are not "ultrabasic" (Wyllie, 1967). It should be noted that one clinopyroxenite (sample 3, Table 2) from the Ultramafic-Gabbroic transition zone has lower MgO, and higher Al_2O_3 and SiO_2 contents than the ultramafic pillows, implying that the pillows are closer to the ultramafic/ultrabasic clan than is the clinopyroxenite.

Ultramafic lavas are also known from Turkey (Bailey and McCallien, 1953), Troodos massif, Cyprus (Gass, 1958; Searle and Vokes, 1969), Barberton Mountain Land, South Africa (Viljoen and Viljoen, 1969a, 1969b), Munro Township, Ontario (Pyke *et al.*, 1973), and the Rambler mine area, Newfoundland (Gale, 1973). The Betts Cove occurrence shows similarity with that of Troodos and Barberton Mountain in being closely associated with mafic pillows. The gradational contact between lavas and ultramafic dykes/sills and the development of isolated pillow structures within the sills is remarkably similar to the relationships reported from Barberton Mountain (Viljoen and Viljoen, 1969a). Unlike Troodos, layering within individual pillows does not occur in Betts Cove nor do the spinifex textures reported from Ontario and South Africa.

Mineralogically, the Betts Cove ultramafic pillows are most similar to those of Barberton Mountain in which tremolite, chlorite, and antigorite are the chief constituents, although the last two are not common in the creamy white pillows in Betts Cove. The mineralogy of the mafic volcanic rocks ("komatiite" of Viljoen and Viljoen, 1969a) in Barberton Mountain compares closely with the type-3 mafic pillows associated with the ultramafic pillows at Partridge Pond in the present area.

B.5.3 Pillow Breccia

Pillow breccia occurs near the base of the Pillow Lava Member. In Betts Cove it was mapped in the Mount Misery - Foote Pond area, in the vicinity of Betts Cove mine, and to the east of Betts Bight Pond (Fig. 2). The breccia is very irregular in development and varies

from a few meters to over 50 meters in thickness. It consists of angular pillow fragments set in a fine-grained greyish yellow matrix. The latter in places includes a high content of red chert. The pillow fragments show good chilled margins and variolitic textures. The breccia is greyish brown on weathered and pale greyish green on fresh surfaces. Where the pillow fragments are small the breccia has a rubbly appearance.

Pillow breccia is a more important constituent of the Pillow Lava Member in the Tilt Cove area than it is in the Betts Cove area. Excellent exposures of greenish pillow breccia occur on the cliff face along the road joining the Government wharf and the northern end of Winser Lake (Fig. 33). Here the breccia is highly sheared and has a rusty appearance due to associated sulfide mineralization.

In thin sections the broken pillows within the breccia are altered to chlorite and saussurite. The breccia matrix consists of tiny epidote anhedra and an extremely fine-grained turbid material.

B.5.4 Sill Complex

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. In the basal part of this member they are generally a couple of meters thick. In the Red Cliff Pond - West Pond area, on the other hand, sills as thick as 200 meters have been mapped (Fig. 2). These sills are medium-grained in the central part

Figure 33: Pillow breccia showing broken pillows in a brecciated mafic matrix. Near Tilt Cove wharf.

and fine-grained on the margins.

The best examples of the sill complex occur about 600 meters north of Fly Pond and also in a lenticular pillow lava outcrop at Partridge Pond (Fig. 2). The latter occurs as a patch within the Ultramafic Member and is tentatively interpreted* as a downfaulted part of the main Pillow Lava Member. Here about ten sheeted sills, without any intervening pillow lavas are exposed on a cliff face. These sills vary in thickness from 0.5 meter to about 5 meters. The uppermost sill grades upwards into mafic and ultramafic pillows. Similar gradational contacts between sills and pillows are also seen in the East Pond area. Isolated, complete as well as partially developed pillow structures can be seen within the sills. These are ellipsoidal in shape and are defined by concentrations of white-weathering varioles, features that commonly occur along the margins of the pillows in the Snooks Arm Group. This, coupled with the remarkable compositional similarity between the sills and the associated pillows, shows that the sills acted as feeders to the pillow lavas.

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. Perknite and highly sheared serpentinite sills occur on the northwestern shore of West Pond and eastern shore of Red Cliff Pond respectively (Fig. 2). Porphyritic as well as ophitic textures are common in the diabasic sills.

---

*The internal setting, dips, and the compositions of the pillow lava/sill assemblages are quite identical at both places.

A diorite sill, about 5 meters thick, occurs within the sheeted sill complex at Partridge Pond. It is a grey, medium-grained rock with distinct chilled margins. The constituents are about 75 percent sodic plagioclase and 25 percent brown hornblende. The latter is of primary origin, unlike the common amphibole in the Betts Cove ophiolite which is secondary after pyroxene.

B.5.5 Discussion

Pillow lavas form the uppermost part of all ophiolite suites. Presence of pillows shows that the lava erupted under water. The pillow lava portion of the ophiolites therefore simply represents the extrusive equivalent of the underlying sheeted dykes, the latter having acted as feeders. Petrologically, therefore, the sheeted dyke/pillow lava boundary is insignificant. This is clearly shown in the Betts Cove area by the petrochemical similarity between the two and also by some of the partially developed pillow structures within the sills/dykes.

The concentration of pillow breccia at the base of the Pillow Lava Member in Betts Cove indicates that the brecciation was caused by escaping gaseous material (fluidization) as the dykes reached the surface. This is in contrast to pillow breccias which develop from slump, or flow of pillows and matrix material during extrusion. The latter type occurs on top of the pillow lava pile and is overlain by well bedded tuffs, as in Quadra Island, British Columbia (Carlisle, 1961).

B.6 Felsitic Material in the Betts Cove Ophiolite

A felsitic rock, occurring at the base of the Pillow Lava Member, is exposed about 500 meters east of the Betts Cove mine (Fig. 2).

It has a near-vertical attitude, with a thickness of about 3 meters and can be traced for a distance of about 80 meters. It is terminated at both ends by faults. A similar rock occurs within the sheeted dykes along the Betts Cove mine trail. It is also about 3 meters thick and can only be traced for about 30 meters. At both localities, this rock strikes oblique to the trend of the nearest sheeted dykes.

Contacts with the adjacent rocks are sheared. In places this felsite shows thin layers that are roughly parallel to the contacts with adjacent rocks. It is a creamy white aphanitic rock, in which only a few quartz crystals are discernible with naked eye. Microscopic study shows that it consists of an extremely fine aggregate of quartz, felspars, and muscovite, with quartz and alkali felspars as minute phenocrysts.

As this felsitic rock is very limited in extent and shows no evidence of independent "roots", it is interpreted as a differentiated silicic phase within the dominantly mafic rocks of the Betts Cove Ophiolite.

C. BOBBY COVE AND BALSAM BUD COVE FORMATIONS

C.1 Field Relationships

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 meters 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 meters. 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. An account of these rocks follows.

C.2 Andesitic Agglomerate, Tuff, and Flysch

These rocks constitute a sizeable part of Bobby Cove Formation but only a minor part of Balsam Bud Cove Formation. Probably the best and most accessible exposure lies along the East Pond - Snooks Arm trail.

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 grey andesitic rock with pyroxene, and less commonly amphibole phenocrysts up to 1.5 centimeters across (Fig. 34). Some of these also contain vesicles and carbonate-bearing amygdules that are moderately elongated. The ellipsoidal fragments have their longer axes aligned in a common direction, generally parallel to the bedding of the adjacent sediments. Bedding within the agglomerate is generally either lacking or poorly developed, however, there are scattered thin units of well bedded graywacke and argillite. At Buttonhole Cove, the agglomerate contains large lumps of porphyritic andesite, gabbro, foliated amphibolite, pyroxenite, and hornblendite.

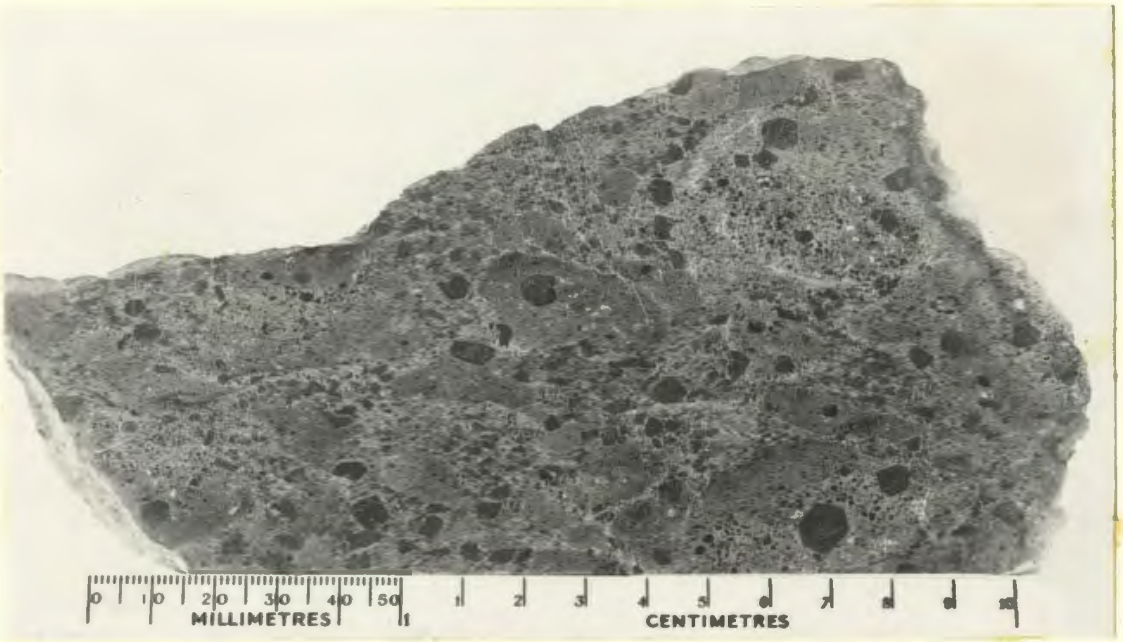


Figure 34: Handspecimen of andesitic agglomerate showing poorly-defined ellipsoidal units of andesite that contain pyroxene phenocrysts (dark spots). Tiny round blobs are amygdules. Sample from near the outlet of East Pond.

The ellipsoidal fragments consist of plagioclase and large clinopyroxene phenocrysts set in an epidote-rich semi-opaque brownish groundmass. The plagioclase phenocrysts have an average size of 1 millimeter and show a crude trachytic texture. Their composition lies in the oligoclase-andesine range. Pyroxene crystals are partly altered to amphibole and chlorite. Some of these crystals show zoning.

Andesitic tuff is perhaps the most common pyroclastic rock in the Snooks Arm Group. Along the East Pond - Snooks Arm trail the agglomerate grades into the tuff, accompanied by a gradual decrease in the size of the associated pyroxene phenocrysts. It contains trains of deeply weathered calcareous material that are parallel to the bedding of the nearby sediments. Bedding in the tuff, although not perfect, is better developed than in the agglomerate. The degree of sorting is highly variable. Some of the poorly sorted tuffs contain wisps, flames, and lenses of foliated amphibolite. They also possess isolated elliptical patches of andesite in which pyroxene phenocrysts are coarse at the centre and fine outwards, suggesting a faster rate of cooling at the margins.

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. It consists of andesite fragments that possess zoned plagioclase phenocrysts; tiny pyroxene and amphibole crystals occur as phenocrysts within the andesitic fragments and also as separate clasts. Silicic material is scarce or absent. In places it shows current-bedding and crude graded bedding.

The abundance of shattered debris, relative scarcity of vesicular material, and a crude size-gradation in the pyroclastic rocks of the Snooks Arm Group shows that these deposits are of subacqueous origin (cf. Parsons, 1967). From the foregoing it is noted that these pyroclastic rocks are mostly quartz-free. Similar rocks are reported from the early Miocene of the New Hebrides by Mitchell (1970) who interprets them as island-arc deposits and suggests that similar ancient arcs could be indicated by quartz-free turbidites containing abundant calc-alkaline (andesitic) detritus.

C.3 Graywacke

Graywacke is generally associated with argillite. Thickness of the graywacke units varies from a few centimeters to several meters. Size-graded beds are common. The average grain-size is about 1 to 2 millimeters although much larger clasts are relatively common. The grains are generally angular to subangular. Lithologically, the graywacke may be classified into the following three types.

C.3.1 Andesitic Graywacke

It consists of over 60 percent andesitic clasts with subordinate chlorite, carbonates, epidote, and plagioclase grains. Some of the andesitic fragments show good trachytic texture. These fragments occur in a chloritic matrix. The rock contains sparse or no quartz clasts.

C.3.2 Felspathic Graywacke

This is a light-coloured dominantly quartz-feldspar rock. Rock fragments, present only in minor amounts, are chiefly of andesitic composition. Feldspar includes both plagioclase and orthoclase. Quartz grains show undulatory extinction.

C.3.3 Lithic Graywacke

This is the most common type of graywacke. It contains roughly similar amounts of felspathic material and rock fragments. Carbonate and chlorite are very common. The lithic fragments are chiefly of andesitic material with minor clasts of granitic and granodioritic rocks. Wispy units of some nearly opaque material are present in almost every lithic graywacke. This material has one cleavage with the suggestion of a second strain-slip fabric. These clasts may have been derived from the nearby polyphase deformed Fleur de Lys terrain. Fleur de Lys metamorphosed and deformed clasts in the Snooks Arm graywacke have been reported by Church (1969).

C.4 Chert and Argillite

Chert and argillite invariably occur as thinly bedded units. In some parts such beds are only a few centimeters thick and, being of different colours, they produce a ribbon pattern. Cherty argillite and argillaceous chert also occur. The chert is red, maroon, light green, grey or buff. In thin sections, aggregates of extremely small, optically indeterminate turbid material contain scattered opaques and chalcedony grains.

The argillite has the same colours as the chert. Being soft and fine-grained, it shows the best-developed cleavage of all rocks of the Snooks Arm Group. The light coloured argillite consists of quartz, sericite, mica and some chlorite and epidote. The greenish argillite is rich in actinolitic amphibole. Red and darker varieties contain a high proportion of extremely fine-grained opaque material, presumably hematite and magnetite. In the argillite samples flattening of quartz

and felspar along the cleavage is very common.

C.5 Conglomerate and Breccia

These 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 over 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 (Fig. 35). The silicic clasts in both of them show a remarkable similarity to the rhyolitic rocks of the Cape St. John Group exposed in the Red Cliff Pond - Brents Cove Area. The conglomerates also contain fragments of quartz, felspar, and andesite. The matrix is rich in carbonate and chlorite.

C.6 Sedimentary Structures

Pre-consolidation sedimentary structures are moderately plentiful in the Bobby Cove and Balsam Bud Cove Formations. They are exposed at a number of places, especially near the terminal of the Snooks Arm road and at a few places in Wild Bight. They can be best studied on partially weathered surfaces. Relatively simple lenticular as well as highly contorted thin units of graywacke occur in a groundmass of argillite or finer-grained graywacke (Fig. 36). The contorted and irregular units are up to 30 centimeters in length and a couple of

Figure 35: Breccia of Balsam Bud Cove Formation showing angular clasts of rhyolite (white) that strongly resemble rhyolite of the nearby Cape St. John Group. North shore of Snooks Arm.

Figure 36: Slump folds in the sediments of Bobby Cove Formation. The coarse material is graywacke and the fine is argillite. Near the terminal of Snooks Arm road.

centimeters in thickness. The contorted as well as the associated lenticular units are confined to certain evenly bedded units. Large contorted blocks of graywacke with good bedding are found randomly oriented in a groundmass of similar or different lithology. The bedding in these blocks is also contorted and it closely follows the contorted margins of the block, indicating that the unconsolidated blocks were detached and slumped while sedimentation was in progress. The complexly folded thin units within the uniformly dipping strata suggest that these structures were developed during sedimentation. Slumping of coarse-grained units under the action of gravity seems to account for the development of these structures.

Where overlain by graywacke, convolute bedding is common in the argillite. The downfacing convolute surface of the graywacke is mostly asymmetric and irregular. This is interpreted as due to pre-consolidation adjustment within different sedimentary units. Small-scale cross-bedding occurs in the argillite exposed along the northern shore of Venams Bights and the northeastern shore of Wild Bight. The individual cross-bedded units range from a few millimeters to about 12 millimeters. Graded bedding is common in the graywacke. Flame structures, in which "flames" of argillite protrude into the overlying graywacke, occur at several places.

C.7 Pillow Lava

Thin flows of basaltic pillow lava occur at several levels within the Bobby Cove and the Balsam Bud Cove Formations. They vary in thickness from a few meters to tens of meters. In the Tilt Cove area, for instance, a 300 meter thick pillow lava unit extends from

Scrape Point and gradually tapers towards Venams Bight. Most of these flows seem to have elongated lenticular shapes.

The pillows are medium to dark grey in colour. At Buttonhole Cove a hematite(?)-rich maroon porphyritic pillow lava flow occurs. The lava consists of saussuritized plagioclase, some pyroxene, and a large amount of chlorite and carbonate. Potash felspar, epidote, and opaques occur as accessories.

C.8 Sill Complex

The Bobby Cove and Balsam Bud Cove Formations include a high proportion of diabase sills. In some places, for example along the Long Pond - Venams Bight trail, the sills constitute over 50 percent of the stratigraphic column. Generally the sills are concordant with the associated sediments. The thickness of individual sills varies from a few meters to as much as 200 meters. An exceptionally thick sill is exposed on the western shore of Beaver Cove. This sill also contains small screens of pillow lava. All the sills are coarse- to medium- grained at the centre and gradually become finer towards the margin.

The sills are diabasic in composition. They consist of plagioclase, clinopyroxene, secondary chlorite, and carbonate with minor quartz and potash felspar. Epidote and opaques always occur as accessories. The rocks show ophitic, subophitic, and porphyritic textures.

C.9 Discussion

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 sills, like those in the Pillow Lava Member of the Betts Cove Ophiolite, may have had the attitude of dykes before intruding the Bobby Cove and Balsam Bud Cove Formations.

The strong similarity between the rhyolitic clasts in the conglomerate/breccia and the rhyolites of the nearby Cape St. John Group suggests a northwesterly "continental" source of the sediments. On the other hand, the graywacke and conglomerate/breccia also contain andesitic clasts of island-arc affinity. This detritus could have been derived from the island-arc type andesitic flysch of the Beaver Cove Group or, less certainly, from the Pacquet Harbour - Rambler terrain of the Fleur de Lys Supergroup in the northwest. However, more definitive island-arc type assemblages have been reported from the Moretons Harbour (Strong and Payne, 1973) and Long Island (Kean, 1973) areas in the southeast. These assemblages, either equivalent to, or somewhat younger than, the Snooks Arm Group, may be underlain by a similar but older island-arc rocks which acted as a provenance for part of the Snooks Arm sediments. This postulated island-arc may now by lying beneath the waters of Green Bay. The subject is discussed in more detail in Chapter VI.

D. VENAMS BIGHT AND ROUND HARBOUR BASALTS

D.1 Field Relationships

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 shows the dip of the associated pillow lavas. Excellent pillow structures in the Round Harbour Basalt are exposed around Snooks Head (Fig. 37). They occur in a variety of shapes and sizes and single pillows up to 4 meters long occur.

D.2 Venams Bight Basalt

These pillow lavas, on fresh surfaces, are generally medium-grey and slightly darker than those of the Betts Cove Pillow Lava Member.

As seen in thin sections, they consist of plagioclase, clinopyroxene, amphibole, and chlorite with minor epidote, carbonate and opaques. They commonly show subophitic texture. Trachytic texture, although not very common, occurs in some specimens. Phenocrysts, mainly sodic plagioclase and less commonly clinopyroxene, are present in nearly all specimens and make 2 to 10 percent of the rock. Size of the phenocrysts varies from less than a millimeter to over 6 millimeters. The groundmass consists of chlorite, a fine-grained turbid material, and tiny granules of magnetite. The turbid material, optically indeterminate, renders the rock semi-opaque in thin sections.

Figure 37: Pillow lavas of the Round Harbour Basalt. Note the relatively shallow dip of the pillows, due to their location at the trough of a syncline. Snooks Head.

Plagioclase and its pseudomorphs, constitute 40 to 60 percent of the rock. It has the composition of sodic andesine (An_{35}) and is altered to sericite, chlorite, epidote, and carbonate. It occurs as slender crystals commonly with two-lamellae twins. Clinopyroxene content ranges from 25 to 40 percent. Euhedral crystals are seen only where plagioclase laths are not associated with it. Some clinopyroxene crystals show undulatory extinction. Partial alteration to tremolitic amphibole and chlorite is common.

Quartz is an accessory, occurring as subhedral grains and as aggregates filling the spaces between pyroxene and plagioclase crystals. Quartz of secondary origin occurs in veins and cavities. Chlorite, an alteration product of pyroxene, forms a variable proportion of the rock. Samples contain up to 60 percent chlorite, occur in veins and patches or as perfect pseudomorphs after pyroxene. It is massive, light green, and nearly isotropic. Epidote forms aggregates of tiny clear crystals closely associated with chlorite and quartz. Calcite either forms pseudomorphs after plagioclase or occurs as irregular diffuse patches.

Mineralogy and textures suggest that these pillows are basalts or andesitic basalts. The low An content of the plagioclase was probably caused by albitization. Chemical analyses (Chapter IV) show that these are tholeiites with high total alkalis and high Na/K ratio, both of which may reflect Na metasomatism.

D.3 Round Harbour Basalt

The pillow lavas of this formation are dark grey to nearly black, a feature that distinguishes them from other volcanic rocks of the

Snooks Arm Group.

These lavas consist of plagioclase, clinopyroxene, chlorite, minor quartz and variable but minor amounts of epidote, magnetite, and carbonate. Pillows of this formation differ from those of the other two in being less altered, containing plagioclase of more calcic composition, and in containing higher amounts of magnetite. Porphyritic, subophitic, and trachytic textures are common (Fig. 38). Pillows with conspicuous plagioclase phenocrysts are exposed at Snooks Head.

Plagioclase (An_{48} - An_{58}) content varies from 30 to 60 percent. The crystals are thin and elongate and commonly show two-lamellae twins. Plagioclase phenocrysts are completely altered to a turbid aggregate of epidote, sericite, chlorite, carbonate, etc. Some of the less altered phenocrysts have a composition in the range of An_{10} - An_{20}. In one sample zoned plagioclase was observed. Zoning is also indicated in others by concentric bands of fresh and saussuritized plagioclase. Clinopyroxene is 30 to 60 percent of the rock. Some crystals show twinning, zoning, and hourglass-like structure. It is altered to amphibole and chlorite in variable degrees. Magnetite constitutes as much as 20 percent of some of the samples and is responsible for the dark colour of the pillows. It occurs as euhedral crystals, tiny granules or stringers.

Mineralogy, texture, and chemical composition (Chapter IV) suggest that these pillow lavas are tholeiitic basalts.

D.4 Sill Complex

Sills probably form 25 to 40 percent of the Venams Bight and Round Harbour Basalts. In places they are over 60 percent and locally

Figure 38: Photomicrograph of a typical subophitic texture in the pillow lavas of Round Harbour Basalt. Light grains are clinopyroxene, grey laths are plagioclase. Crossed nicols (X40).

Figure 39: Photomicrograph of a diabase sill showing two vesicles, one with semi-radiating needles of pumpellyite. Sample from Snooks Head. Crossed nicols (X15).

give rise to a "sheeted" sill complex. Good examples of the latter occur at Venams Bight and to a lesser extent at Snooks Head. At Venams Bight, earlier sills are split apart by the younger ones. In some cases thin screens of pillow lava also occur along the contacts between adjoining sills. Thin chert laminae occur parallel to some sills. The thickness of individual sills varies from about a meter to over 12 meters but most of them are in the 5 to 7 meter range. The sills are concordant with the pillowed flows, and in some places alternate with them. They are medium-grained in the centre and become finer and finally glassy towards the margins. The margins of the sills are shaped by the cusps between adjacent pillows, producing wavy rather than straight borders. This is taken as one of the evidences for the sills being the same age as the pillows and having acted as feeders to them.

Sills in both formations are basaltic in composition. They consist of sodic plagioclase, clinopyroxene, and chlorite. Quartz, epidote, carbonate, and opaques occur as accessories. They show subophitic texture, quite similar to that in the associated pillow lava. Plagioclase is moderately to completely saussuritized. Some of the plagioclase crystals are zoned and some show undulatory extinction. Clinopyroxene is partly altered to amphibole and chlorite. Chlorite also occurs in the cavities/vesicles. One sample from Snooks Head contains light green radiating needles of pumpellyite within the chlorite-bearing vesicles (Fig. 39).

Some sills in the Round Harbour Basalt show striking mineralogical and textural similarity to those in the underlying Balsam Bud Cove

Formation and Venams Bight Basalt. It is therefore probable that some of these sills/dykes also intruded the underlying formations while "feeding" the Round Harbour Basalt on top.

The sills in some places, e.g. Venams Bight, contain autobreccia that occurs as veins and pods. The breccia consists of angular fragments of diabase with some chert and carbonates.

D.5 Pillow Breccia

In the Venams Bight Basalt, pillow breccia occurs at Venams Bight and in Round Harbour road cuts. In Venams Bight the breccia is about 10 meters thick and contains complete as well as partial pillows in a fragmented chert-rich matrix. In the Round Harbour road cuts the breccia comprises pillow fragments in a jasper-rich matrix. The matrix also includes delicately laminated red mudstone.

D.6 Discussion

In the Round Harbour Basalt, especially in Snooks Head area, the sills and pillow lava units show a crude alternation which, in conjunction with the petrographical similarity between the two, suggests that each sill probably produced a thin pillow lava unit that lies on its top. This implies that the generation of both of these pillow lava formations took place in tens, perhaps hundreds, of small intrusive-extrusive phases that are represented by the individual sills/pillow lava pairs. The sills probably started as dykes (vertical) at depth and became sills (subhorizontal) after entering the open surface at the bottom of the basin. The upper portion of these sills was a lava with pillowy structures.

The significance and genetic aspects of the Venams Bight,

Balsam Bud Cove, and Round Harbour Formations, that conformably overlie the Betts Cove Ophiolite, but are not a part of the ophiolites, are discussed in Chapter V.

E. LATE INTRUSIVE ROCKS IN THE SNOOKS ARM GROUP

The Snooks Arm Group is intruded by sills and dykes that are predominantly silicic in composition. These are most abundant in the Tilt Cove area. Here they occur as plugs and irregular sills and dykes. Their width ranges from less than a meter to over 10 meters. Some of these, e.g. a dyke exposed on the cliff face at Tilt Cove wharf, cut across the geological contacts. Such dykes exhibit faint mineral banding parallel to their walls and show distinct chilled margins. Local shearing of these intrusions and of the country rocks has developed strong foliation. Some of these foliated silicic intrusive rocks have been erroneously interpreted as "rafts" of the pre-Ordovician Cape St. John Group within the Snooks Arm rocks by Dewey and Bird (1971).

An extensive granitic sill runs along the contact between the Bobby Cove Formation and the underlying Pillow Lava Member of the Betts Cove Ophiolite. Silicic intrusive and extrusive rocks also occur as lenticular units within the Balsam Bud Cove Formation and Round Harbour Basalt.

Nearly all these silicic sills and dykes consist chiefly of quartz plagioclase and potash felspar with minor chlorite and micas. Porphyritic texture is common with quartz and felspars forming the phenocrysts. Felspars are completely sericitized. Quartz grains are embayed by sericitic material.

From the lack of any intermediate (dioritic) material in the vicinity of these dykes and sills and from their highly irregular trends, it seems unlikely that these could represent differentiated phases of the dominantly mafic rocks of the Snooks Arm Group. Hence they are tentatively interpreted as post-Snooks Arm Group (Siluro-Devonian?) in age.

F. STRUCTURE, DEFORMATION, AND METAMORPHISM

F.1. Folds

A cross-section along the Red Cliff Pond - Snooks Head join shows that the beds are overturned or subvertical to the north-west and gradually assume shallower dips towards the southeast (Fig. 2). A somewhat similar variation takes place along the Tilt Cove - Venams Bight - Round Harbour coastal section. In these areas the strike of the beds varies from easterly to northeasterly. In the southwestern part of the Snooks Arm Group, i.e. along Kitty Pond - Bobby Cove - Wild Bight join, two significant differences exist: (i) The dip of the Pillow Lava Member in the Betts Cove area is much shallower than in the West Pond area. (ii) The strikes vary from northerly in the north to southeasterly and even easterly near Bobby Cove. This is best seen along the Betts Island - Buttonhole Cove - Bobby Cove coastal section. In this section flexures and tight folds are quite common whereas in the northeast such structures are rare or absent.

Folds are best developed in Bobby Cove. They have an easterly plunge of about 40° and a wavelength of about 500 meters. Minor folds also occur along the eastern shore of Wild Bight, in Venams Bight, and

along cuts in the Round Harbour road. At all these places the plunge of the folds is consistently to the east.

A slaty cleavage which commonly dips to the southeast at a somewhat steeper angle than the bedding, indicates a synclinal axis to the southeast. Detailed documentation of bedding and cleavage orientations shows that the cleavage consistently strikes to the east despite sharp swings in the bedding. This suggests that the minor folds are auxiliary to the main syncline and that the dominant (axial plane) slaty cleavage in the Snooks Arm Group was developed through a single deformation.

From the foregoing account it is concluded that the Snooks Arm Group forms the northern limb of a slightly overturned, east-plunging syncline as first suggested by Neale (1957). The strike of beds and flows in the Betts Island - Bobby Cove - Wild Bight region shows that the axis of such a syncline passes through this join. The nose of the syncline would lie in the vicinity of Betts Bight.

It has been suggested (Neale and Kennedy, 1967; Marten, 1971) that the southern limb of the syncline is formed by the Western Arm Group exposed along the southern shore of Green Bay. This group consists of a pillow lava-tuff sequence quite similar to that of the Snooks Arm Group, and the strata have a moderate to steep dip to the north. The following two features of the Western Arm Group (Marten, 1971, Fig. 2) strengthen the suggestion that these two groups form opposite limbs of a syncline: (i) In the western part of the Western Arm area the bedding trends swing from east-northeast to west, indicating the probable position of the nose of the syncline. This

syncline, according to Marten, has a northeasterly plunge of 35°.
(ii) The penetrative cleavage consistently trends east-northeast despite changes in the trend of the bedding; this attitude of the (axial plane) cleavage is quite similar to that in the Snooks Arm Group.

The Snooks Arm and the Western Arm Groups are separated from each other by the Green Bay Fault, now covered by the waters of the Bay. Regional correlations show that it is a strike-slip dextral fault with a horizontal displacement of about 25 kilometers. Similar dextral wrench-faults, perhaps related to the Green Bay Fault, are described from the Western Arm area by Marten (1971) who attributes the rotation of bedding and cleavage in the Harrys Harbour area to one such fault.

F.2 Faults

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 the total. The most prominent one extends from Pittman Bight to Beaver Cove and marks the contact between the Snooks Arm Group and the adjacent Fleur de Lys 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 Opiolite 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 a crude parallelism to each other, indicating some possible genetic relationship. Such a relationship has been observed and more completely established in the Western Arm Group by Marten (1971).

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 mylonization.

F.3 Deformation

The Snooks Arm Group possesses one slaty cleavage indicating that it has been subjected to a single phase of deformation. Since no younger strata lie on top of this group, direct evidence for the age of deformation does not exist. The only deformed rocks of post-Snooks Arm age nearby, are those of the Siluro-Devonian Mic Mac Sequence. This Sequence consists of basaltic and rhyolitic flows and subaerial sediments and, like the Snooks Arm Group, has one main

cleavage (Neale and Kennedy, 1967). This similarity in structural history suggests that the Snooks Arm Group was deformed at the same time as the Mic Mac Sequence and therefore a post-Silurian age of deformation is postulated. Since the most pervasive deformation in the Newfoundland Appalachians has been during the Acadian Orogenic event (Williams *et al.*, 1972) it is probable that the deformation of the Snooks Arm Group is also related to this pre-Carboniferous, poorly dated Mid Paleozoic event.

F.4 Metamorphism

The Snooks Arm Group has been subjected to two types of metamorphism: burial and regional. The burial metamorphism is more pervasive at greater depths. Thus, the youngest formation (Round Harbour Basalt) contains pumpellyite (Fig. 39) indicating zeolite facies metamorphism operative under low temperatures and intermediate pressures (cf. Winkler, 1967). The alteration of pyroxene and plagioclase also increases with depth; e.g. in the Round Harbour Basalt they are only slightly altered whereas in the underlying formations, particularly the Betts Cove Ophiolite, they are almost completely uralitized and/or saussuritized. The bastite-talc alteration, such as in the Betts Cove ultramafic rocks, is stated to be produced under temperatures and pressures equivalent to those of upper greenschist and lower amphibolite facies metamorphism (O'Hara, 1967), further suggesting that the metamorphism increased with depth in the Snooks Arm Group.

Regional metamorphism was of greenschist facies, as indicated by the commonly occurring quartz-chlorite-sodic plagioclase-epidote assemblage. Growth of chlorite along slaty cleavage shows that this

metamorphism was related to the deformation that produced the cleavage.

The burial metamorphism is interpreted as an early event, prior to the emplacement of the Snooks Arm Group. Regional metamorphism was a later event, presumably related to Acadian (Mid-Upper Devonian?) orogeny.

CHAPTER IV

GEOCHEMISTRY

A. GENERAL STATEMENT

The results of 53 major-element analyses of igneous rock samples of the Snooks Arm Group are presented in Table 2. The main purpose of these analyses was to determine whether the mafic pillow lavas of this group are calc-alkaline or tholeiitic. Ultramafic and related pillow lava samples were analysed specifically to determine their differences from the mafic lavas. Analyses of the plutonic ultramafic rocks, gabbro, and sheeted dykes are very few and represent only an analytical reconnaissance. All samples of the ophiolite were taken from the Betts Cove area as the rocks there are fresher and the suite is better developed than in the Tilt Cove area. Figure 40 shows the location of the analysed samples.

B. THE BETTS COVE OPHIOLITE

B.1 Ultramafic, Gabbroic, and Sheeted Dyke Members

Three samples of unpillowed ultramafic rocks were analysed from the Betts Cove Ophiolite (samples 1, 2, 3; Table 2). Sample 1 is a serpentinized lherzolite from the layered sequence. Sample 2 is a serpentinite occuring at the base of the pillow lava outcrop at Partridge Pond. This sample, perhaps originally a pyroxene-dunite, shows a distinctly higher MgO content than the lherzolite. Alkalis in both of them are either below detection levels or occur only as traces. The third ultramafic rock (sample 3) is a greenish grey,

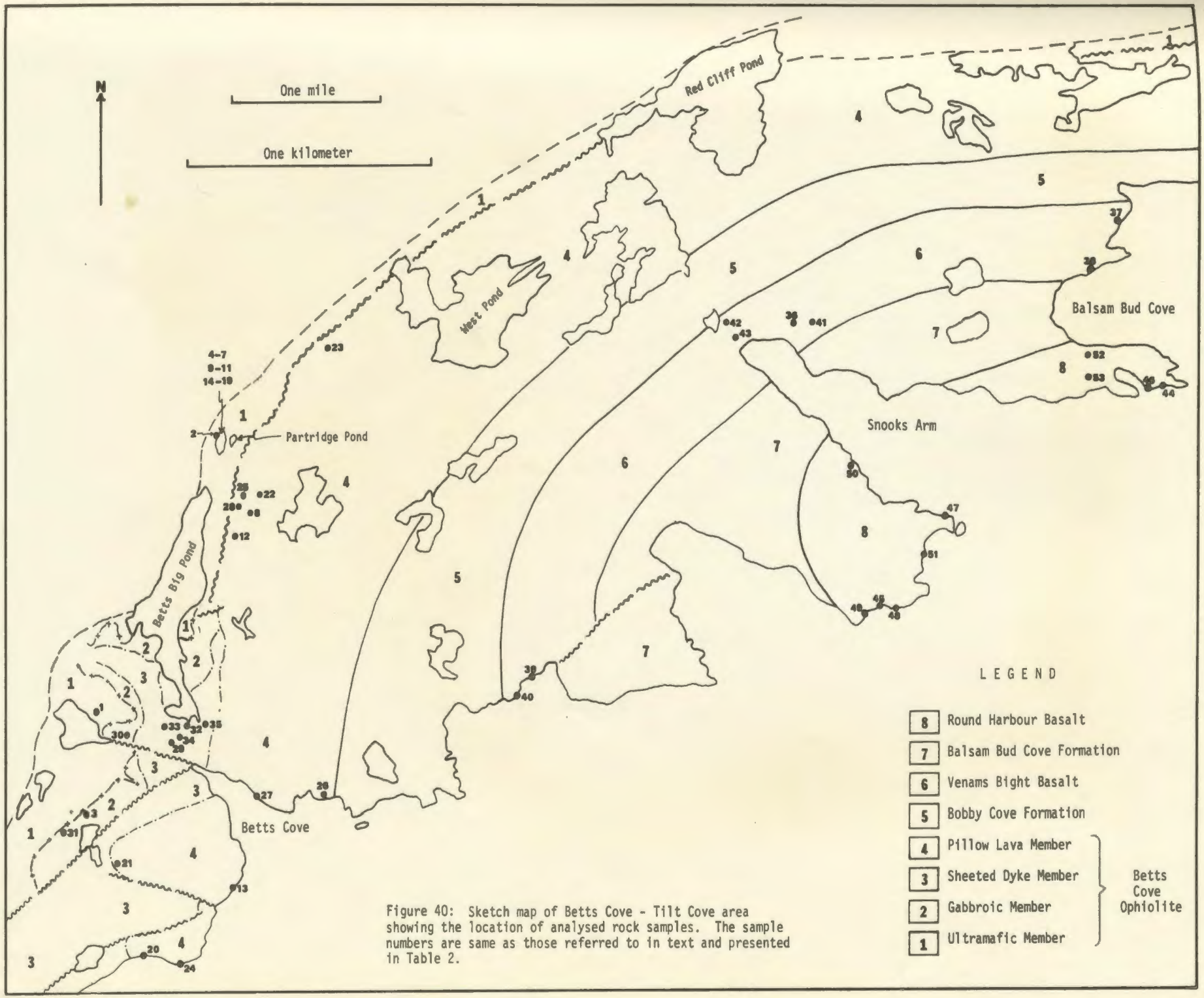


Figure 40: Sketch map of Betts Cove - Tilt Cove area showing the location of analysed rock samples. The sample numbers are same as those referred to in text and presented in Table 2.

TABLE 2

Chemical analyses of igneous rocks of the Snooks Arm Group*

	UM 1	UM 2	UM 3	UMP 4	UMP 5	UMP 6	UP 7	UP 8	UP 9	UP 10	UP 11	UP 12	UP 13	UP 14	UP 15	UP 16	UP 17	UP 18
SiO_2	46.10	46.30	50.30	35.20	44.90	32.60	46.20	44.50	56.50	58.50	50.30	48.60	46.40	51.00	47.70	54.90	48.70	46.20
TiO_2	ND	ND	<0.10	ND	ND	ND	<0.10	<0.20	ND	ND	0.10	<0.10	<0.10	<0.20	<0.20	<0.20	<0.20	<0.20
Al_2O_3	1.00	<0.50	6.10	<1.00	1.30	<1.00	9.10	10.60	0.70	0.30	9.20	8.40	11.50	10.20	10.40	11.20	10.20	9.20
Fe_2O_3	2.31	1.87	1.11	0.83	1.60	1.05	1.03	2.31	0.75	0.51	0.73	1.08	1.43	1.67	1.57	2.16	1.50	1.72
FeO	6.42	4.29	5.28	5.70	5.44	5.33	9.30	10.34	4.93	4.94	9.49	7.82	7.62	6.70	8.02	5.75	7.11	8.17
MnO	0.16	0.11	0.14	0.10	0.07	0.09	0.17	0.23	0.09	0.08	0.18	0.16	0.21	0.16	0.18	0.14	0.16	0.17
MgO	30.38	37.19	17.08	26.68	26.46	26.89	19.91	18.49	21.84	22.09	18.32	17.03	15.36	16.40	15.88	13.43	17.81	20.43
CaO	2.90	T	12.81	13.95	9.98	14.24	8.21	7.78	11.75	13.40	7.80	9.26	8.80	8.66	7.54	5.13	7.02	8.13
Na_2O	T	<0.05	0.62	0.05	0.07	0.05	0.16	0.30	T	T	1.01	0.07	0.13	0.96	0.40	3.43	0.15	0.17
K_2O	T	T	0.70	T	T	T	0.38	0.16	T	T	0.15	0.40	1.67	1.37	0.65	0.08	0.72	0.34
P_2O_5	ND	ND	ND	0.22	ND	ND	ND	T	ND	T	ND	0.03	ND	ND	ND	T	T	T
L.I.	8.50	11.90	2.60	16.90	10.00	17.90	4.60	5.80	2.50	2.20	4.40	6.00	5.50	3.70	5.40	4.30	5.90	5.50
Total	97.77	101.66	96.74	99.63	99.82	98.19	99.06	100.51	99.06	101.93	101.68	97.05	98.62	100.82	97.74	100.52	99.27	100.03
CIPW Norms																		
Q	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	1.86	0.08	--
Or	--	--	4.39	--	--	--	2.37	0.99	--	--	0.91	2.59	10.60	8.34	4.16	0.49	4.56	2.12
Ab	--	0.18	5.57	--	0.65	--	1.43	2.68	--	--	8.78	0.65	1.18	8.36	3.66	30.16	1.35	1.52
An	3.05	--	12.52	1.37	3.60	1.41	24.32	28.61	1.35	0.56	20.69	23.51	27.75	20.05	26.70	15.51	26.80	24.68
Aug	10.30	--	43.66	23.13	40.64	10.74	15.04	9.75	33.67	37.79	15.33	21.37	15.31	19.30	11.07	8.72	8.37	14.34
Hy	62.37	64.01	27.28	--	24.43	--	35.91	33.79	1.72	--	42.68	46.43	30.08	33.99	50.65	39.99	56.48	37.01
Ol	20.51	32.49	4.84	57.44	28.07	63.18	19.23	20.62	62.48	60.94	10.30	3.52	12.73	7.45	1.27	--	--	17.66
Mt	3.75	3.01	1.71	1.44	2.58	1.88	1.58	3.53	0.77	0.51	1.08	1.71	2.22	2.49	2.46	3.25	2.32	2.63
Ilm	--	--	--	--	--	--	0.10	--	--	--	0.19	0.10	0.10	--	--	--	--	--
Ap	--	--	--	0.61	--	--	--	--	--	--	--	0.07	--	--	--	--	--	--

ND Not determined
T Trace
L.I. "Loss on ignition"

UM Ultramafic rock
UMP Brown-weathering ultramafic rock with pillow-like shape (from Partridge Pond)
UP Ultramafic and related pillow lava (perknite and felsic perknite)

* Analyst: Mrs. G. Andrews, using atomic absorption spectroscope (Perkin-Elmer model 303) equipped with recorder readout.
See Appendix for petrographical description of the samples.

TABLE 2 (Continued)

	MP 19	MP 20	MP 21	MP 22	MP 23	MP 24	MP 25	MP 26	MP 27	MP 28	G 29	G 30	D 31	D 32	D 33	D 34	D 35	VB 36
SiO_2	57.50	54.40	54.40	54.80	52.90	53.20	54.20	48.60	52.20	53.70	54.40	52.70	50.10	49.60	52.60	51.70	51.10	48.50
TiO_2	<0.10	0.30	0.10	0.10	0.10	0.30	<0.20	1.60	1.20	<0.20	<0.10	<0.10	<0.10	<0.10	<0.10	<0.10	<0.10	1.70
Al_2O_3	15.50	12.90	15.60	15.00	13.50	15.50	15.70	15.80	15.70	14.80	14.30	16.90	13.60	11.20	13.90	12.90	10.70	14.30
Fe_2O_3	0.85	0.99	0.80	0.13	1.26	1.14	0.95	3.27	4.79	1.36	1.18	1.46	0.86	2.16	0.80	2.25	1.61	2.86
FeO	6.28	5.99	7.40	5.73	7.50	6.84	5.16	6.72	5.56	6.27	6.04	3.50	6.78	8.27	6.00	6.32	6.33	10.12
MnO	0.15	0.14	0.13	0.12	0.11	0.25	0.08	0.71	0.12	0.08	0.16	0.10	0.16	0.19	0.14	0.16	0.16	0.21
MgO	9.68	9.04	8.23	9.60	11.96	8.42	7.61	7.39	6.28	9.40	10.11	8.45	11.30	12.29	8.73	8.37	12.32	5.59
CaO	5.59	8.92	2.06	3.12	5.40	6.09	5.71	7.96	4.02	4.91	5.01	9.23	8.86	7.60	6.54	9.50	7.79	11.23
Na_2O	3.78	4.60	4.50	4.25	4.16	5.46	6.22	3.74	6.56	5.31	1.62	2.32	3.76	2.20	2.93	3.32	0.92	3.96
K_2O	0.25	0.04	0.06	0.27	0.11	0.07	0.13	1.07	0.28	0.10	3.51	1.47	0.06	<0.05	1.92	<0.05	0.19	0.28
P_2O_5	0.03	0.06	ND	0.05	ND	0.05	ND	0.14	0.08	T	T	T	T	T	ND	T	0.06	0.27
L.I.	2.00	1.80	5.50	3.90	3.50	2.80	2.20	2.80	3.60	3.40	3.60	3.20	3.30	3.70	2.70	3.50	3.90	2.50
Total	101.61	99.18	98.78	97.07	100.50	100.12	97.90	99.80	100.39	99.33	99.93	99.33	98.81	97.21	96.25	98.02	95.08	101.52
CIPW Norms																		
Q	3.86	--	5.78	3.67	--	--	--	--	--	--	1.80	1.93	--	--	--	0.82	9.08	--
Or	1.48	0.24	0.38	1.71	0.67	0.42	0.80	6.52	1.71	0.61	21.55	9.04	0.37	--	12.14	--	1.23	1.67
Ab	32.09	39.96	40.81	38.59	36.28	47.46	52.89	32.61	56.48	46.83	14.22	20.41	33.31	19.90	26.49	29.71	8.53	28.11
An	24.67	14.82	10.96	16.26	18.39	18.06	15.18	23.87	12.98	16.94	22.18	32.61	21.00	22.13	20.41	21.47	26.87	20.62
Aug	2.38	24.38	--	--	7.43	10.26	11.63	12.82	5.65	6.81	2.91	11.96	20.18	14.69	11.54	22.96	12.09	28.00
Hy	34.09	10.08	35.91	36.90	21.73	2.55	--	1.29	--	12.72	35.54	21.81	1.63	38.84	26.02	21.57	39.46	--
Ol	--	8.29	--	--	13.39	18.81	16.92	14.49	12.97	14.00	--	--	22.18	1.08	2.14	--	--	10.40
Mt	1.23	1.47	1.24	0.20	1.88	1.69	1.43	4.88	7.17	2.05	1.77	2.20	1.30	3.34	1.24	3.45	2.56	4.18
Ilm	0.09	0.58	0.20	0.20	0.19	0.58	--	3.13	2.35	--	--	--	--	--	--	--	--	3.26
Ap	0.07	0.14	--	0.12	--	0.11	--	0.33	0.19	--	--	--	--	--	--	--	0.15	0.63

MP Mafic pillow lava of Betts Cove Ophiolite

G Gabbro (sample 29 is a gabbro screen within the Sheeted Dyke Member)

D Sheeted dyke (sample 31 is a diabase dyke cutting layered ultramafic sequence)

VB Pillow lava of Venams Bight Basalt.

TABLE 2 (Continued)

	VB 37	VB 38	VB 39	VB 40	VB 41	VB 42	VB 43	RH 44	RH 45	RH 46	RH 47	RH 48	RH 49	RH 50	RH 51	RH 52	RH 53
SiO_2	48.50	48.50	46.30	46.40	46.60	48.60	47.20	48.00	47.00	47.50	45.50	49.30	48.90	48.40	48.80	47.40	51.10
TiO_2	1.60	1.70	1.50	1.60	2.10	1.70	2.00	1.90	1.50	1.70	1.20	1.60	1.50	1.70	1.50	1.80	2.20
Al_2O_3	14.90	16.40	14.70	14.70	14.20	14.60	14.10	15.40	15.20	14.90	18.50	15.90	14.30	15.10	14.90	14.30	14.10
Fe_2O_3	2.81	3.18	4.68	5.41	2.31	5.26	5.56	4.32	2.82	4.04	2.37	2.92	2.71	2.86	2.58	5.45	3.54
FeO	8.19	8.33	6.91	7.13	10.26	7.12	7.68	7.51	8.50	7.48	5.92	7.64	9.41	8.23	9.50	7.04	8.83
MnO	0.20	0.19	0.20	0.20	0.20	0.20	0.22	0.21	0.19	0.22	0.13	0.17	0.21	0.14	0.15	0.23	0.28
MgO	6.76	7.16	6.85	6.42	6.02	6.28	5.88	5.89	6.15	6.45	4.46	5.95	6.71	6.79	7.63	6.00	6.25
CaO	10.34	10.03	10.82	11.33	7.30	10.55	9.61	9.72	12.22	11.43	13.75	12.38	9.43	11.40	8.58	10.20	7.84
Na_2O	3.53	3.11	3.01	3.07	4.15	3.38	3.31	3.84	3.24	3.18	2.40	2.62	3.63	2.61	3.88	3.35	3.96
K_2O	0.26	0.41	0.28	0.22	0.53	0.19	0.32	0.62	0.25	0.39	0.63	0.22	0.08	0.30	0.13	0.39	0.45
P_2O_5	0.15	0.20	0.16	0.15	0.29	0.17	0.23	0.20	0.17	0.21	0.13	0.15	0.14	0.16	0.11	0.22	0.26
L.I.	2.70	3.50	3.00	2.50	2.70	2.50	3.20	0.90	2.50	0.8	5.40	1.80	2.50	2.30	2.90	1.10	2.10
Total	99.94	100.51	98.51	99.33	98.56	99.15	100.11	99.71	99.74	98.30	100.39	100.65	99.52	99.99	100.66	97.48	100.91
CIPW Norms																	
Q	--	--	--	--	--	--	0.90	--	--	--	--	0.38	--	--	--	--	0.17
Or	1.58	2.50	1.73	1.34	3.26	1.16	1.95	3.71	1.52	2.36	3.92	1.31	0.48	1.81	0.78	2.39	2.69
Ab	30.71	27.12	26.66	26.82	36.58	29.58	28.89	32.88	23.62	27.59	17.37	22.42	31.65	22.60	33.57	29.40	33.907
An	24.72	30.49	26.98	26.52	19.33	24.94	23.39	23.23	26.93	25.87	39.84	31.33	23.18	29.27	23.38	23.68	19.60
Aug	22.01	15.81	22.62	24.55	13.33	22.58	19.57	19.64	28.05	24.99	25.35	24.29	19.76	22.42	15.87	21.98	14.72
Hy	2.49	0.33	5.07	3.72	6.58	5.57	12.48	2.65	--	0.86	--	12.54	8.62	13.58	3.93	8.24	18.86
Ol	10.80	15.18	6.44	5.43	12.32	4.50	--	7.41	9.84	8.46	5.00	--	8.96	2.35	15.44	2.00	--
Mt	4.19	4.75	7.10	8.10	3.49	7.89	8.31	6.33	4.20	6.00	3.61	4.28	4.05	4.24	3.82	8.19	5.19
Ilm	3.12	3.32	2.98	3.13	4.15	3.34	3.92	3.65	2.93	3.31	2.39	3.07	2.93	3.30	2.91	3.54	4.22
Ap	0.35	0.47	0.38	0.35	0.70	0.40	0.55	0.47	0.40	0.50	0.31	0.35	0.33	0.38	0.26	0.52	0.61

RH Pillow lava of Round Harbour Basalt.

clinopyroxenite from the ultramafic-gabbro transition zone. It shows relatively low MgO and high CaO contents.

Two gabbro samples were analysed, one from a screen (sample 29) within the sheeted dykes and the other (sample 30) from the uppermost part of the Gabbroic Member. Both of them show a marked contrast from other mafic rocks of the Snooks Arm Group in having a higher content of K_2O. This is reflected in the accessory potash felspar observed in the mode. Variation diagrams however suggest that these K_2O values are anomalous and might have been related to alteration (Fig. 44).

Out of five dyke samples, one (sample 31) is a coarse diabase cutting the layered ultramafic rocks and the rest are from the main Sheeted Dyke Member. The felsic perknite dykes (samples 32 and 35), in accordance with their mineralogy, show somewhat higher MgO and lower Na_2O contents than the diabase dykes (samples 33 and 34).

B.2 Pillow Lava Member

B.2.1 Mafic Pillows

All of the ten analysed samples of mafic pillow lavas are rich in Na_2O (3.74 to 6.56 percent) and poor in K_2O (of the order 0.2 percent). This, in conjunction with their mineralogy, suggests that these lavas can be described as spilites, i.e. soda-rich basalts (Turner and Verhoogen, 1960; Hughes, 1972). The author contends that the high Na_2O content, which is in agreement with the sodic plagioclase observed in the mode, was caused by subsequent metamorphism and metasomatism. In the AFM diagram these pillow lavas fall in the calc-alkaline field (Fig.41). This is taken as mainly due to high secondary Na_2O content;

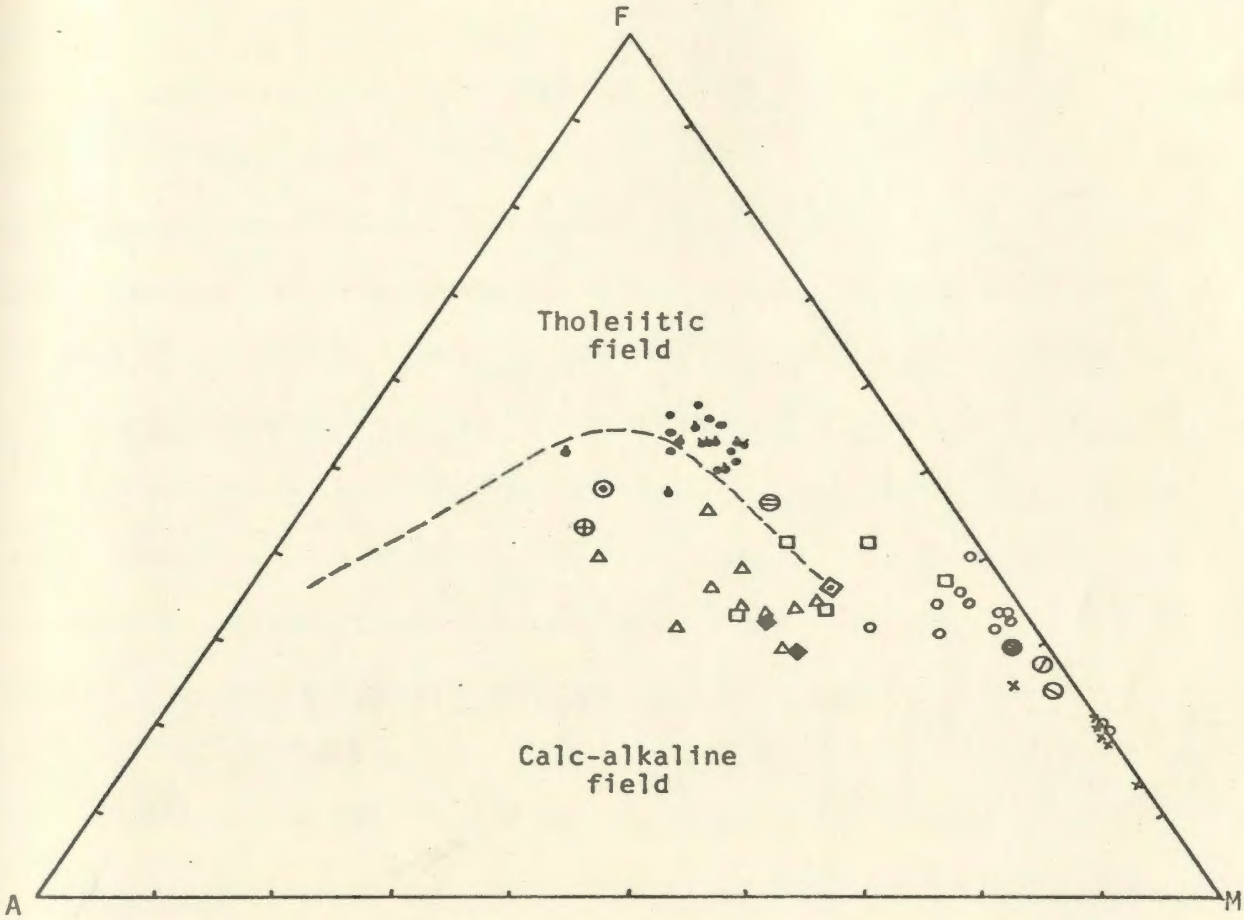


Figure 41: AFM plots of the igneous rocks of the Snooks Arm Group.
A = total alkalis; F = total iron as FeO; M = MgO; all in weight percent.
The dashed line serves to separate tholeiitic and calc-alkaline compositions, after alkaline compositions have been eliminated (cf. Irvine and Baragar, 1971).

- Round Harbour Basalt
- Venams Bight Basalt
- Betts Cove Ophiolite

} Mafic pillow lava

- Ultramafic and related pillow lava
- Sheeted dyke
- Gabbro
- Ultramafic rock

- Average spilite (Papezik and Fleming, 1967).
- Average calc-alkaline lava (Jakes and Gill, 1970).
- Average oceanic tholeiite (Engel et al., 1965).
- Average pyroxenite (Nockolds, 1954).
- Gabbro sample from Indian Ocean (Engel and Fisher, 1969).
- Ultramafic pillow lava from Komati, South Africa (Viljoen and Viljoen, 1969a).
- Ultramafic pillow lava from Troodos, Cyprus (Searle and Vokes, 1969).

it can be seen from this diagram that the average spilite also falls in the calc-alkaline field. In the alkali:silica diagram they all lie in the tholeiite field (Fig. 42). The low normative An content of these pillows puts them in the basalt/tholeiitic andesite boundary region of the normative-plagioclase:colour-index plot (Fig. 43a). For the same reason the majority of these pillows lie too close to the albitic end of the normative-plagioclase:alumina plot (Fig. 43b) and hence this plot can not be applied to distinguish tholeiite from the calc-alkaline series (cf. Irvine and Baragar, 1971, Fig. 6).

The ranges of SiO_2, Al_2O_3, and TiO_2 in these pillow lavas are 46 to 57 percent, 12 to 15.8 percent, and less than 0.3 percent, respectively, values lower than those typical of rocks assigned to the calc-alkaline series (Jakes and Gill, 1970).

B.2.2 Ultramafic and Related Pillows

Twelve samples of ultramafic (perknite) and related (felsic perknite) pillows were analysed, most of which were taken from the lenticular pillow lava outcrop at Partridge Pond. All of these consistently differ from the associated mafic pillows in their higher MgO, much lower Na_2O, and extremely low TiO_2 contents. The ranges of MgO and SiO_2 are 15 to 22 percent and 44.5 to 58.5 percent respectively. The average Al_2O_3 content is about 9 percent although in two samples (9 and 10, Table 2) it constitutes only 0.7 and 0.3 percent. These two samples have the highest MgO contents and form the best examples of ultramfic pillows from the Betts Cove area. Two felsic perknites (samples 14 and 16) are somewhat rich in alkalis. For comparison, plots of average pyroxenite (Nockolds, 1954) and those of two ultramafic

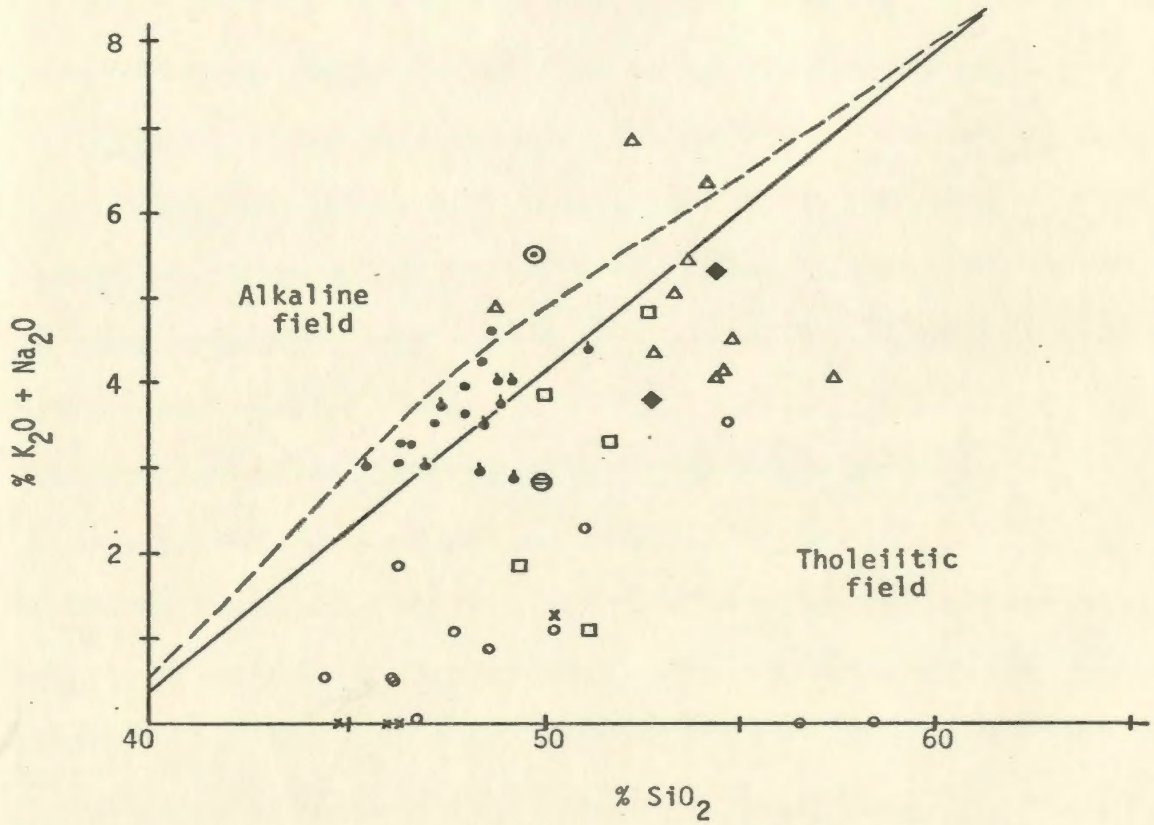


Figure 42: Alkalis:silica plot for the igneous rocks of the Snooks Arm Group. Data in weight percent. Solid line separates the Hawaiian tholeiitic lavas from alkali basalts (MacDonald and Katsura, 1964). Dashed curve is the line suggested by Irvine and Baragar (1971) for making a general distinction between alkaline and subalkaline compositions. See Figure 41 for explanation of symbols.

pillows, one each from Barberton Mountain (Viljoen and Viljoen, 1969a) and Troodos (Searle and Vokes, 1969), are shown in the AFM diagram (Fig. 41). They have considerably higher MgO and lower TiO_2 contents than the average basaltic komatiites described from the Rambler Mine area, Newfoundland (Gale, 1973) and Barberton Mountain, South Africa, (Viljoen and Viljoen, 1969b); Al_2O_3 is present in nearly equal amounts.

The chemical composition, as shown in the variation diagrams (Fig. 44), and the mineralogical composition, show that there is no sharp demarcation between ultramafic and mafic pillows. The felsic perknites, with gradually increasing amounts of felsic minerals, grade into the mafic pillows.

Analyses of three brown-weathering ultramafic rocks with pillow-like shapes are presented in Table 2 (samples 4, 5, 6). As described in the previous chapter, these are intimately associated with the perknite pillows at Partridge Pond. They have low SiO_2 and almost negligible Al_2O_3 contents. In all of these the "loss on ignition" is very high. This may be due to the escape of CO_2 from the carbonates which forms an important component of these rocks. High CO_2 and H_2O contents are also reported from the Troodos ultramafic pillows (Searle and Vokes, 1969). These pillow-like rocks from Betts Cove occupy the same place in the AFM diagram as the lherzolite sample from the Ultramafic Member (Fig. 41).

C. VENAMS BIGHT AND ROUND HARBOUR BASALTS

Eight samples of the Venams Bight Basalt and ten of the Round Harbour Basalt were analysed. These samples come from nearly every

level and every part of these formations (Fig. 40). Samples from both formations show close similarity in their chemical composition. Unlike the Betts Cove pillow lava, in most plots they are concentrated within a small area reflecting their chemical homogeneity. They differ from the Betts Cove pillow lava in the following respects:

(i) generally lower SiO_2 content, varying from 45 to 49 percent;

(ii) consistently higher TiO_2;

(iii) somewhat lower, and less variable Na_2O;

(iv) lower, and less variable, MgO. (Nearly all of the analysed samples are in the 5 to 7 percent range);

(v) higher normative An content (An_{35}-An_{57}) as compared to that of the Betts Cove pillow lava (An_{18}-An_{35}).

In the alkali:silica diagram, the majority of the 18 samples lie on, or slightly towards the alkali side of, the alkali-tholeiite dividing line of MacDonald and Katsura (1964). However, Irvine and Baragar (1971) have suggested that this dividing line should be shifted slightly towards the alkaline field and should be moderately curved (Fig. 42). Following their modified dividing line, all the analysed pillow lava samples of these two formations lie within, although close to the boundary of the tholeiite field. It should be noted that these samples have a relatively low K_2O content and, hence, the high total alkalis is due to the enrichment in Na_2O.

In the normative plagioclase:alumina plot, nearly all of these samples again lie in the tholeiite field (Fig. 43b). Two samples occur anomalously in the calc-alkaline field, most probably due to the presence of plagioclase phenocrysts (cf. Irvine and Baragar, 1971). In the

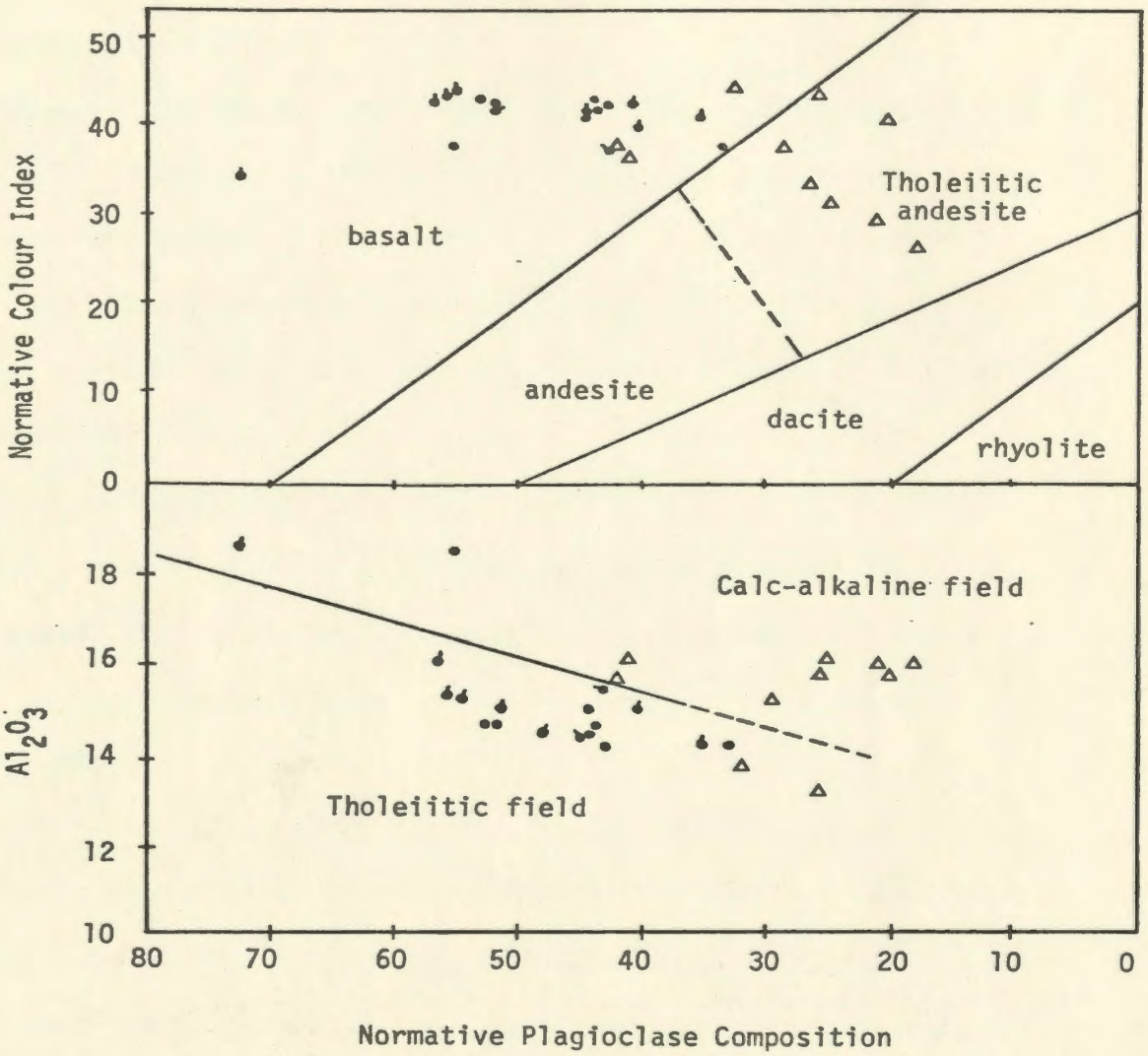


Figure 43: (a) Normative plagioclase: colour index and (b) normative plagioclase: Al_2O_3 plots for the mafic pillow lavas of the Snooks Arm Group (cf. Irvine and Baragar, 1971). See Figure 41 for the explanation of symbols.

normative-plagioclase:colour-index plot they all lie in the basalt field (Fig. 43a). In both of these plots they show a linear trend parallel to the plagioclase abscissa indicating enrichment in normative anorthite content.

In the AFM diagram they are notably concentrated within the tholeiite field (Fig. 41). They also show a higher FeO content than the average oceanic tholeiite.

All the above cited plots, along with the following two observations suggest that the Venams Bight and Round Harbour Basalts are tholeiitic basalts.

(i) A rather low SiO_2 content.

(ii) A TiO_2 content in the 1.2 to 2.20 percent range which, according to Jakes and Gill (1970), suggests an "abyssal tholeiite" type of volcanic rock.

D. AFM DIAGRAM

The AFM (total alkalis:total-iron-as-FeO:magnesia) plots are given in Figure 41. The ultramafic rocks show a moderately-defined linear trend towards the ultramfic and related pillows. However, at a point 5 percent A and 60 percent M, this trend is terminated except for two dyke samples that lie along its projection. The remainder of the mafic rocks (three dykes, two gabbros and nearly all the mafic pillows) of the Betts Cove Ophiolite lie towards the total alkali (A) side of the above trend and towards the calc-alkali side of the Hawaiian alkalic trend. They do not show the linear trend indicative of fractionation. This departure from the main trend was probably caused by metamorphism and metasomatism resulting in Na_2O enrichment.

No fractionation trend for the Betts Cove Ophiolite can therefore be discerned from the AFM diagram.

Plots of average spilite, oceanic tholeiite, calc-alkaline lava, oceanic gabbro, pyroxenite and two ultramafic pillows are presented for comparison.

E. VARIATION DIAGRAMS

Variation of total FeO/(MgO + total FeO) with other major elements is shown in Figure 44, from which the following observations can be made.

1. The Betts Cove Ophiolite and the Venams Bight/Round Harbour Basalts belong to two different groups of magma. This can be seen from the distinct grouping and separate linear trend of the basalts in the TiO_2, CaO, Al_2O_3, FeO, Na_2O, and K_2O plots. The fractionation trends in the two groups of magma show marked contrast, which is especially noticeable in the TiO_2, CaO, Al_2O_3, and FeO plots.
2. Well defined fractionation trends shown by the CaO, Al_2O_3, and MgO plots suggest that the ophiolite suite at Betts Cove was derived by the fractionation of a single parent magma. This is in agreement with conclusions drawn earlier on the basis of field evidence.
3. The Na_2O and K_2O plots show a wide scatter that may reflect alteration related to metamorphism and metasomatism. The Na_2O plot occupies a large area without any well defined trend. It is noticed that the Na_2O content is maximum in the mafic pillows and gradually decreases in the dykes and gabbro,

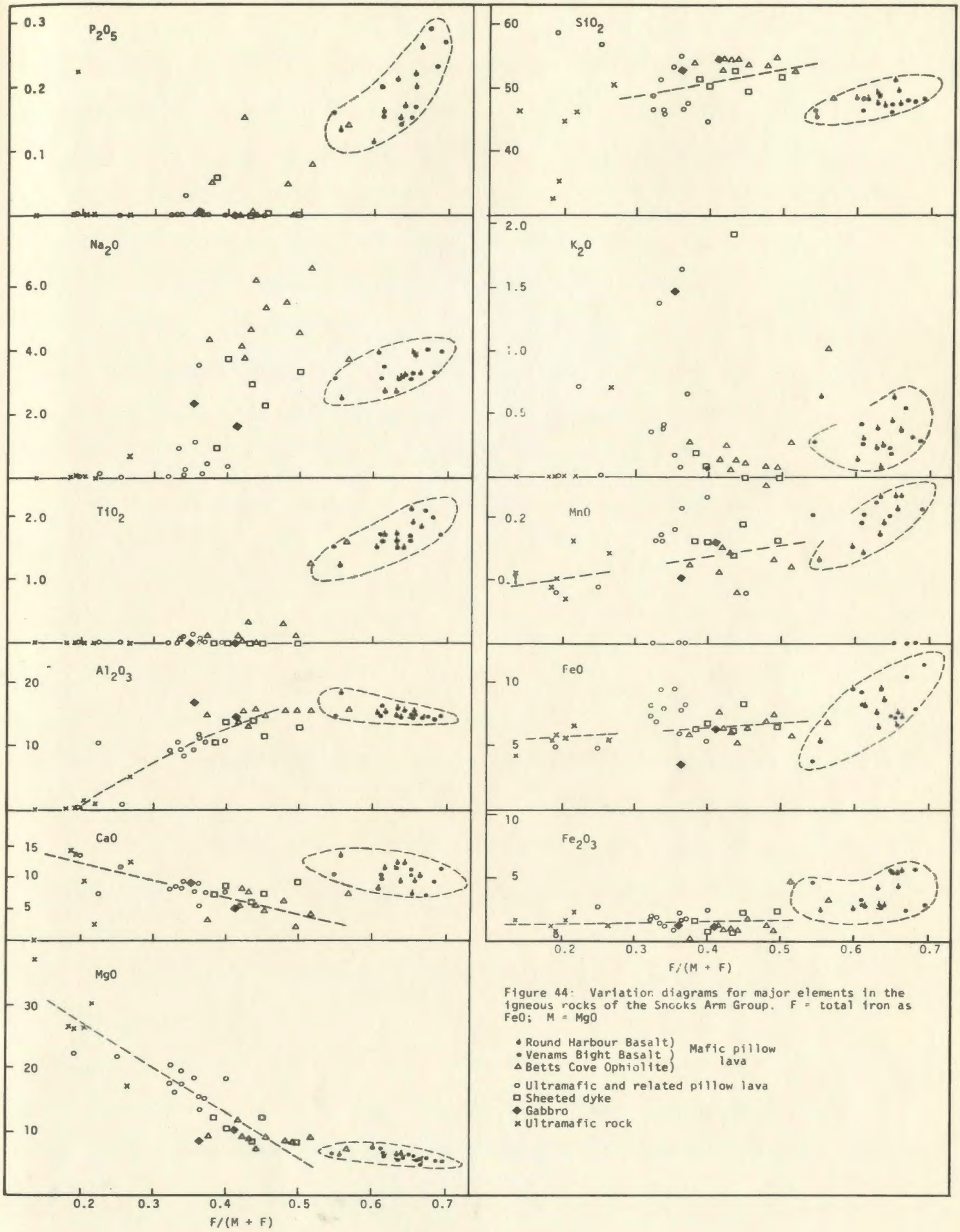


Figure 44: Variation diagrams for major elements in the igneous rocks of the Snooks Arm Group. F = total iron as FeO; M = MgO

- Round Harbour Basalt)
- Venams Bight Basalt) Mafic pillow lava
- Betts Cove Ophiolite)
- Ultramafic and related pillow lava
- Sheeted dyke
- Gabbro
- Ultramafic rock

indicating that the alteration became less effective at greater depths or in less permeable rocks.

4. The Venams Bight and Round Harbour Basalts are coherent in all the plots, indicating a chemical homogeneity and presumably much less alteration than the Betts Cove pillow lava.
5. The plots of Venams Bight and Round Harbour Basalts show a linear pattern which may represent plagioclase fractionation within the lava. This is also substantiated by an identical linear trend in the normative plagioclase plot of these rocks (Fig. 43). Since plagioclase phenocrysts are abundant in these pillow lavas, it is likely that the variation in plagioclase composition was a dominant process during fractionation.

F. SUMMARY AND DISCUSSION

The Betts Cove ophiolite suite was formed by the fractionation of a single parent magma. The pillow lavas of this suite are oceanic tholeiites (now spilites) in which the soda enrichment seems to have been related to metamorphic amd metasomatic processes at the oceanic ridge. These processes produced the high total alkalis that cause departures from some possible fractionation trends in the AFM and other variation diagrams.

Pillow lavas of the Venams Bight and Round Harbour Basalts are tholeiites which were derived from a different magma than that of the Betts Cove Ophiolite. The plots of these basalts in the alkali:silica diagram lie close to, but on the alkali side of, the alkali/tholeiite dividing line of MacDonald and Katsura (1964). This, coupled with

relatively higher TiO_2 contents, suggests that these tholeiites may be of alkaline affinity. Thus the Venams Bight and Round Harbour pillow lavas may have originated from a deeper source than rocks of the Betts Cove Ophiolite and, hence, may represent an off-axis volcanism such as that suggested for the Upper Pillow Lavas of Troodos ophiolite (Gass and Smewing, 1973). The Venams Bight and Round Harbour Basalts are chemically homogeneous. A progressive change in plagioclase composition was probably a dominant process during their fractionation.

CHAPTER V

ORIGIN, CORRELATION AND TECTONIC EVOLUTION OF THE SNOOKS ARM GROUP

A. SNOOKS ARM GROUP AS OCEANIC CRUST AND MANTLE

The presence of sheeted dykes, as stated in Chapter III, perhaps constitutes the best evidence that the Betts Cove ophiolite represents a fragment of oceanic crust and mantle. The origin of a complex consisting of nearly 100 percent dykes can be best explained by the mechanism of sea-floor spreading (cf. Gass, 1968). The internal stratigraphy and the highly transitional contacts between various members of the ophiolite suite are other important features that closely resemble interpretations of seismic data and dredge hauls from modern oceanic crust and mantle (Cann, 1970). Since the Bobby Cove, Balsam Bud Cove, Venams Bight, and Round Harbour Formations conformably overlie the Betts Cove Ophiolite, the whole Snooks Arm Group should represent a single suite probably transitional from oceanic crust (at the base) to island arc (at higher levels). Also Snooks Arm volcanic rocks are chemically similar to those dredged from modern ocean floors.

Moores and Vine (1971) and Dewey and Bird (1971, p. 3186) have discussed some possible criteria for distinguishing the slow-spreading oceanic ridges from the fast-spreading ones. Moores and Vine, drawing from their observations of the Troodos ophiolite, suggest that if well developed sheeted dykes are present and if tectonic banding is

normal to the crust-mantle transition, the ophiolite was produced at a slow-spreading ridge. From this it is inferred that the Betts Cove ophiolite was produced at a slow-spreading ridge although tectonic banding is not pronounced in this case. Other features, based on Dewey and Bird's (1971) criteria, that suggest a relatively slow-spreading rate for the Betts Cove ophiolite are: (i) complex Mohorovicic transition zone and (ii) higher and much better concentrated base-metal deposits (Upadhyay and Strong, 1973).

B. CONCEPT OF A MARGINAL OCEAN BASIN

Marginal ocean basins are those semi-isolated basins, or series of basins, that have intermediate to normal water depths and that lie between continents and on the concave side of island arc systems such as those of the western Pacific. For detailed internal stratigraphy, structure, and evolution of marginal ocean basins the reader is referred to the papers by Menard (1967), Karig (1971a), and Pakham and Falvey (1971).

The following features suggest that during Early Ordovician time the Snooks Arm Group formed the floor of a small ocean basin that was marginal to the main Proto-Atlantic Ocean (Wilson, 1966).

1. The Andesitic agglomerate, tuff, and flysch of the Snooks Arm Group are typical of those deposits derived from island arcs (Mitchell, 1970; Mitchell and Reading, 1971) and therefore they indicate the presence of a nearby island arc. This proximity to an island arc may be indicative of a marginal ocean basin.
2. (a) The Betts Cove Ophiolite, at least as presently preserved, is too thin to represent a main oceanic crust (see Fig. 46).

(b) Modern floors of major oceans consist of a suite of magmatic rocks which is overlain by a thin cover of sediments here considered equivalent to the Betts Cove Ophiolite and the Bobby Cove Formation respectively. These major oceans do not generally contain deposits comparable to the alternating sedimentary and volcanic rocks of the uppermost three formations of the Snooks Arm Group. Such thick deposits are, however, known to occur in marginal basins such as the Aleutian Basin (Shor, 1964), the sea of Okhotsk (Averyanov *et al.*, 1961), and the Balaeric Basin (Hersey, 1965).

3. The rhyolitic and quartz-felspar porphyry clasts in the Snooks Arm sediments strongly resemble the nearby Cape St. John (Fleur de Lys Supergroup) lithologies, indicating that the basin was very close to a continental margin. The high angularity of these clasts also suggests a very short transport, further attesting to a nearby source.
4. The orientation of sheeted dykes in Newfoundland ophiolites (Upadhyay *et al.*, 1971; Strong, 1972; Williams and Malpas, 1972) is oblique to the suggested axis (ridge) of the main Proto-Atlantic Ocean (Wilson, 1966; Bird *et al.*, 1971). It is possible that such irregular and oblique attitudes of dykes are caused by diffuse spreading at the less sharply defined ridges generally believed to characterize marginal basins.

The structure and origin of marginal basins is less well understood than those of main oceans. It has been suggested that

marginal basins also contain ridges or ridge-like features that may be extinct or active (van der Linden, 1969). Magnetic anomaly patterns in such basins are complex. They generally show anomalously high heat flow (Sleep and Toksöz, 1971; Packham and Falvey, 1971). These basins are generally understood to have developed by a diffuse, slow-spreading process whereby the island arcs move away from the continental margins (Karig, 1971a, b). This is in accordance with the fact that the crust of marginal basins is younger than some orogenic events of adjacent arcs. The spreading is apparently caused by thermal upwellings behind the arcs (Karig, 1971a; Matsuda and Uyeda, 1971). These upwellings probably originate by partial melting of the down-going lithosphere slab. It is envisaged that this process would be of a more sporadic nature and would probably occur along patchy diffuse zones rather than along sharply defined ridges. If subduction does not occur at the marginal basin/continent interface, a continuous spreading of the basin would ultimately result in the migration of the island arc away from the continent. Dewey and Bird (1971) have suggested that with the expansion of the marginal basin the locus of spreading is likely to shift away from the continental margin and, as a result, the arc/trench belt would move towards the main ocean. This would give rise to an asymmetric form of spreading whereby the addition of new oceanic crust and mantle is concentrated in a linear zone closer to the bordering island arc. Sleep and Toksöz (1971, Fig. 1a) have suggested that the spreading may be triggered by hydrodynamic forces generated by the sinking of the down-going slab and producing a flow pattern in the asthenosphere, which, in turn, acts on the lithosphere. It has been

argued (Moores, 1970) that the spreading in marginal basins may continue to the extent that the arcs may sweep across the main ocean and finally collide with the opposing continental margin. Dewey and Bird (1971) have disputed such a notion and, instead, suggest that at a particular stage the marginal basins begin to contract by the development of subduction zone(s) at or near their transition to island arcs and are eventually obducted during the ensuing continent/arc collision.

C. GENESIS OF BETTS COVE OPHIOLITE

Any model for the genesis of the Betts Cove ophiolite should be consistent with the following observations.

(1) The cumulate ultramafic rocks, in the upper part, are interlayered with gabbro and gradually become insignificant within the Gabbroic Member.

(2) Some of the diabasic and ultramafic dykes of the sheeted complex cut the layered ultramafic-gabbro sequence.

(3) In Betts Cove area, the gabbro, which is generally leucocratic and coarser than the diabase dykes, occurs as lenticular or podiform screens which lack chilled margins (Figs. 25, 30).

(4) The gabbro does not cut the pillow lava whereas the (sheeted) dykes do. This, in conjunction with observation (3), suggests that the gabbro could not have been intrusive into the dyke-pillow lava assemblage.

(5) The layering in the ultramafic-gabbro sequence is at high angles to the plane of sheeted dykes.

(6) High-pressure mineral assemblages such as garnet or spinel are

not known to occur.

The above observations suggest, among other things, that (a) the cumulate ultramafites and gabbro were formed by differentiation of a single magma under low-pressure conditions and (b) a hiatus existed between the formation of the ultramafic-gabbro sequence and that of the sheeted dyke-pillow lava assemblage. This hiatus is similar to that in the Canyon Mountain (Thayer and Himmelberg, 1968) and Troodos (Gass, 1968) ophiolites. However, in contrast, an intrusive relationship between gabbro and other members of the ophiolite suite has been reported from the Semail complex, Oman (Reinhardt, 1969) and from Liguria (Bezzi and Piccardo, 1971).

In the model proposed here it is envisaged that the parent material for the formation of the Betts Cove ophiolite suite was derived from depth, through partial melting of the upper mantle. The mantle material may have been pyrolitic* in composition. The process of partial melting, as suggested by Green and Ringwood (1967), can be summarized in the following steps:

(1) Diapirs of solid pyrolite rise adiabatically from a depth of 100 kilometers or more.

(2) The generation of magma through partial melting of the rising pyrolite takes place at much shallower depths, perhaps about 30 kilometers below surface (Green and Ringwood, 1967; Thayer, 1969).

---

*Pyrolite (Green and Ringwood, 1967) is a postulated hypothetical mantle rock whose composition is equivalent to 3 parts anhydrous dunite and 1 part tholeiitic basalt.

(3) The partially molten mass continues to rise adiabatically probably with the liquid remaining in equilibrium with the residual unmelted crystals of the mush.

(4) This liquid is segregated from the mush and is emplaced in small chambers. The mush, which is left behind, would represent the "depleted" upper mantle. These chambers would be at the same depth as the Mohorovicic discontinuity in oceanic crust or, as interpreted here, the depth of cumulate ultramafites in ophiolites, which is generally in the range of 6 to 10 kilometers.

(5) Differentiation within the magma chamber gives rise to a part of the ophiolite suite.

Stage (5) is of special concern here and is elaborated below to explain the origin of Betts Cove ophiolite (Fig. 45).

The entire process summarized above, took place in a tensional environment and marked the initiation of sea-floor spreading. The high mafic to ultramafic ratio in the Betts Cove ophiolite suggests that the magma that gave rise to the ophiolite suite was basaltic in composition. The magma was emplaced in a lopolith-like chamber. Differentiation of this magma took place in two phases, separated by a considerable time interval. This is suggested by the "cold" screens of gabbro and ultramafic rocks that were picked up by the dykes of younger phase. The first phase produced cumulate ultramafic rocks at the base, overlain successively by a pyroxenite-gabbro transition zone and then the gabbro layer. This ultramafic-gabbro mush was overlain by some of the remaining undifferentiated liquid (Fig. 45a). During

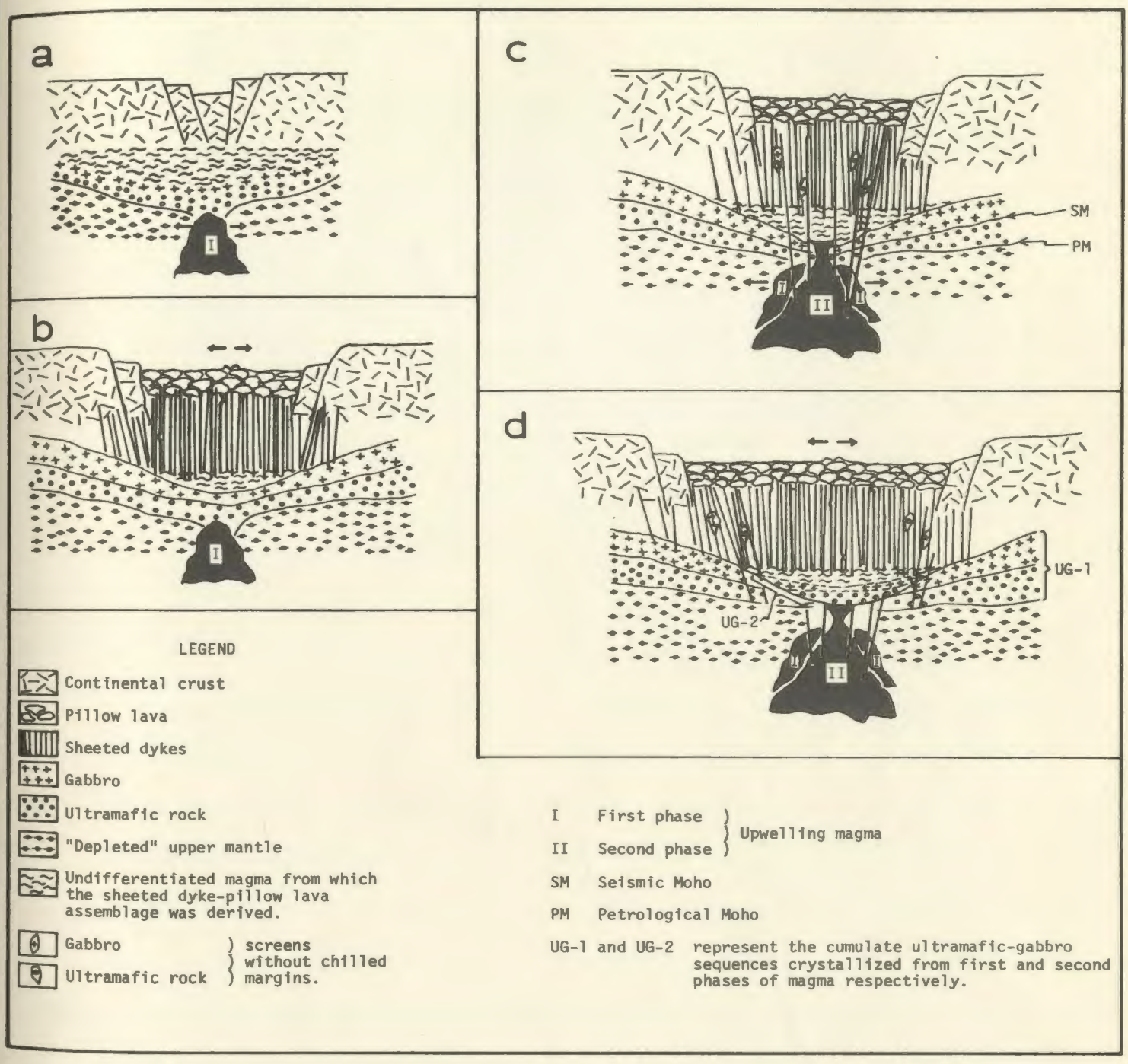


Figure 45: Schematic sections illustrating the proposed model for the genesis of Betts Cove ophiolite. Discussion in text.

the second phase of differentiation, sheeted dykes and pillow lavas were derived from this remaining liquid. The presence of extremely thin layers that continue along strike for tens of meters and of size-graded layers in the Betts Cove ultramafic rocks, suggest that the ultramafic-gabbro sequence was formed through gravitative differentiation in a relatively calm environment. Minor flow and movement within the magma chamber is indicated by the cross-laminated ultramafic layers. In a similar model proposed for the evolution of Troodos ophiolite, Greenbaum (1972) has suggested that the thermal gradients were more or less horizontal so that the development of vertical convection currents was at a minimum.

The formation of the sheeted dykes-pillow lava assemblage took place in a tensional environment so that the block-faulted "cover" of continental rocks above the magma chamber began to distend and finally broke apart (Fig. 45b). This provided room for the injection of dykes, the emplacement of which was further facilitated by the pressure built within the underlying magma chamber. These dykes, after reaching the surface, became flows which developed pillow structures. The dykes which initially followed the vertical to subvertical tension cracks were at a high angle to the layering in the underlying ultramafic-gabbro sequence. As the tension prevailed more dykes intruded along planes that were parallel to the preceeding dykes. Since every dyke would tend to make room for itself, a lateral spreading must result. Since all the dykes originate from the basaltic liquid on top of the gabbro mush, the contact between gabbro and sheeted dykes would be generally gradational, a feature which is observed in Betts Cove.

A partial segregation of the predominantly basaltic liquid into mafic and ultramafic parts might have taken place so that the dykes and pillows of diabasic and ultramafic composition would be derived from the liquids of corresponding composition. Those dykes that happened to originate from the trough of the magma chamber and which were intruded obliquely to the generally vertical dykes, might also have cut the ultramafic-gabbro wall rocks and picked up their pieces as screens.

As the basaltic liquid is consumed in the formation of dykes and pillow lava, replenishment would take place through another upwelling of partially molten pyrolite (Fig. 45c). The differentiation of this melt would take place again in two phases exactly as described earlier for the first phase (Fig. 45d). This new plume of magma, in order to make room for itself, would cause distention and necking apart of the layered ultramafic-gabbro sequence related to the first phase of magma. If this sequence was not fully solidified, as is envisaged here, it would wedge out towards the trough of the chamber so that all the cumulate ultramafic-gabbro sequences related to different phases of magma emplacement would combine to produce a somewhat continuous and subhorizontal contact zone between the ultramafic rocks and gabbro. Evidence for a gradual thinning (wedging) of the gabbroic part towards oceanic ridges, has also been presented by LePichon (1969). Such renewed phases of magma emplacement are necessary to cause spreading in the gabbroic-ultramafic part of the ophiolite suite. Similar repetitive phases are also known to occur at the modern mid-oceanic ridges.

When a new phase of magma is emplaced, some of it would intrude the pre-existing ultramafic-gabbro sequence in the form of dykes, even before entering the magma chamber (Fig. 45c). These dykes might also incorporate and carry small pieces of pre-consolidated ultramafic rocks and gabbro to higher levels so that they would appear as screens within the sheeted dykes. The author finds this a more realistic mechanism for emplacement of the screens than the one suggested earlier which involves oblique dykes cutting through the wall rocks.

The process of sea-floor spreading continues through such repeated phases of magma generation and subsequent differentiation and emplacement. In order to cope with the spreading resulting from the formation of sheeted dykes at higher levels, tension fractures may develop in the underlying ultramafic-gabbro sequence. These fractures may be channels to the ensuing new phase of magma coming from depth. Spreading of the whole ophiolite, therefore, would take place in two distinct parts. The major spreading in the lower part, involving cumulate ultramafic-gabbro sequence and the underlying "depleted" mantle, would occur during the emplacement of the partially molten pyrolite body. In the upper parts, spreading would take place during the formation of sheeted dykes. Temporally, the two might alternate, overlap or, less commonly, be simultaneous.

During upwelling of the partially molten pyrolite, spreading may take place within the "depleted" material beneath the cumulate ultramafic rocks. This would also result in considerable lateral gliding within the basal parts of the cumulate ultramafic rocks and would give rise to tectonites in which individual crystals are stretched

and tapered parallel to the layering (Avé Lallemant and Carter, 1970; Nicolas *et al.*, 1971). The growth of such tectonites would tend to be less prominent at higher levels.

D. GENESIS OF VENAMS BIGHT AND ROUND HARBOUR BASALTS

Marginal basins experience repeated volcanism and sedimentation as stated earlier in this chapter. It is envisaged here that the magma which feeds this volcanism is derived by partial melting of a lithospheric plate undergoing subduction behind a bordering island arc. The author contends that the deposition of sedimentary material represents relative quiescence caused either by a temporary pause in the movement of the down-going lithospheric plate or by a "smooth" subduction so that no frictional heat can be generated to produce the hot diapiric upwellings (Hasebe *et al.*, 1970). From a study of the Mariana Island arc system, Karig (1971b) has suggested that the tectonic activity in that area is discontinuous. Prevalence of a tectonic quiescence during the sedimentary episodes in the Snooks Arm Group therefore would not be unexpected. The pillow lava-sill association in these is quite similar to that in the Betts Cove Pillow Lava Member where the sills swing and assume dyke-like attitudes in the underlying Sheeted Dyke Member. From this similarity it is speculated that the sills in the Venams Bight and Round Harbour Basalts probably also had the attitude of dykes at depth. These dykes are not exposed and it is possible that they are underlain by the waters of Green Bay to the south. However, at one place on the shore of Snooks Arm a diabase dyke was found that cuts across the bedding of the sediments, indicating the postulated dyke-like attitude of the sills

at depth. Likewise, diabase dykes/sills running along as well as across the strike of the Round Harbour Basalt (e.g. at Snooks Head) show that changes in attitude from dyke to sill were common.

Chemically, the Venams Bight and Round Harbour Basalts are tholeiites. They do not belong to the same magma group as the Betts Cove ophiolite. Field evidence suggests that the diabase dykes/sills that fed these pillow lavas also intruded the underlying sedimentary/pyroclastic formations. This is supported by the striking petrographical similarity between some of the sills in these formations and those in the overlying pillow lavas.

E. NEW CONCEPTS CONCERNING OPHIOLITE SUITES

E.1. Introduction

Lotti (1886) is regarded as the first to recognize a distinct suite of ultramafic rocks, gabbro, pillow lava, and chert from his work on the island of Elba. Steinmann's (1927) work on the Mediterranean ophiolites, however, remains the definitive early account on the subject. Steinmann defined ophiolite as an ordered suite of rocks with serpentinite at the base, overlain successively by gabbro, pillow lava, and chert. Benson (1926), thinking along the same lines as Steinmann, referred to the ophiolites as "Green Rocks of the Alpine Type" and emphasized the close spatial and genetic relationship between coarse-grained plutonic rocks at the base and fine-grained volcanic rocks on top.

In the context of sea-floor spreading and plate tectonics, the ophiolite suite has been reviewed in recent years by Moores (1970) Coleman (1971a), Dewey and Bird (1971), and Church and Stevens (1971).

It is now generally held that an ophiolite suite is a sequence of rocks from partially serpentinized ultramafic rocks, through gabbro, with or without sheeted dykes, overlain by pillow lava. This sequence is capped by sediments in most cases. Their occurrence in orogenic belts, transitional contacts between various members, presence of extrusive rocks on top, and their generally faulted base, are some of the field criteria that help distinguish them from layered complexes such as Muskox, Bushveld, Skaergaard, and Stillwater. Metamorphic aureoles may or may not be developed around ophiolites (Thayer, 1960; Malpas *et al.*, 1973; Williams and Smyth, 1973).

Due to intense tectonism in orogenic belts, ophiolites may not be preserved as a complete suite and, instead, their various members may occur as disjunct, shredded slices in mélange terrains. If there is any evidence for a consanguity among these isolated members, the prefix "ophiolitic" should be used to describe them.

E.2 Origin of Ophiolites

Three pertinent hypotheses for the origin of ophiolites are summarized and discussed below.

E.2.1 Plutovolcanic Hypothesis

This hypothesis assumes that a massive submarine mafic lava flow is extruded on the ocean floor, the outer skin of which is chilled to form pillow lavas and the inner core, due to slower crystallization, gives rise to coarse-grained plutonic rocks. The chief proponents of this hypothesis are Bailey and McCallien (1953), Brunn (1954, 1960), and Maxwell and Azzaroli (1962). It has also been applied to the ophiolites of southern Quebec by Lamarche (1972).

There are several objections (e.g. Vuagnat, 1963; Thayer, 1967) to this hypothesis. (i) It does not explain the mode of emplacement of sheeted dykes that intrude the ultramafic,gabbro and pillow lava members under tensional environment. (ii) Geological and geophysical evidence suggests that the magmatic material at the oceanic ridges is emplaced as thin linear slices at more or less regular intervals. Extrusion of a huge tongue of lava in a single phase therefore seems unlikely. (iii) A large magmatic flow on surface should give rise to a complete envelope of pillow lava and perhaps at the bottom of such flow a reverse succession (i.e. plutonic rocks on top and pillow lava at the bottom) should be observed. The present author is not aware of any evidence for these features. (iv) On the basis of a high ultramafic/mafic ratio in some ophiolites, Thayer (1967) has argued against mafic magma as the parent magma.

E.2.2 Oceanic Plate Hypothesis

According to this hypothesis the magmatic material is derived from below the oceanic crust and a hiatus exists between the ultramafic part (mantle) and the gabbro-pillow lava assemblage (oceanic crust). The gabbro and pillow lavas were presumably intruded through the ultramafic rocks whose presence on the upper parts of the oceanic plate, according to the hypothesis, would be coincidental. This hypothesis has been advocated by De Rover (1957), Gass and Masson-Smith (1963) and Hess (1964, 1965).

Two factors argue against this hypothesis: (i) In most ophiolites a gradational contact exists between ultramafic rocks, gabbro, sheeted dykes (if present), and pillow lava. (ii) Chemical

evidence from Vourinos and Kizil Dagh (Nicolas, 1966), Troodos (Bottcher, 1969), Papua (Davies, 1971), and Betts Cove (see Chapter IV), suggests that in these ophiolites a continuous differentiation trend exists from ultramafic rocks through gabbro to pillow lava.

E.2.3 Mantle Fusion Hypothesis

This hypothesis postulates that the material for the formation of oceanic crust is derived by partial fusion of the mantle which is emplaced as a deforming and differentiating solid-liquid mass at the oceanic ridges (Moores, 1969; Maxwell, 1969). It assumes that all members of the ophiolite suite are consanguineous and cogenetic.

This hypothesis accounts for most of the field relationships observed in ophiolites although it is not always substantiated by the chemical data (e.g. Moores, 1969). The proposed model for the origin of Betts Cove ophiolite is based on this hypothesis and the details of the postulated process have been presented earlier in this chapter.

E.3 Emplacement of Ophiolites

It has been suggested (Dewey and Bird, 1971) that a considerable time gap may exist between the origin and the emplacement of an ophiolite suite. Some field observations such as the apparent lack (? destruction) of metamorphic aureoles, presence of basal thrust contacts, and occurrence as disjunct slices in mélange terrains do suggest that some ophiolites were probably emplaced by tectonic processes and did not necessarily originate at the site of their present position.

Emplacement of ophiolites may be initiated by the closing of an ocean/geosyncline and may be finally accomplished at the time of collision between the bordering continent and the contracting oceanic

crust at high-angle faults. When the oceanic plate meets the continental margin either or both of the following two things may happen: (i) The oceanic plate may be thrust (subducted) beneath the continent along the trenches and destroyed at depth. (ii) It may be thrust (obducted) over the continent and remain there as an ophiolite outcrop. However, obduction of shredded slices detached from the main oceanic plate may also accompany a major subduction process. This is stated to be the case in the western California Cordillera (Bailey *et al.*, 1970). When the oceanic plate collides against the continent, whether related to subduction or obduction, high-pressure blue schists may be developed (Ernst *et al.*, 1970). If a relatively "hot" oceanic plate, probably originating during the inception of sea-floor spreading (Church and Stevens, 1971), is obducted over a continent, a high-temperature metamorphic aureole may develop (Malpas *et al.*, 1973; Williams and Smyth, 1973).

More extensive treatment of this subject is beyond the scope of the present work and the interested reader is referred to the papers by Dietz (1963), Dewey and Bird (1970, 1971), Coleman (1971a), and Dickinson (1971), on which the foregoing account is largely based.

F. COMPARISON WITH MODERN OCEANIC CRUST AND OCEAN FLOOR

Seismic and dredging work in oceanic areas in recent years has provided a coherent picture of the probable structure and internal stratigraphy of oceanic crust (e.g. Cann, 1968; 1970; Christensen, 1970). Figure 46 (column 1) shows the postulated sequence (layers) within a typical oceanic crust and the seismic velocities therein. Layers 1 and 2 consist of sediments and pillow lava respectively.

The composition of layer 3 is controversial. According to Hess (1962), it consists of serpentinized peridotite. Others (e.g. Christensen, 1970; Cann, 1968) have suggested that layer 3 is more likely to be comprised of amphibolite, gabbro, or a combination of gabbro, diabase, and their metamorphosed equivalents. Layer 3 is underlain by a material in which the seismic velocity is 8.1 km/sec and this material is interpreted to be of peridotitic composition and forms the upper mantle. An abrupt change in seismic velocity from 6.8 km/sec to 8.1 km/sec takes place at the layer 3/upper mantle interface and this interface is referred to as the Mohorovicic discontinuity or simply as the "Moho".

The Ultramafic, Gabbroic plus Sheeted Dyke, and Pillow Lava Members of the Betts Cove Ophiolite are considered to be equivalent to the upper mantle, layer 3, and layer 2 respectively of the oceanic crust; Bobby Cove Formation is correlated with layer 1, the partly consolidated sedimentary unit that overlies the pillow lava of the oceanic crust (Fig. 46, column 6). It will be noticed that the Betts Cove ophiolite, if we disregard the ultramafic part, has a thickness only about one third of that of a typical oceanic crust. Marginal basins, e.g. the Sea of Okhotsk generally have much thicker sedimentary layers. Some marginal basins such as the Aleutian and the Balearic basins contain many more layers than the typical oceanic crust (Fig. 46). This may be a manifestation of repeated volcanism and sedimentation as in the Snooks Arm Group. Judging from the seismic velocities, layers with "gabbroic velocities" (around 6.8 km/sec) are much thicker in the marginal basins than the Gabbroic Member of the Betts Cove

Ophiolite.

Most of the dredge hauls recovered from ocean floors are sediments and basalts, although some ultramafic samples have also been obtained chiefly from fracture and fault zones. The composition and other aspects of these samples can be duplicated within rocks of both the Betts Cove Ophiolite and the Snooks Arm Group as described below.

(1) At Palmer Ridge in the northeast Atlantic, suites of serpentinite, amphibolite, gabbro, and limestone have been found in four different hauls. This association has been interpreted as depositional rather than intrusive (Cann and Funnel, 1967). A similar suite consisting of peridotite, gabbro/greenschist, and basalt has been reported from the Mid-Atlantic Ridge (Bonatti, 1968). Both of these suites are similar in their apparent internal stratigraphy and transitional contacts to various members of the Betts Cove and other ophiolites.

(2) Crystal layering has been reported in the partly serpentinized ultramafic rocks from the Mid-Atlantic Ridge (Aumento and Loubat, 1971).

(3) Ultramafic and gabbroic rocks with cumulate textures are known from the Romanche Fracture in the Atlantic Ocean (Melson, 1969).

(4) Serpentinized ultramafic rocks similar to those of Betts Cove, have been described from the Indian Ocean (Chernysheva and Berzrukov, 1966; Udintsev, 1969) as well as the Atlantic Ocean (Aumento and Loubat, 1971). Serpentinite breccia of unspecified origin, occurs at the Palmer Ridge (Cann and Funnel, 1967).

(5) Alteration of ultramafic rocks, such as repeated phases of

serpentinization, bastite pseudomorphs after orthopyroxene, talc-carbonate growth, etc. (Aumento and Loubat, 1971) are quite similar to the alteration of Betts Cove ultramafic rocks.

(6) Uralitization of pyroxenes, which is the most common alteration in Betts Cove area, is also recorded from the floor of the Atlantic Ocean (Nicholls *et al.*, 1964; Aumento and Loubat, 1971).

(7) Rocks consisting entirely of talc, actinolite, and chlorite, in association with spilitic pillows at the Carlsberg Ridge (Cann and Vine, 1966) are comparable to the ultramafic and related pillows described from Betts Cove except that in the latter talc is somewhat less and chlorite is uncommon.

(8) Rodingites, like those of Betts Cove, also occur at the Mid-Atlantic Ridge (Aumento and Loubat, 1971).

(9) Dolerite and gabbro have been dredged from the Atlantic Ocean (Nicholls *et al.*, 1964; Cann and Funnel, 1967). Such dolerites, in which chilled margins are rarely seen due to the generally small size of the samples, may be equivalent to the sheeted dykes in ophiolites. However, Engel and Fisher (1969) have reported a gabbro sample cut by a 15-centimeter diabase dyke from the Indian Ocean. Similarly, ultramafic rocks intruded by gabbro and dolerite are known from the same ocean (Udintsev, 1969). Both of these relations are observed in the Betts Cove ophiolite.

(10) Diorites, in association with basalt, gabbro, and serpentinite, occur at the High Fractured Plateau of the Mid-Atlantic Ridge

(Aumento, 1969).

(11) The rhyolitic material in the Betts Cove ophiolite, even though in small amount, can be compared with the rhyolites and dacites of central Iceland (e.g. Carmichael, 1964) where the Mid-Atlantic Ridge rises above the sea level. Other silicic flows from the oceanic areas include a "differentiated andesitic variety" from the East Pacific Rise, which, from Hart's (1969) data, has been classified as "dacite" by Davies (1971).

(12) The composition and association of the spilites of the Betts Cove pillow lava are similar to those recorded from the Indian Ocean (Cann and Vine, 1966; Cann, 1969). Likewise, the tholeiites of both Venams Bight and Round Harbour Basalts are similar to innumerable oceanic basalts (Engel *et al.*, 1965; Hekinian, 1968; Aumento, 1968; Kay *et al.*, 1970).

G. COMPARISON WITH OTHER OPHIOLITES

Cross-sections of various well known ophiolites are shown in Figure 46. All of them have identical stratigraphy although the thicknesses vary from one suite to another. All these ophiolites, except the Vourinos and Papuan occurrences, contain sheeted dykes. The bases of most of them are faulted. In some of them, e.g. Bay of Islands (Malpas *et al.*, 1973; Williams and Smyth, 1973), Oman (Reinhardt, 1969) and Papua (Davies, 1968), garnet-amphibolite zones are developed at the base. The Betts Cove ophiolite differs from the others in two respects: (i) The gabbro unit is anomalously thin and (ii) the sediments and volcanic rocks on top of this ophiolite are not known from other ophiolites. The total thickness of the Betts Cove

Figure 46: Comparative sections of oceanic crust and mantle (columns 1-5), and ophiolites and related island arc assemblages (columns 6-20). (1) Typical section of oceanic crust and mantle (Dewey and Bird, 1971). (2) Balearic basin (Hersey, 1965). (3) Sea of Japan (Menard, 1967). (4) Aleutian basin (Shor, 1964). (5) Sea of Okhotsk (Averyanov et al., 1961). (6) Snooks Arm Group. (7) Western Arm Group with Lush's Bight pillow lavas (LB) at the base; a: Harry's Harbour Section, b: Western Arm Section (Marten, 1971). (8) Cutwell Group (Kean, 1973). (9) Lush's Bight Group, Pilleys Island area (Strong, 1972). (10) Moretons Harbour Group (Strong and Payne, 1973. (11) Nippers Harbour Group. (12) Baie Verte ophiolite (W.S.F. Kidd, in Dewey and Bird, 1971). (13) Mings Bight ophiolite (R. Norman, personal communication, 1973). (14) North Arm Mountain section, Bay of Islands complex (J. Malpas, personal communication, 1973). (15) Hare Bay ophiolite (Williams, 1971). (16) Semail complex, Oman (Reinhardt, 1969). (17) Troodos complex, Cyprus (Gass, 1968; Moores and Vine, 1971). (18) Papuan ophiolite (Davies, 1971). (19) Canyon Mountain complex, Oregon (Thayer and Himmelberg, 1968). (20) Vourinos complex, Greece (Moores, 1969).

A to E: Notations for the correlation between Snooks Arm and Western Arm Groups (see Table 3 for names of formations, etc.). Figures in columns 1 to 5 represent seismic velocities in km/sec.

ophiolite compares closely with that of Troodos. Vourinos, Canyon Mountain, Papua, and Bay of Islands differ in having significant amounts of diorite in the gabbro unit. The sheeted dyke column has a common height of about 700 meters in all of them. The total thickness of the pillow lava unit is of the order of 1 to 1.5 kilometer(s), although in Mings Bight (Fig. 46, column 13) it appears to be much thinner. The pillow lavas of Troodos were extruded in two phases, separated by an unconformity (Gass, 1968). The deep-water sediments and pyroclastic rocks that invariably cap the ophiolite suites have a thickness in the range of 500 meters. In some cases, e.g. Canyon Mountain Complex, this cap may be missing due to faulting or erosion.

Practically all the ophiolites under review are cumulates, although pervasive alteration and shearing may obscure these features. They invariably lie immediately beneath the gabbro unit. In some of the Newfoundland ophiolites, such as Bay of Islands (Williams and Malpas, 1972) and Betts Cove, dyke breccia is intimately associated with the sheeted dykes and forms an integral part of the ophiolite suite. Similar breccia may be present elsewhere but has been hitherto unrecognized. Significant comparisons based on petrography, mineralogy, and chemistry were made in the previous two chapters.

H. OPHIOLITIC ULTRAMAFIC ROCKS IN RELATION TO MANTLE MATERIAL

The mantle material which underlies oceanic crust is interpreted to be of ultramafic composition. In the context of information available from dredged samples and from ophiolites, this mantle can be divided into the following three types:

(1) The primitive mantle that has not undergone any "depletion"

through partial melting. The composition of this material may be typified by the lherzolitic inclusions found in basalts. Some spinel-bearing garnet peridotites occurring in the ophiolites of Morocco (Milliard, 1959) are stated to have originated at a pressure of over 10 kilobars (O'Hara, 1963) which would correspond to a depth of about 30 kilometers. Since this depth is too far below the uppermost parts of the mantle, Dewey and Bird (1971) have suggested that such high-pressure assemblages may represent primitive mantle inclusions rapidly carried up and frozen into the "depleted" cap.

(2) The "depleted" mantle from which a basaltic magma has been extracted to give rise to the cumulate ultramafic and other members of the ophiolites/oceanic crust. This type is generally believed to consist of dunite and harzburgite and to lack layering or cumulate textures. Stratigraphically it would lie immediately below the cumulate ultramafic member (Fig. 45). It is known to occur in the ophiolites of Papua (Davies, 1971),Troodos (Moores and Vine, 1971; Gass and Smewing, 1973), and Bay of Islands (Malpas, 1973). There is little or no evidence for the occurrence of "depleted" mantle in the Betts Cove sequence. However, some parts of the highly serpentinized sequence exposed along the Burtons Pond - Axe Pond join lack evidence of layering and may be of this type.

(3) The cumulate ultramafic rocks formed through the differentiation of basaltic magma extracted from the mantle at depth. These may be interlayered with gabbroic rocks towards the top. They are formed on top of the "depleted" upper mantle and invariably constitute the basal part of ophiolite suites.

The break known as the Moho can actually be one of the two divisions depending on the criteria for its recognition (Malpas, 1973): (i) The seismic Moho that marks the contact between gabbro and the underlying cumulate ultramafic rocks. (ii) The petrological Moho that occurs at the contact between the cumulate ultramafic member and the underlying "depleted" ultramafic material. In Betts Cove, as only cumulate ultramafic rocks are present, only seismic Moho is recognized. In other ophiolites such as Papua, Troodos, and Bay of Islands both seismic and petrological Mohos are recognized.

I. OPHIOLITIC AND RELATED ISLAND ARC TYPE ROCKS IN NEWFOUNDLAND

Practically all the ultramafic-mafic complexes in the Appalachian-Caledonian belt are now considered to be ophiolites, which, in turn, are interpreted as the remnants of a Proto-Atlantic Ocean (Bird *et al.*, 1971; Church, 1972). These ophiolites, numbering about 20, occur as disjunct isolated units throughout this mountain belt. In Newfoundland they occur in four main regions (Fig. 47). (i) The transported Bay of Islands and Hare Bay ophiolites on the western platform. (ii) The narrow linear Baie Verte-Mings Bight ophiolite belt in the Burlington Peninsula. (iii) The Betts Cove-Pilleys Island-Moretons Harbour ophiolitic and related island arc type terrain (Notre Dame Bay Supergroup) surrounding the Notre Dame Bay waters. (iv) The isolated, generally linear, ultramafic-mafic complexes in, and to the southwest of Gander River region.

The columnar sections in Figure 46 depict the correlation of the internal stratigraphy of the Snooks Arm Group and that of the other ophiolites of Newfoundland. A comparative summary of each of

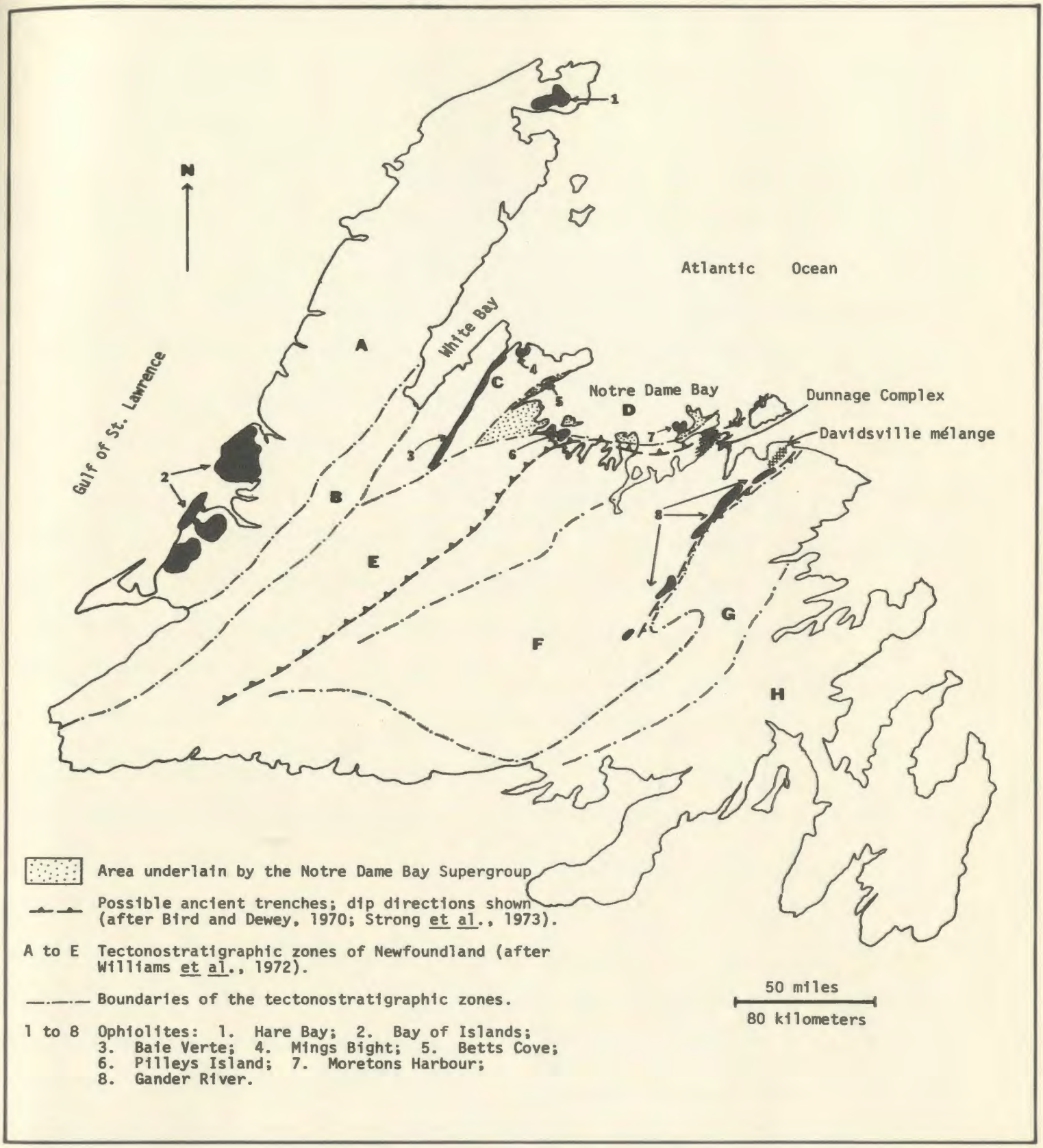


Figure 47: Major tectonic elements of Newfoundland.

these suites is presented below.

I.1 Bay of Islands and Hare Bay Ophiolites

In the Bay of Islands area (Smith, 1958) the ophiolites occur in four separate transported massifs. In at least two of these complete ophiolite suites are developed. The base is marked by a thrust contact. It differs from other Newfoundland ophiolites in having a high-temperature metamorphic aureole at the base of the ultramafic member. Both ultramafic and gabbro members are much thicker than the respective parts of Betts Cove ophiolite. The Bay of Islands ultramafic rocks also possess high-pressure mineral assemblages (Church and Stevens, 1971) and well developed tectonites. The top part of the ultramafic member and the overlying gabbro are layered and show cumulate textures. The sheeted dyke member (Williams and Malpas, 1972) contains a much higher proportion of dyke breccia than present in Betts Cove. The dykes have easterly to southeasterly trends. The dykes grade into pillow lavas which, in turn, are capped by deep-sea sediments.

Two ultramafic bodies with faulted base occur in the Hare Bay area. They have no gabbro, dykes, pillow lavas or sediments on top. From tectonostratigraphic relations they are also interpreted as a part of the transported ophiolite suite (Williams, 1971).

The west Newfoundland ophiolites are also of Lower Ordovician age (Williams, 1971). These ophiolites were transported from the east by gravity sliding (Rodgers and Neale, 1963). Dewey and Bird (1971) have suggested that the provenance of these ophiolites lies in the White Bay area. However, recent finding of a gneissic basement in that region by M. deWit (personal communication, 1972) makes this

suggestion untenable. The other possible sources are the Baie Verte or the Betts Cove/Notre Dame Bay ophiolitic terrain.

I.2 Baie Verte - Mings Bight Ophiolites

A narrow, linear, and almost continuous ophiolite belt extends between Baie Verte and the Trans Canada Highway near Sheffield Lake. It has fault contacts at the base (west) as well as at the top (east) (Neale and Kennedy, 1967). It is greatly dissected by tectonic slides, and the internal sequence is very complex. Except possibly sheeted dykes, all other members of a complete ophiolite suite are present (Fig. 46). Occurrence of a conglomerate bed between gabbro and pillow lava (W. S. F. Kidd in Dewey and Bird, 1971, p. 3199) represents an unexplained feature which is not common in other ophiolites.

The Mings Bight ophiolite, probably a disjunct part of the main Baie Verte belt, comprises all members including sheeted dykes. The internal stratigraphy has been highly modified by later faults. (R. Norman, personal communication, 1972). The Mings Bight ophiolite is much like the Betts Cove ophiolite except that it includes an apparently thicker gabbro member.

Church and Stevens (1971) have suggested that the Baie Verte ophiolite may represent a faulted and steeply folded extension of the Betts Cove ophiolite sheet. The author, however, considers the steep suture-like aspect of the Baie Verte ophiolite a very significant feature which, in conjunction with its tight synclinal structure, gives the impression that this belt was formed by squeezing from the two opposite sides (cf. Kennedy, 1973). Thus, alternative suggestion (Dewey and Bird, 1971) that the Baie Verte ophiolite originated in a

marginal basin at its present site, the closure of which gave rise to the Baie Verte ophiolite suture, appears more reasonable than the suggestion of Church and Stevens.

On the basis of similar rock types and structural history the Baie Verte (Group) ophiolite was originally interpreted as the same age as the Snooks Arm Group (Neale and Kennedy, 1967). From a detailed structural study, however, Kennedy and Phillips (1971) have since suggested that the ultramafic rocks of the Baie Verte belt are pre-Ordovician in age. This is further discussed later in this chapter.

I.3 Ophiolites of the Notre Dame Bay Supergroup

The term "Notre Dame Bay Supergroup" is proposed here to include all the Ordovician ophiolites and related island arc type rocks of the Notre Dame Bay region* (Fig. 47). The boundary of this supergroup is marked, for the most part, by the easterly trending Lukes Arm Fault which separates it from the predominantly pyroclastic and sedimentary rocks of Ordovician to Silurian age in the south. To the west it is bounded by the fault at the base of the Snooks Arm Group. This supergroup includes most of zone "D" rocks of Williams *et al.*, (1972) and also nearly all of the "Lush's Bight terrain" of Horne and Helwig (1969). The Notre Dame Bay Supergroup, as proposed here, includes the following groups.

---

*Strong and Payne (1973) have referred to the Notre Dame Bay ophiolites as "Lush's Bight Supergroup". The present author prefers the term "Notre Dame Bay Supergroup" for the following two reasons: (i) "Lush's Bight" has been used in a variety of contexts in the geological literature and is likely to cause confusion. (ii) Notre Dame Bay Supergroup is geographically inclusive and self-explanatory term.

(1) Snooks Arm Group

(2) Lush's Bight Group

(3) Western Arm Group

(4) Catcher Pond Group

(5) Cutwell Group

(6) Moretons Harbour Group

(7) (?) Nippers Harbour Group

The reason for incorporating these groups into a supergroup lies in the recent findings that all of these are of Ordovician age and represent remnant oceanic crust or related island arcs. It can be shown that in the Notre Dame Bay region a transition from main oceanic crust (Snooks Arm and Lush's Bight Groups), through mixed oceanic-island arc (? Moretons Harbour Group) to a complete island arc (Cutwell and Catcher Pond Groups) type assemblages exists.

Figure 46 shows the cross-sections of various groups within the Notre Dame Bay Supergroup. A section of the Catcher Pond Group is not included as the detailed succession within this group is not known. A brief account of these groups follows.

(A) The Lush's Bight Group (MacLean, 1947) consists of 4570 meters of pillowed meta-basalt at the base (Little Bay Head Section) which is conformably overlain by the Western Arm Group. It has however been pointed out (Neale and Nash, 1963) that there may have been more repetitions of beds by tight folding than previously recognized. Some of the Lush's Bight pillow lavas have been identified as low-potash oceanic tholeiites (Smitheringale, 1972; Strong, 1973b). In Pilleys Island the Lush's Bight Group is known to consist of

pegmatitic gabbro (e.g. the Brighton Complex), sheeted dykes, pillow lavas and pyroclastic rocks, although the stratigraphic order has not been preserved as the contacts are marked by vertical to subvertical faults (Strong, 1972). The sheeted dykes have a north to northwest trend.

(B) The Western Arm Group has the same age (MacLean, 1947) and lithology (Marten, 1971) as the Snooks Arm Group, although it lacks the ophiolite part corresponding to the Betts Cove Ophiolite. A probable correlation between these two groups, which apparently occupied the two limbs of a fault-disrupted syncline, is shown in Table 3 and in Figure 46.

The Pillow Lava Member of the Betts Cove Ophiolite is correlated here with the Lush's Bight pillow lavas mainly because of its conformable contact with the overlying Western Arm Group and its oceanic tholeiitic composition. No counterpart of the Round Harbour Basalt exists in the Western Arm Group. It should be noted that the Lush's Bight pillow lava pile is much thicker than the Pillow Lava Member of Betts Cove Ophiolite. There are two possible explanations for this thickness change: (i) The Lush's Bight pillow lava unit gradually thins towards the north, as seems to have happened from the Western Arm Section in the south to the Harry's Harbour Section in the north (Table 3), so that the latter has a thickness much closer to that of the Snooks Arm Group. (ii) The apparent thickness may have been exaggerated due to faulting (Strong, 1972).

(C) The Catcher Pond Group (Neale and Nash, 1963) consists of rhyolitic flows, pillowed andesites, and amygdaloidal

TABLE 3

STRATIGRAPHIC CORRELATION BETWEEN SNOOKS ARM AND WESTERN ARM GROUPS

Snooks Arm Group		Notation used in Figure 46	Western Arm Group (Marten, 1971)		
Formation Name	Thickness (Meters)		Formation Name	Thickness (Meters)	
				Harry's Harbour	Western Arm Section
Round Harbour Basalt	1000	E	?	?	?
Balsam Bud Cove Formation	750	D	Western Head Agglomerate	300	1000
			Welsh Cove Tuff	330	450
Venams Bight Basalt	500	C	Bill Hill Basalt	500	700
Bobby Cove Formation	500	B	Skeleton Pond Tuff	850	950
Pillow Lava Member of Betts Cove Ophiolite	1200	A	Pillow lavas of Lush's Bight Group	?	?

basalts with minor pyroclastic rocks. Trilobites of Lower Ordovician age have been described from the thin beds of limestone that are interbedded with the lavas (Dean, 1970). Although no chemical data on these rocks is yet available, from their lithological similarity with the Cutwell Group (described below) they seem to represent an island arc type assemblage.

(D) The Cutwell Group (Espenshade, 1937), exposed in the Long Island area, consists of an intrusive complex (gabbro and diabase dykes) at the base, overlain by thick agglomerate, sediment, and pillow lava units (Kean, 1973). The total thickness (about 5 kilometers, as re-defined by Kean), stratigraphy, and rock types are similar to the volcanic and sedimentary/pyroclastic rocks of the Snooks Arm Group. The pillow lava unit in the upper part of the Cutwell Group is andesitic as compared to the basaltic pillows in the lower levels. From the stratigraphy and petrochemistry this group has been interpreted as a remnant island arc (Kean, 1973). It contains fossils of lower Middle Ordovician age (Williams, 1962) and is the youngest group within the Notre Dame Bay Supergroup.

(E) In the Moretons Harbour area, the eastern part of Notre Dame Bay, an 8-kilometer pillow lava-sheeted dyke-tuff succession has been designated as the Moretons Harbour Group (Strong and Payne, 1973), formerly included in the Lush's Bight Group (Williams, 1964b). Thin siliceous sedimentary units are common in the two pillow lava formations at the bottom. Pillow lavas occur as screens within sheeted dykes. The dykes are overlain successively by a dominantly tuffaceous and a basaltic pillow lava formation. Ultramafic rocks and gabbro are

not seen. On the basis of an exceedingly thick (6 kilometers) volcanic pile, non-theoliitic chemistry of the sheeted dykes, and the presence of volcaniclastic sediments, Strong and Payne (1973) have concluded that the Moretons Harbour Group represents an ancient island-arc assemblage possibly with some transition towards oceanic crust at the base.

(F) The Nippers Harbour Group, exposed to the southwest of the Snooks Arm Group, is part of an ophiolite suite consisting chiefly of gabbro and diabase dykes with isolated pillow lava outcrops (Fig. 46). It is faulted against the Snooks Arm Group. No concrete evidence is yet available to indicate whether it is pre-Ordovician or Ordovician in age. Interpretations of some field relationships (p. 27) suggest a pre-Ordovician age, which, although not conclusive, is tentatively accepted in this thesis. This and other aspects of the Nippers Harbour Group have been described in Chapter II.

I.4 Gander River Ophiolites

A narrow, discontinuous belt of ultramafic and gabbroic rocks with associated mafic volcanic rocks runs parallel to the Gander River along the eastern margin of Zone F, and can be traced 150 kilometers southwestward. The ultramafic rocks, mainly pyroxenites and their serpentinized equivalents, occur as isolated bodies within Ordovician volcanic rocks and dark slates. They trend parallel to the regional strike and have been assigned an Ordovician age (Jenness, 1958). The nature of the contact between ultramafic and gabbroic rocks is not known generally due to lack of exposures, but sills and dykes of pyroxenites, as in Betts Cove, are known to cut both serpentinites and

gabbro (Jenness, 1958). Lack of detailed information does not permit any meaningful comparisons with the Snooks Arm Group. However, the author suggests that these ultramafic-mafic bodies represent an ophiolite which, due to extensive thrusting and faulting, occurs as disjunct and isolated units.

J. TECTONIC EVOLUTION AND EMPLACEMENT OF THE SNOOKS ARM GROUP

J.1 Introduction and Review of Ideas

J. Tuzo Wilson (1966) suggested that the site of the present Appalachian-Caledonian mountain belt was originally occupied by a "Proto-Atlantic" ocean during Early Paleozoic time. This ocean closed during Middle to Late Paleozoic through subduction along its western margin, giving rise to an island arc system and finally producing a fold mountain belt. Since Wilson's suggestion, various models have been proposed for the evolution of this orogen. From the occurrence of sheeted dykes in Newfoundland, it is generally agreed that the Proto-Atlantic was formed by the process of sea-floor spreading and also that nearly all of the ultramafic-mafic complexes (ophiolites) in the Appalachian orogen may represent crust of that ocean. However, differences of opinion do exist regarding the age(s), place of origin, and mode of emplacement of the ophiolites and therefore at this stage no particular model can be taken as a final one. The author therefore summarizes pertinent ideas and then presents a model which he considers most plausible, especially with reference to the emplacement of the Snooks Arm Group.

Bird and Dewey's (1970) model, perhaps the most comprehensive proposed so far, envisages a late Precambrian to Ordovician expansion,

followed by Ordovician through Devonian contraction of the Proto-Atlantic ocean. They have postulated a northwest* dipping subduction zone, which passes approximately through the trend of the Luke Arm Fault in Notre Dame Bay, Red Indian Lake in central Newfoundland, and Rose Blanche-LaPoile Bay on the south coast (Fig. 47). Their evidence for the remnant trench is the Dunnage Complex in Notre Dame Bay (Kay, 1967; Horne, 1969). This complex is a mélange of breccias and conglomerates with gigantic slide blocks. It contains limestone lenses with Middle Cambrian fauna. Bird and Dewey (1970) maintain that the Proto-Atlantic began to open in the Late Precambrian through distention of a North American/African continent. During Late Cambrian-Early Ordovician plate-loss was initiated and a northwest dipping subduction zone (Dunnage Complex) began to form. This closing of the Proto-Atlantic continued until Middle Devonian, when the oceanic plate was consumed beneath the island arc and a collision between opposing continental margins occurred. This collision resulted in the main (Acadian) deformation in the Newfoundland Appalachians.

In a later paper (Dewey and Bird, 1971), these authors have suggested that the Baie Verte and Betts Cove ophiolites represent marginal basins, separated from the main Proto-Atlantic by an island arc or a microcontinent. They have suggested that remnants of pre-Ordovician oceanic crust are represented by the Nippers Harbour Group and by the Little Bay Head Section of the Lushs Bight Group.

In contrast, Church and Stevens (1971) have suggested that the

---

*All the directions in this chapter are referred to the present, that is "northwest" for Devonian time means present geographical northwest which may be different from the northwest of that time.

ophiolites could have been emplaced at the time of their origin, if the ocean-ridge was close to the continental margin. Although they suggest various alternative hypotheses regarding the origin and emplacement of Newfoundland ophiolites, they favour the notion that all these ophiolites once formed a single sheet that was subsequently folded during Acadian orogeny. They propose a southeast dipping subduction zone for the closing of the Proto-Atlantic. This is substantiated by, and apparently based upon, the occurrence of island arc type volcanic rocks that lie to the south of the Caledonian axis in Britain (Fitton and Hughes, 1970). Church and Stevens propose an Early Ordovician emplacement of the Betts Cove ophiolite because the graywacke in the Snooks Arm Group contains ultramafic detritus which, according to these authors, may have been derived from the basal ultramafic rocks of the same group that were already thrust or obducted on to land. Arguments against the derivation of the ultramafic detritus from the ultramafic member of the same (Snooks Arm) group are the following: (1) The map- and stratigraphic- distance between the ultramafic-detritus bearing graywacke and the basal ultramafic member of the Snooks Arm Group is about 3 kilometers and 2 kilometers respectively. These distances, in the author's view, are too small to produce a situation where the basal ultramafic member could form an on-land thrust sheet and at the same time the deep-water sediments could be deposited at the bottom of the basin.

Kennedy (1973) has recently proposed a model which is based on the divergent facing directions of F_2 folds in the Fleur de Lys Supergroup. He suggests that a small ocean basin opened in pre-Ordovician

time within the Fleur de Lys Supergroup above a westerly dipping subduction zone. The pre-Ordovician oceanic crust is represented by the ultramafic rocks of the Baie Verte - Mings Bight belt and the related island arc assemblage by the Pacquet Harbour and Cape St. John Groups. Subsequent pre-Ordovician closure of this basin and its two-way obduction on to the Fleur de Lys continental rocks led to the formation of the divergently facing F_2 folds on two sides of the Baie Verte ultramafic belt.

A new phase of ocean generation began, according to Kennedy's model, in the Lower Ordovician after the closure and obduction of the older basin. On the Burlington Peninsula this younger oceanic crust is represented by the Snooks Arm Group and the pillow lavas, tuffs and sediments of the Baie Verte Group that are exposed adjacent to the eastern boundary of the Baie Verte ultramafic belt. He thus separates the ultramafic rocks from the pillow lava-tuff-sediment assemblage along the Baie Verte belt and assigns them pre-Ordovician and Ordovician ages respectively. The latter, he contends, got to their present position by obduction from the Betts Cove area during a Devonian closure of the main ocean to the southeast.

Kennedy's model satisfactorily explains the origin of the divergently facing folds in the Fleur de Lys Supergroup. However, his proposed distinction of the ultramafic rocks from the pillow lava-tuff-sediment assemblage, both being so intimately associated, striking parallel to one another and together forming an apparent complete ophiolite suite, should be viewed with caution. He also fails to explain why northward-facing folds did not develop in the Fleur de Lys

rocks when the Ordovician oceanic plate was obducted from Betts Cove in the south to Baie Verte in the north.

On the basis of chemical data from granitod rocks and distribution of mineral deposits Strong *et al.*, (1973) have suggested an east dipping subduction zone for the closure of the Proto-Atlantic. They have found a continuous increase in the K_2O content in the rocks of Zone F to Zone H. Compilation of the distribution of economic minerals in eastern Newfoundland shows the following pattern (Strong, 1973a): Zn-Pb-Cu-Ag-Au in the Central Mobile Belt (Zones E, F); Cu-Mo in Gander Lake belt (Zone G); F-Ba-Zn-Ph-(?)Sn at St. Lawrence (Zone H). They consider this pattern broadly similar to that of western North and South America (Sillitoe, 1972a) and, by analogy, propose an east dipping subduction zone in Newfoundland.

J.2 The Proposed Model

The model adopted in this thesis to depict the tectonic evolution and emplacement of the Snooks Arm Group and other ophiolitic rocks of Newfoundland is a synthesis of, and stems from, several of the proposals mentioned above. The author, however, wishes to emphasize that this model is tentative and may be profoundly modified as more information becomes available.

J.2.1 Subduction Zones

The present proposal requires that the Proto-Atlantic closed through subduction at both the eastern (or southeastern) and the western (or northwestern) margins. Evidence for the east dipping subduction zone includes the following features:

(i) In the British Caledonides, which are the north-eastward

extension of the Appalachians, island arc type volcanic rocks occur to the southeast (Fitton and Hughes, 1970).

(ii) Strong *et al.*, (1973) present evidence of increase in K_2O content from the Gander Lake area eastwards, and, together with the zoned mineral belts of eastern Newfoundland, this may imply an east dipping subduction zone.

(iii) The patchy ultramafic/mafic rocks (Williams, 1967b) extending southwestward from Gander River form a remarkably narrow linear belt that generally occurs at the contact of Zones F and G. From this outcrop pattern and geological setting, the author interprets them as shredded oceanic crust fragments that were scraped from a down-going lithospheric plate. These are not the large sheet-like plates such as the ophiolites of western Newfoundland and Notre Dame Bay, but represent small fragments analogous to the Franciscan ophiolites of California (Bailey *et al.*, 1970; Dewey and Bird, 1970). The position of the Gander River ophiolites can be explained if the lithospheric plate was consumed along an east dipping subduction zone.

(iv) Mélange zones separating the Middle Ordovician sedimentary-volcanic terrain from the polyphase deformed pre-Middle Ordovician metasedimentary terrain have been reported from the Gander River area (Kennedy and McGonigal, 1972). The mélange contains deformed blocks of metasediments that vary in size from a few centimeters to tens of meters. The author interprets this mélange as a Paleozoic trench deposit comparable in some respects to the Dunnage Complex in northeast Newfoundland.

A west dipping subduction zone is evidenced by the following:

(i) The Dunnage Complex (mélange) shows some evidence of being a fossil trench deposit (Bird and Dewey, 1970).

(ii) The position of the island arc assemblages in Moretons Harbour (Strong and Payne, 1973) and Long Island (Kean, 1973), indicates, and is consistent with, a west dipping subduction zone, if the Dunnage Complex is a trench deposit.

(iii) The Buchans area, to the west of Red Indian Lake through which the postulated remnant trench passes, is underlain by island arc type calc-alkaline volcanic rocks (J. Thurlow, personal communication, 1973) implying a west dipping subduction zone.

J.2.2 Tectonic Evolution

The tectonic evolution of the Proto-Atlantic Ocean and various stages that led to the emplacement of the Snooks Arm Group are shown in Figure 48. Each stage (a to e) is elaborated below.

(a) In Late Precambrian a composite American/African continent began to distend, initiating the opening of the Proto-Atlantic Ocean. At a later stage, subduction of this oceanic plate began along a west dipping subduction zone. From late Precambrian through Middle Cambrian the Fleur de Lys Supergroup was deposited. The basal part of this supergroup consisted of continental slope and rise sediments; on the higher stratigraphic levels it was made up of the island arc type Beaver Cove, Pacquet Harbour, and Cape St. John Groups that were generated on top of the down-going lithospheric plate. This island arc assemblage partly rested on the marginal oceanic crust probably represented by the Nippers Harbour Group and the Reddits Cove Gabbro. The Fleur de Lys

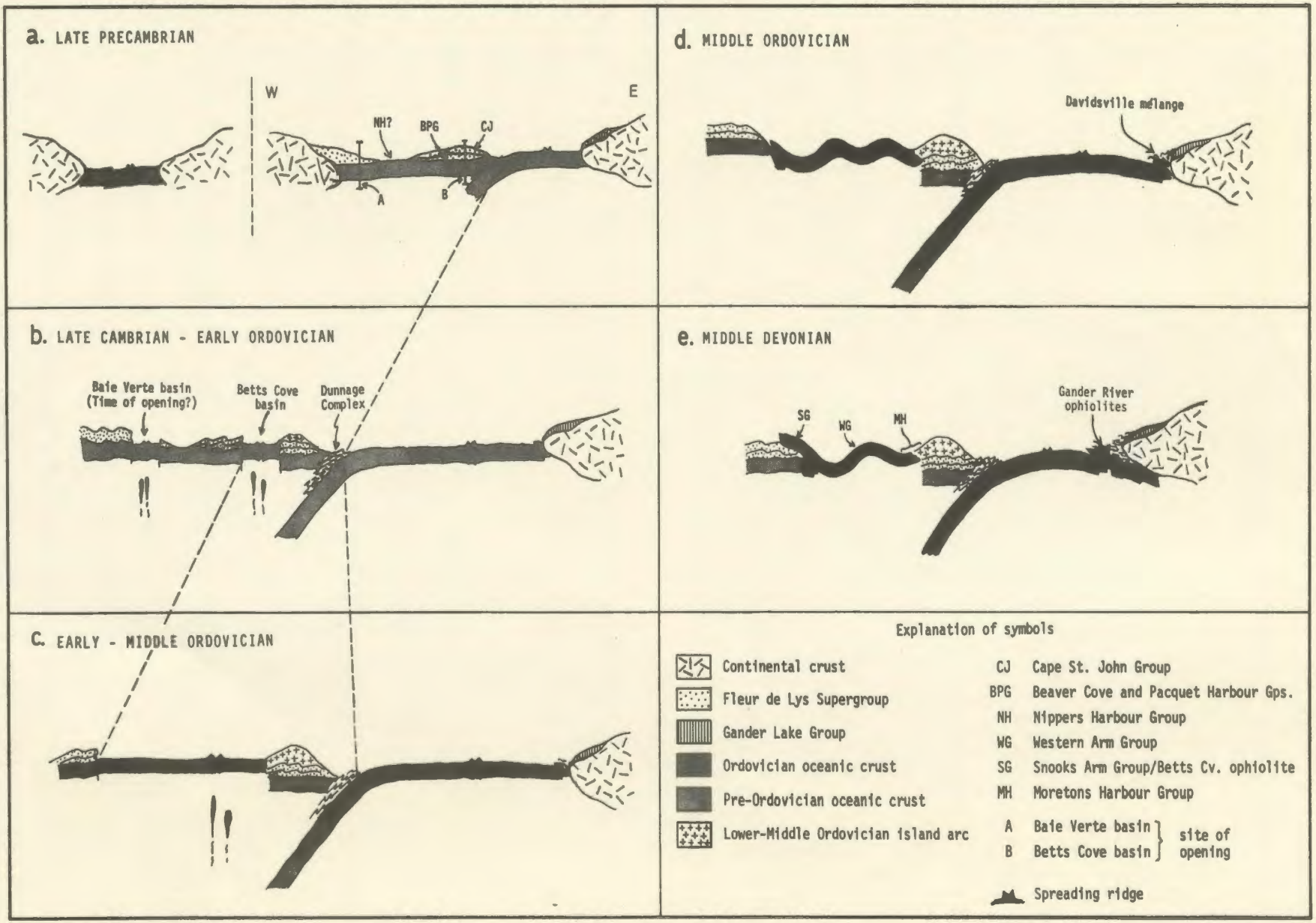


Figure 48: Schematic sections illustrating the tectonic evolution of the Snooks Arm Group and some other ophiolites of Newfoundland.

rocks and the associated oceanic crust were deformed in Upper Cambrian time (Williams *et al.*, 1972).

On the eastern side, the Gander Lake Group, chiefly a continental rise clastic sequence, was deposited on Grenvillian gneissic basement and was deformed in pre-Middle Ordovician time (Williams *et al.*, 1972).

(b) During latest Cambrian and Early Ordovician the opening of the Betts Cove and perhaps Baie Verte marginal basins began. It is envisaged that the rifting of the Fleur de Lys Supergroup caused the opening of the Betts Cove basin so that a part of it moved eastward as the opening progressed. The (presumed) trench deposits of the Dunnage Complex began to form at this stage. Adjacent to this trench island arcs were built perhaps on top of the rifted Fleur de Lys rocks. The pre-Ordovician oceanic crust was thus first covered by the Beaver Cove-Pacquet Harbour-Cape St. John Groups and then, on the postulated rifted fragment in the east, by the Ordovician island arc assemblages. The pre-Ordovician part of the oceanic plate was gradually consumed in the trench and was replaced by successively younger ones as the plate-accretion at the ridge continued. The magma for the generation of the Betts Cove basin was derived from the melting of the down-going main lithospheric plate.

(c) From Early to Middle Ordovician time the Betts Cove marginal basin became wider through sea-floor spreading. By Middle Ordovician time it had attained its maximum width and included the present-day ophiolitic rocks of Notre Dame Bay. Island arc assemblages such as those of Moretons Harbour, Cutwell, Western Arm, and upper levels of Snooks Arm Group were deposited. The postulated subduction zone dipping to the the east beneath the

Gander Lake Group, probably had its inception in Middle Ordovician time, as indicated by the Middle Ordovician sediments and mélange zone containing giant blocks of polydeformed metasediments (Kennedy and McGonigal, 1972).

(d) From Middle Ordovician to Early Devonian, the marginal basin closed in stages. The Bay of Islands and Hare Bay allochthons, whose movement began in the latter part of stage (c), were emplaced in the Middle Ordovician (Rogers, 1965; Stevens, 1970). The source area for these allochthons may be either the Baie Verte or the Betts Cove-Notre Dame Bay region. The Gander River ophiolites, postulated to have been scraped from a down-going plate, were emplaced mainly at this stage.

(e) In Middle Devonian time, the Snooks Arm Group was finally obducted onto the Fleur de Lys continental margin. The folding and deformation of the Notre Dame Bay ophiolitic rocks culminated at this stage giving rise to the main syncline of which Snooks Arm and Western Arm Groups form the two opposite limbs. This collision brought the Gander Lake metasedimentary belt closer to the Fleur de Lys terrain. This probably marks the time of the Acadian orogeny, pervasive through most of the Newfoundland Appalachians.

Pre-Ordovician oceanic crust had been completely consumed in the trenches by this time. The present Zone F rocks in east-central Newfoundland may be underlain by the Ordovician oceanic crust.

CHAPTER VI

ECONOMIC GEOLOGY

A. INTRODUCTION

Betts Cove and Tilt Cove are the two localities in the area underlain by the Snooks Arm Group where copper has been mined in the past. The Betts Cove deposits were discovered in 1860 and mined between 1875 and 1885, yielding 130,000 tons of hand-picked ore averaging about 10 percent copper along with 2450 tons of pyrite (Douglas *et al.*, 1940). The Tilt Cove deposits were discovered in 1857 and were mined during two periods from 1864 to 1917 and from 1957 to 1967, yielding several million tons of copper ore.

Apart from Betts Cove and Tilt Cove areas, important pyrite and/or copper showings also occur within the pillow lavas around the eastern and central parts of Long Pond and near the pillow lava/sediment contact, on the north shore of Snooks Arm (Fig. 2). The present study was largely concentrated on the Betts Cove deposits and therefore the ore genesis model presented later in this chapter is based chiefly on the information from the Betts Cove occurrences.

B. THE TILT COVE AREA

All the open-cut workings of the Tilt Cove area occur within the Pillow Lava Member of the ophiolite suite (Fig. 2). Several other small occurrences are also confined to the same pillow lavas. Pillow breccia in this area seems to have been particularly susceptible to sulfide mineralization. This is exhibited by the heavily pyritized

rusty-looking pillow breccia exposed along the cliff face between Winser Lake and the Government wharf.

The economic geology of this area has been described in varying degrees of detail by Snelgrove (1931), Douglas *et al.*, (1940), Donoghue *et al.*, (1959) and Baird (1959) from which the following account was prepared. The host rocks for the ore are chloritized, sheared, and brecciated pillow lavas and agglomerates. The ore bodies are of two types: (i) massive, fine-grained, steeply-dipping pyrite bodies with chalcopyrite and (ii) diffused stockworks of stringers, veins, and irregular clusters of pyrite and chalcopyrite. The chief minerals are pyrite and chalcopyrite with minor magnetite, sphalerite, pyrrhotite, native silver, and gold. Specularite and limonite occur in the vicinity of strong post-ore slips within the massive sulfide zones. Nickel mineralization at the contact of talcose ultramafic rocks and mafic pillows has been reported by Papezik (1964).

Polished sections of Tilt Cove ore studied by the author show textures indicative of extensive replacement. The following paragenesis is suggested: magnetite-pyrite-pyrrhotite-chalcopyrite. Chalcopyrite replaces nearly all the earlier minerals.

Occurrence of Tilt Cove ore bodies in the vicinity of intrusive silicic rocks and along fault zones, led earlier workers to suggest an epigenetic origin whereby the ore fluids were derived from the intrusive rocks and the fault zones acted as localizers. The author contends that the ore bodies are syngenetic and, as discussed later in detail, their present occurrence along fault zones is related to post-mineralization events.

C. THE BETTS COVE AREA

C.1 Field Relationships

The locations of the sulfide deposits in the Betts Cove area is shown in Figure 2. Most important deposits occur in two regions: (i) at and around the abandoned Betts Cove mine and (ii) at Mount Misery. Significant deposits also occur near the southern end of Burtons Pond and at several places along the Fault Cove-Betts Cove mine join. It will be noticed from Figure 2 that the sulfides are concentrated along or near the contact of sheeted dykes and the overlying pillow lavas. Pyrite is sparsely disseminated throughout gabbro, sheeted dykes and pillow lavas; chalcopyrite and pyrite along with green stains of malachite are particularly conspicuous in sheeted dykes. The sulfides occur in all forms ranging from banded massive lenses to disseminated zones. The largest sulfide bodies occur within chloritized fault zones and ore samples from such areas are strongly foliated.

In the pillow lavas the maximum concentration of sulfides occurs in the spaces between pillows. The sulfide-rich interspaces are now foliated, producing an augen structure around the massive pillows (Fig. 49). The concentration of sulfides between pillows is taken as a primary volcanogenic feature which was further accentuated by deformation as the sulfide-bearing interspaces are more susceptible to shearing and remobilization than the interiors of the massive pillows. Poorly mineralized and heavily mineralized layers, about 50 centimeters thick, with sharp contacts, are seen at the Betts Cove mine face. This may also represent some primary depositional feature, now somewhat modified by shearing.

Figure 49: Mineralized pillow lava outcrop in which sulfide-rich matrix is strongly sheared and produces augen structure around the massive pillows. Betts Cove mine.

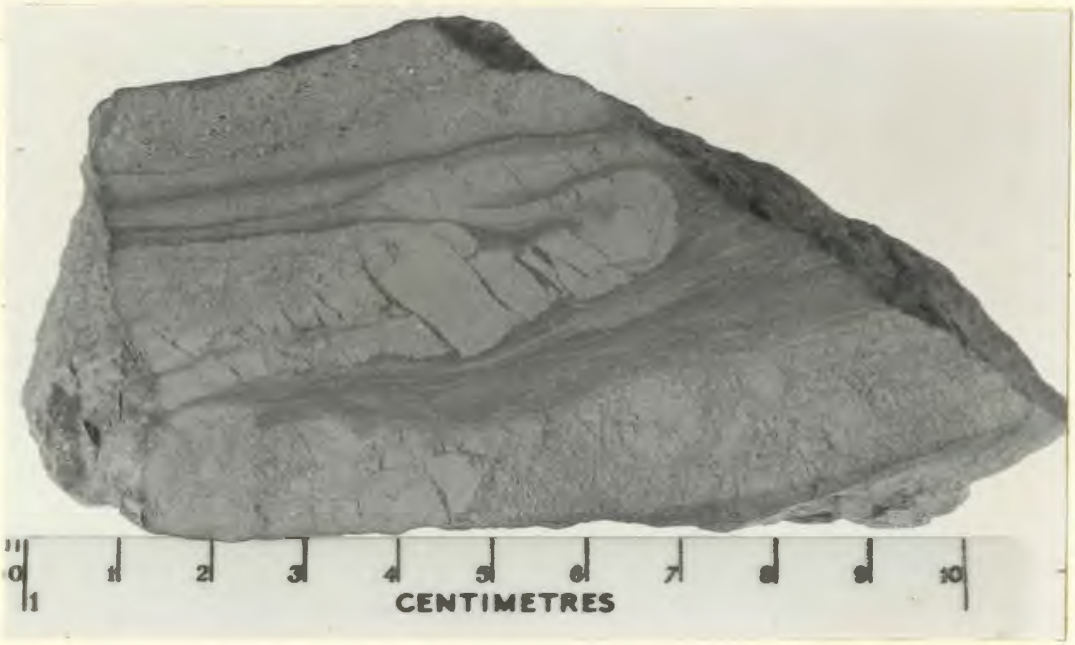


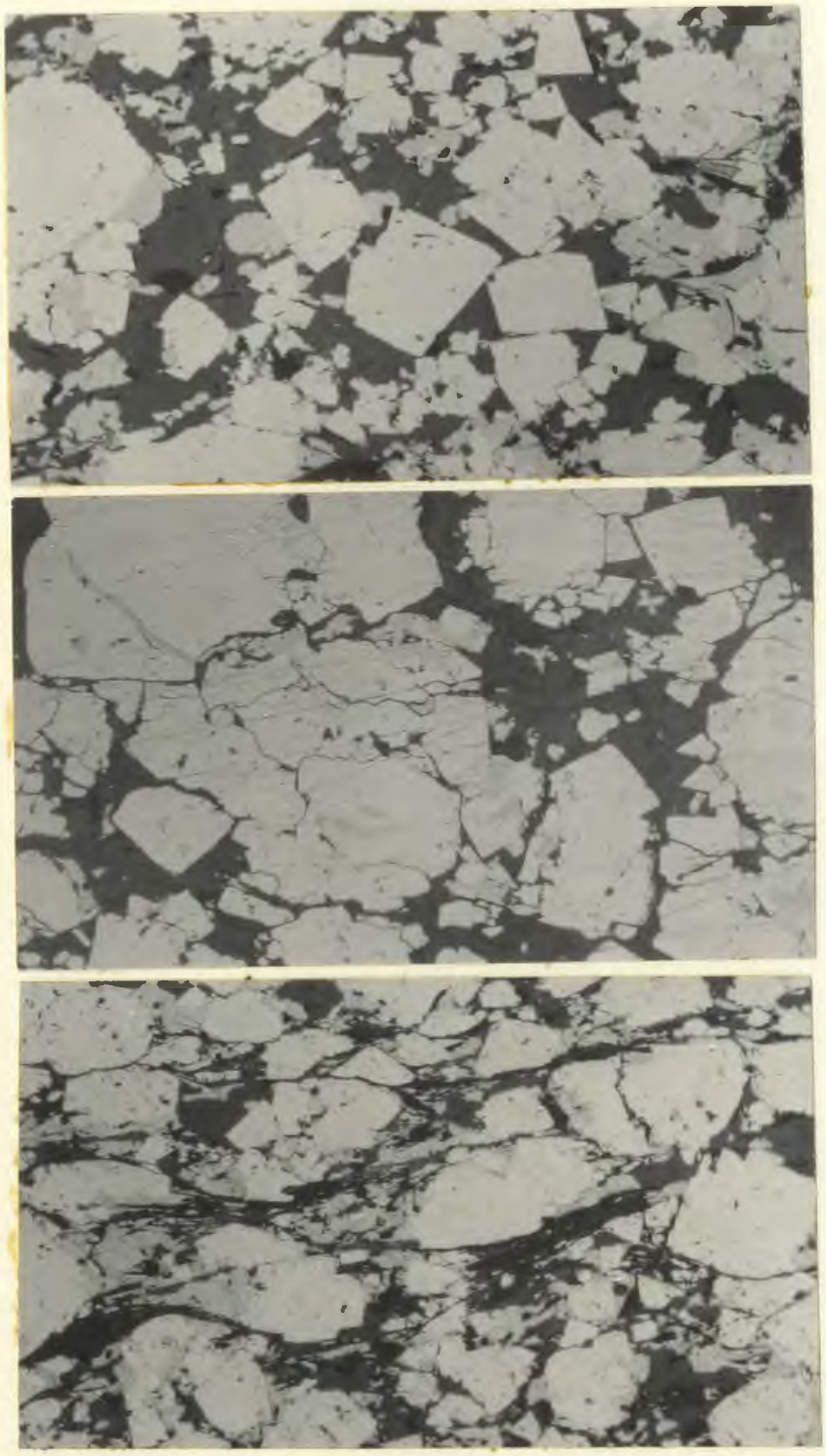
Figure 50: Sedimentary slump fold in a massive sulfide specimen. Light material is pyrite, dark is sphalerite. Betts Cove mine dump.

Massive and banded sulfides are best seen in the material from the mine dump, where numerous blocks show sedimentary lamination and slump folds which are indicative of initial sedimentary deposition of sulfides (Fig. 50). Such primary features were later modified by deformation and remobilization as suggested by veins and stringers of sulfides within the associated silicate minerals.

C.2 Mineralography

Microscopic studies of the Betts Cove ore show that pyrite is by far the most abundant sulfide mineral, followed by chalcopyrite and local concentrations of sphalerite; native iron (Deutsch and Rao, 1972) occurs elsewhere in the Betts Cove pillow lavas. Although there appears to be a definite paragenetic sequence, with pyrite being the earliest phase to crystallise generally as dispersed euhedral crystals in a sphalerite-chalcopyrite matrix (Fig. 51a)., pyrite, sphalerite and chalcopyrite form individual laminae with sharp contacts that are fairly consistent in thickness. In the case of pyrite, for instance, crystals show a constant size within any particular lamina. Chalcopyrite occurs as disseminated stringers and also interstitial to euhedral pyrite crystals, but most commonly fills fractures within highly brecciated pyrite grains (Fig. 51b). Sphalerite appears last in the sequence surrounding and filling fractures in both pyrite and chalcopyrite.

The hand specimens of undeformed rocks show more clearly that these sulfides are interbedded in alternating layers rich in one particular mineral (Fig. 50), indicating repeated precipitation of each phase. The paragenetic sequence described above is interpreted as



a.

b.

c.

Figure 51: Photomicrographs of Betts Cove ore samples. a: Primary euhedral pyrite crystals and associated chalcopyrite in a silicate matrix (X45). b: Fractured and granulated pyrite crystals with remobilized chalcopyrite "healing" the fractures (X85). c: Strongly foliated pyrite with minor chalcopyrite, in a silicate matrix (X45). All under reflected plane polarized light. The light material is pyrite, medium is chalcopyrite and the dark is silicate gangue. Note the gradation from undeformed (a), through moderately deformed (b), to highly deformed (c) sulfides.

reflecting a relative ease of remobilization, with brittle deformation of pyrite and more plastic deformation of chalcopyrite and sphalerite--an interpretation supported by the experiments of Gill (1969) and the observations of Suffel *et al.*, (1971). A complete gradation is observed from unfoliated to highly foliated ore (Fig. 51) on all scales from microscopic to outcrop size. Silicates, particularly quartz, associated with the sulfides are foliated in the same direction as the deformed sulfides.

D. GENESIS OF THE BETTS COVE-TILT COVE ORE DEPOSITS

Earlier workers have attributed the sulfide mineralization in the Betts Cove area to nearby dioritic-gabbroic (Snelgrove, 1931; Douglas *et al.*, 1940) and granitic (Baird, 1931) intrusions producing an epigenetic, hydrothermal mineralization localized along pre-existing faults. Since the gabbroic rocks in the Betts Cove ophiolite occur as "cold" screens within sheeted dykes (i.e. they did not intrude the dykes), they could not have been responsible for any epigenetic mineralization. Furthermore, the concentration of sulfides at the base of the Pillow Lava Member, presence of malachite along the contact of adjacent (sheeted) dykes, slump folds in sulfides, and the concentration of sulfides in the interstices between pillows, indicates a primary syngenetic origin of the sulfides.

It was argued earlier that the Betts Cove-Tilt Cove ophiolite represents an oceanic crust that was generated through plate-accretion at an Ordovician oceanic ridge. The author contends that the sulfides are comagmatic with the mafic portion of the ophiolite that hosts them.

It is suggested that the sea-floor spreading in Early Ordovician time produced the ophiolite suite which was accompanied in the very early stages by volcanic exhalative mineralization to produce typical massive sulfides underlain by disseminated "stockwork" mineralization (Fig. 52). The movement of the sulfide material was essentially along the planes defined by the subvertical sheeted dykes, and as the latter became sills/flows after reaching the surface, i.e. the path changed from near-vertical to subhorizontal, the sulfides probably could not reach beyond the top portion of the Sheeted Dyke Member. The "capping" formed by the subhorizontal pillow lavas on top of the sheeted dykes could have hampered the movement of sulfides towards the lavas and hence causing their concentration at the sheeted dyke/pillow lava contact zone. In a case like Tilt Cove, where sheeted dykes are either absent or poorly developed, the intrusive gabbroic rocks seem to have been converted to flows (pillow lavas) on the surface and hence there the sulfides were deposited in the lava flows.

The mineralization resulted primarily from syngenetic precipitation of sulfides from upwelling geothermal brines during hiatuses in volcanism in an oceanic ridge-type environment, such as that existing in the Red Sea today (c.f. Degens and Ross, 1969). The less important disseminated or "stockwork" mineralization is produced by replacement and cavity-filling in the underlying rocks during ascent of the volcanic exhalations.

Faulting and shearing at a later stage disrupted the original shapes and attitudes of the ore bodies and finally led to their remobilization and emplacement along these fault zones (Fig. 52c).

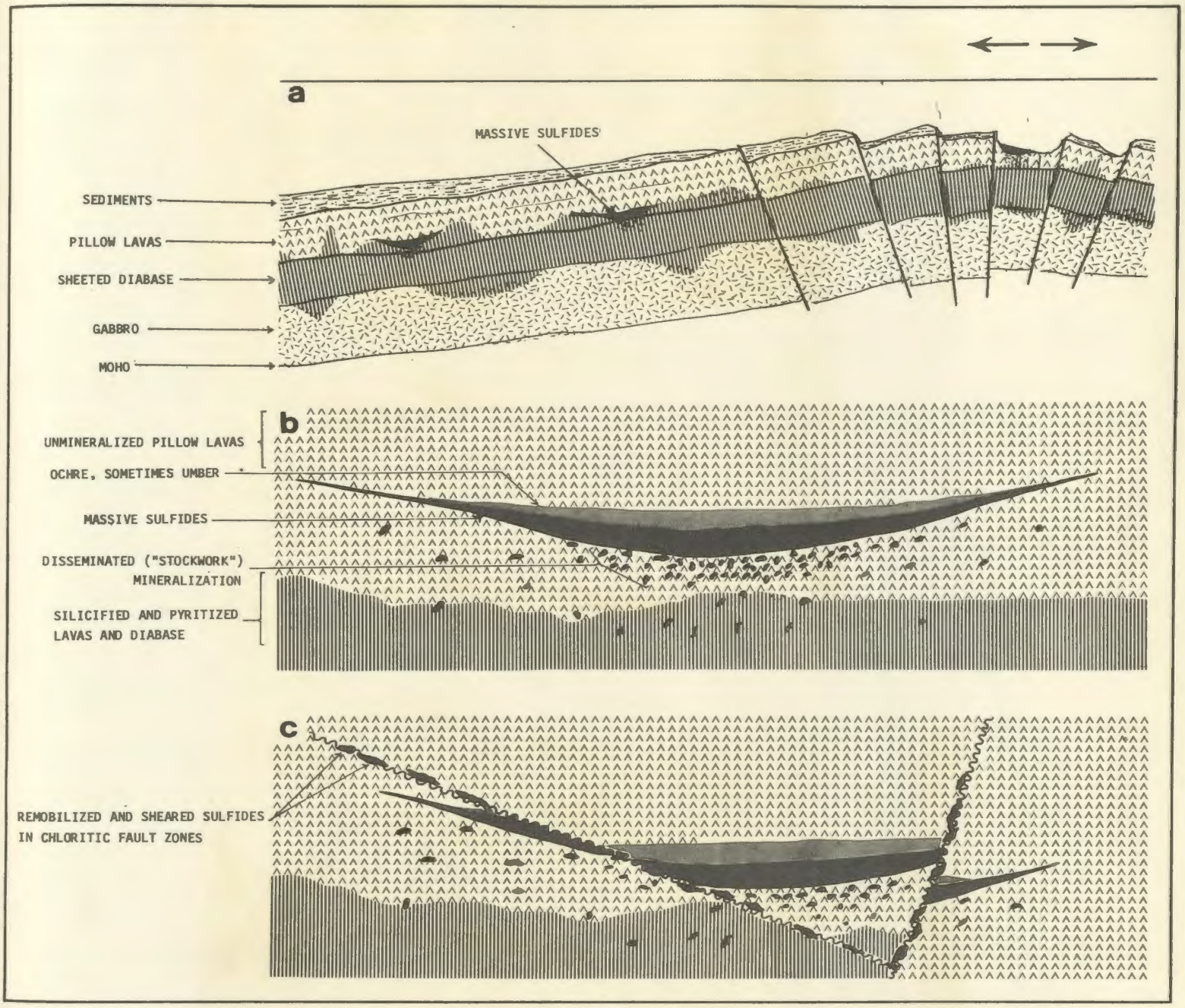


Figure 52: Proposed model for the formation of ophiolite-type massive sulfide deposits (modified after Hutchinson and Searle, 1971; Constantinou and Govett, 1972). *a*. Massive and "stockwork" mineralization produced by volcanic exhalative activity in an oceanic ridge-type environment during sea-floor spreading. Note the position of the sulfides at the base of the pillow lava sequence. *b*. Enlarged schematic cross section of a typical massive sulfide deposit showing sequence from "stockwork" mineralization to ochre. *c*. Deformed and remobilized equivalent of *b*, showing the emplacement of massive sulfides in chloritic fault zones, a feature common in Betts Cove and some other Newfoundland ophiolites.

Being rheologically incompetent, the sulfides would yield to shearing more easily than other material in the ophiolite suite. The concentration of sulfides along a particular level within the ophiolite, therefore, might have played an important role in localizing these faults.

E. DISCUSSION

In terms of overall geological environment, stratigraphic control, sedimentary features, and the simple pyrite-chalcopyrite-sphalerite mineralogy, the Betts Cove copper deposits show close similarity to those of the Troodos ophiolite (Hutchinson and Searle, 1971), although some features such as secondary submarine leaching (Constantinou and Govett, 1972) have not been studied in Betts Cove. The concentration of sulfides at the sheeted dyke/pillow lava contact typical for Cyprus (Searle, 1972), is also observed in several other Newfoundland ophiolites, notably Bay of Islands (R. W. Hutchinson, personal communication, 1972), Pilleys Island and Moretons Harbour (Strong, 1972) (Fig. 46). These similarities in mineralization suggest that the basic process of ore genesis is the same in all these ophiolites although minor differences may occur due to changing physico-chemical conditions from one place to another. The ophiolites, now generally recognized as obducted oceanic crusts and mantle, may provide the keys to understanding metallogenic processes operative in the modern oceans (Hutchinson and Searle, 1971; Sillitoe, 1972b; Upadhyay and Strong, 1973). Although massive sulfide deposits have not yet been found outside the continents, the existence of unusually high metal values in sediments on the ocean floors suggest that the massive

sulfides are emplaced at the oceanic ridges (Bostrom and Peterson, 1966).

A syngenetic origin has also been suggested for a number of sulfide deposits that occur in the predominantly volcanic terrains of Newfoundland (Williams, 1963; Peters, 1967; Upadhyay and Smitheringale, 1972; Strong and Peters, 1973). These deposits show a genetic similarity with those in the ophiolitic rocks, although in the latter the ore bodies are generally much better concentrated along certain stratigraphic contact zones.

REFERENCES

Aumento, F. 1968: The Mid-Atlantic Ridge near 45° N. II. Basalts from the area of Confederation Peak; Can. J. Earth Sci., 5, 1-21.

_____ 1969: Diorites from the Mid-Atlantic Ridge near 45° N.; Science, 165, 1112-1113.

_____ 1972: The oceanic crust of the Mid-Atlantic Ridge at 45° N.; Publ. Earth Physics Branch, Dept. Energy, Mines and Resources, Ottawa, 42(3), 49-53.

_____ and Loubat, H. 1971: The Mid-Atlantic Ridge near 45° N. XVI. Serpentinized ultramafic intrusions; Can. J. Earth Sci., 8, 631-663.

Averyanov, A. G., and others, 1961: Deep seismic sounding in the zone of transition from the Asian continent to the Pacific Ocean during the IGY, Izv. Akad. Nauk USSR, Ser. Geofiz., 169-184.

Ave Lallemant, H. G., and Carter, N. L. 1970: Syntectonic recrystallization of olivine and modes of flow in the upper mantle; Geol. Soc. Amer., Bull., 81, 2203-2220.

Bailey, Sir E. B., and McCallien, W. J. 1953: Serpentinite lavas, the Ankara Mélange and the Anatolian Thrust; Royal Soc. Edinburgh, Trans., 62, Pt. 2, 403-442.

Bailey, E. H., Blake, M. C., Jr., and Jones, D. L. 1970: Onland Mesozoic Oceanic crust in California Coast Ranges; U. S. Geol. Surv. Prof. Pap. 700-C, C70-C81.

Baird, D. M. 1951: The geology of the Burlington Peninsula, Newfoundland; Geol. Surv. Can. Pap. 51-21, 71 p.

_____ 1959: Massive sulfide deposits in Newfoundland; Can. Inst. Min. Met., Trans., 63, 39-42.

Barnes, Ivan, and O'Neil, J. R. 1969: The relationship between fluids in some fresh alpine-type ultramafics and possible modern serpentinization, western United States; Geol. Soc. Amer., Bull., 80, 1947-1960.

Bell, J. M., Clarke, E de C., and Marshall, P. 1911: The geology of the Dunn Mountain Subdivision, Nelson; New Zealand Geol. Surv. Bull. 12.

Benson, W. N. 1926: The tectonic conditions accompanying the intrusion of basic and ultrabasic igneous rocks; Nat. Acad. Sci., Mem. 19, 90 p.

Betz, F., Jr. 1948: Geology and mineral deposits of southern White Bay; Newfoundland Geol. Surv., Bull. 24, 26 p.

Bezzi, A., and Piccardo, G. B. 1971: Structural features of the Ligurian ophiolites; petrologic evidence for the "oceanic" floor of the northern Apennines geosyncline; a contribution to the problem of alpine-type gabbro-peridotite associations; Soc. Geol. Italiana, Mem., 10, 1-63.

Bilgrami, S. A., and Howie, R. A. 1960: Mineralogy and petrology of a rodingite dyke, Hindubagh, Pakistan; Amer. Mineral., 45, 791-801.

Bird, J. M., and Dewey, J. F. 1970: Lithosphere plate-continental margin tectonics and the evolution of the Appalachian orogen; Geol. Soc. Amer., Bull., 81, 1031-1060.

_____ Dewey, J. F., and Kidd, W. S. F. 1971: Proto-Atlantic oceanic crust and mantle: Appalachian/Caledonian ophiolites; Nature, Phys. Sci., 231, 2-31.

Bonatti, E. 1968: Ultramafic rocks from the Mid-Atlantic Ridge; Nature, 219, 363-364.

Bostrom, K., and Peterson, M. N. A. 1966: Precipitates from hydrothermal exhalations on the East Pacific Rise; Econ. Geol., 61, 1258-1265.

Bottcher, W. 1969: Zur Entstehung des magmatischen Troodos-komplexes (Zypern); N. Jb. Miner. Abh., 110(2), 159-187.

Bowen, N. L., and Tuttle, O. F. 1949: The System $MgO-SiO_2-H_2O$; Geol. Soc. Amer., Bull., 60, 439-460.

Brunn, J. H. 1954: Les eruptions ophiolitiques dans le NW de la Grece: leurs relations avec l'orogenese; 19th Int. Geol. Cong., Algiers, sec. XV, fasc. XVII, 19-27.

_____ 1960: Mise en place et differentiation de l'association plutovolcanique du cortege ophiolitique; Rev. Geogr. Phys. Geol. Dynam., 3, 115-132.

Cady, W. M., Albee, A. L., and Chidester, A. H. 1963: Bedrock geology and asbestos deposits of the upper Mississquoi Valley and vicinity, Vermont; U.S. Geol. Surv. Bull. 1122-B, B 1- B 78.

Cann, J. R. 1968: Geological processes at mid-ocean ridge crests; Geoph. J. Roy. Astr. Soc., 15, 331-341.

_____ 1969: Spilites from the Carlsberg Ridge, Indian Ocean; J. Petrol., 10, 1-19.

_____ 1970: New model for the structure of the ocean crust; Nature, 226, 928-930.

_____ and Funnel, B. M. 1967: Palmer Ridge: a section through the upper part of the ocean crust?; Nature, 213, 661-664.

_____ and Vine, F. J. 1966: An area on the crest of the Carlsberg Ridge: petrology and magnetic survey; Phil. Trans. Roy. Soc. Lond. A, 259, 198-219.

Carlisle, D. 1961: Pillow breccias and their aquagene tuffs, Quadra Island, British Columbia; J. Geol., 71, 48-71.

Carmichael, I. S. E. 1964: The petrology of Thingmuli, a Tertiary volcano in eastern Iceland; J. Petrol., 5, 435-460.

Chernysheva, V. L., and Berzrukov, P. L. 1966: Serpentinite from the crest of the Indo-Arabian Ridge; Dokl. Akad. Nauk. SSR, 166, 207-210.

Christensen, N. I. 1970: Composition and evolution of the oceanic crust; Marine Geol., 8, 139-154.

Church, W. R. 1969: Metamorphic rocks of the Burlington Peninsula and adjoining areas of Newfoundland, and their bearing on continental drift in North Atlantic; Amer. Assoc. Petrol. Geol., Mem. 12, 212-233.

_____ 1972: Ophiolite: its definition, origin as oceanic crust, and mode of emplacement in orogenic belts, with special reference to the Appalachians; Publ. Earth Physics Branch, Dept. Energy, Mines and Resources, Ottawa, 42(3), 71-85.

_____ and Stevens, R. K. 1971: Early Paleozoic ophiolite complexes of the Newfoundland Appalachians as mantle-oceanic crust sequences; J. Geoph. Res., 76, 1460-1466.

Coates, H. J. 1970: Structural and metamorphic history of the Pacquet Harbour - Grand Cove area, Burlington Peninsula, Newfoundland; unpublished M.Sc. thesis, Memorial University of Newfoundland, 79 p.

Coleman, R. G. 1966: New Zealand serpentinites and associated metasomatic rocks; New Zealand Geol. Surv., Bull., 76, 102 p.

Coleman, R. G. 1971a: Plate tectonic emplacement of upper mantle peridotites along continental edges; J. Geoph. Res., 76, 1212-1222.

_____ 1971b: Petrologic and geophysical nature of serpentinites; Geol. Soc. Amer., Bull., 82, 897-918.

Constantinou, G., and Govett, G. J. S. 1972: Genesis of sulfide deposits, ochre, and umber of Cyprus; Inst. Mining Met., Trans., Sec. B, 81, B34-B46.

Davies, H. L. 1968: Papuan Ultramafic Belt; 23rd Int. Geol. Cong., Prague, 1, 209-220.

_____ 1971: Peridotite-gabbro-basalt complex in eastern Papua: an overthrust plate of oceanic mantle and crust; Bureau of Mineral Resources, Geology and Geophysics, Australia, Bull. 128, 48 p.

Dean, W. T. 1970: Lower Ordovician trilobites from the vicinity of South Catcher Pond, northeastern Newfoundland; Geol. Surv. Can., Pap. 70-44, 15 p.

Deer, W. A., Howie, R. A., and Zussman, J. 1962: Rock-forming minerals, Vol. 3, Sheet silicates; John Wiley and Sons, Inc., New York.

Degens, E. T. and Ross, D. A. 1969 (Editors): Hot brines and recent heavy metal deposits in the Red Sea; Springler Verlag New York Inc.

DeLaunay, L. 1894: Les richesses minerales de Terre-Neuve; Annales des Mines, 1894, 395-400.

DeRoever, W. P. 1957: Sind sie Alpinotypen peridotit massen vielleicht tektonisch verfrachtete Bruchstucke der Peridotitschale?; Geol. Rundschau, 46, 137-146.

Deutsch, E. R., and Rao, K. V. 1972: Preliminary magnetic study of Newfoundland ophiolites, showing the presence of native iron (abs); Am. Geophys. Union, Trans., 53, p. 734.

Dewey, J. F. 1969: Evolution of the Appalachian/Caledonian orogen; Nature, 222, 124-129.

_____ and Bird, J. M. 1970: Mountain belts and the new global tectonics; J. Geophy. Res., 75, 2625-2647.

_____ and Bird, J. M. 1971: Origin and emplacement of the ophiolite suite: Appalachian ophiolites in Newfoundland; J. Geoph. Res., 76, 3179-3206.

Dickinson, W. R. 1971: Plate tectonic models of geosynclines; Earth Planet. Sci. Lett., 10, 163-174.

Dietz, R. S. 1963: Alpine serpentinites as oceanic rind fragments; Geol. Soc. Amer., Bull., 74, 947-952.

Donoghue, H. G., Adams, W. G. and Harpur, C. E. 1959: Tilt Cove copper operation of the Maritimes Mining Corporation Ltd., Can. Inst. Min. Met., Trans., 42, 54-73.

Douglas, G. V., Williams, D., Rove, O. N., and others, 1940: Copper deposits of Newfoundland; Newfoundland Geol. Surv., Bull. 20.

Engel, A. E. J., Engel, C. G., and Havens, R. G. 1965: Chemical characteristics of oceanic basalts and upper mantle; Geol. Soc. Amer., Bull., 76, 719-734.

Engel, C. G., and Fisher, R. L. 1969: Lherzolite, anorthosite, gabbro, and basalt dredged from the Mid-Indian Ocean Ridge; Science, 166, 1136-1141.

Ernst., W. G., Seki, Y., Onuki, H. and Gilbert, M. C. 1970: Comparative study of low-grade metamorphism in California Coast Ranges and the outer metamorphic belt of Japan; Geol. Soc. Amer., Mem. 124, 270 p.

Espenshade, G. H. 1937: Geology and mineral deposits of the Pilleys Island area; Newfoundland Geol. Surv., Bull. 6.

Fitton, J. G., and Hughes, D. J. 1970: Volcanism and plate tectonics in the British Ordovician; Earth Planet. Sci. Lett., 8, 223-228.

Gale, G. H. 1973: Paleozoic basaltic komatiite and ocean-floor type basalts from northeastern Newfoundland; Earth Planet. Sci. Lett., 18, 22-28.

Garland, J. 1888: Copper mining at Tilt Cove, Newfoundland; Roy. Geol. Soc. Cornwall, Trans., 11, 95-105.

Gass, I. G. 1958: Ultrabasic pillow lavas from Cyprus; Geol. Mag., 95, 241-251.

_____ 1968: Is the Troodos massif of Cyprus a fragment of Mesozoic ocean floor?; Nature, 220, 39-42.

_____ and Masson-Smith, D. M. 1963: The geology and gravity anomalies of the Troodos massif, Cyprus; Phil. Trans. Roy. Soc. Lond. A, 255, 417-467.

_____ and Smewing, J.D. 1973: Intrusion, extrusion, and metamorphism at constructive margins: evidence from Troodos massif, Cyprus; Nature, 242, 26-29.

Gill, J. E. 1969: Experimental deformation and annealing of sulfides and interpretation of ore textures; Econ. Geol., 64, 500-508.

Grange, L. I. 1927: On the rodingites of Nelson; New Zealand Inst. Trans. 58, 160-166.

Green, D. H. 1961: Ultramafic breccias from the Musa Valley, eastern Papua; Geol. Mag., 98, 1-26.

_____ 1964: The petrogenesis of the high-temperature peridotite intrusion in the Lizard area, Cornwall; J. Petrol., 5, 134-188.

_____ and Ringwood, A. E. 1967: The genesis of basaltic magmas; Contr. Miner. Petrol., 15, 102-190.

Greenbaum, D. 1972: Magmatic processes at ocean ridges: evidence from the Troodos massif, Cyprus; Nature Phys. Sci., 238, 18-21.

Hart, S. R. 1969: K, Rb, Cs, Ba contents and Sr isotope ratios of sea-floor basalts, and the nature of sea-water alteration effects (abs.); in Royal Society: abstracts for meeting on petrology of igneous and metamorphic rocks from the ocean floor; 12-14 Nov. 1969, London.

Hasebe, K., Fujii, N., and Uyeda, S. 1970: Thermal processes under island arcs; Tectonophysics, 10, 335-355.

Hekinian, R. 1968: Rocks from the mid-oceanic ridge in the Indian Ocean; Deep Sea Res., 15, 281-303.

Hersey, J. B. 1965: Sedimentary basins of the Mediterranean Sea; in Submarine Geology and Geophysics, Colston Papers, edited by W. F. Whitard and R. Badshaw, 75-91, Butterworths, London.

Hess, H. H. 1962: History of ocean basins; in Petrologic Studies: A Volume in Honour of A. F. Buddington, edited by A. E. J. Engel and B. F. Leonard, Geol. Soc. Amer., 599-620.

_____ 1964: The oceanic crust, the upper mantle and the Mayaguez serpentinized peridotite; in A Study of Serpentinite, edited by C. A. Burke, 169-175, Natl. Acad. Sci.-Natl. Res. Council Publ.

_____ 1965: Mid-oceanic ridges and tectonics of the sea-floor; in Submarine Geology and Geophysics, Closton Papers, edited by W. F. Whitard and R. Bradshaw, 317-334, Butterworth, London.

Horne, G. S. 1969: Early Ordovician chaotic deposits in the central volcanic belt of northeastern Newfoundland; Geol. Soc. Amer., Bull., 80, 2451-2464.

Horne, G. S. and Helwig, J. 1969: Ordovician stratigraphy of Notre Dame Bay, Newfoundland; Amer. Assoc. Petrol. Geol., Mem. 12, 388-407.

Hughes, C. J. 1972: Spilites, keratophyres, and the igneous spectrum; Geol. Mag., 109, 513-527.

Hutchinson, R. W., and Searle, D. R. 1971: Stratabound pyrite deposits in Cyprus and relations to other sulfide ores; Soc. Mining Geol. Japan, Spec. Issue 3, 198-205.

Irvine, T. N. and Baragar, W. R. A. 1971: A guide to the chemical classification of the common volcanic rocks; Can. J. Earth Sci., 8, 523-548.

Jakes, P. and Gill, J. 1970: Rare earth elements and the island arc tholeiite series; Earth Planet. Sci. Lett., 9, 17-28.

Jenness, S. E. 1958: Geology of the Lower Gander River ultrabasic belt, Newfoundland; Newfoundland Geol. Surv., Rept. 12, 58 p.

Kalliokoski, J. 1953: Springdale, Newfoundland; Geol. Surv. Can., Pap. 53-5, 4 p.

Karig, D. E. 1971a: Origin and development of marginal basins in the western Pacific; J. Geoph. Res., 76, 2542-2561.

_____ 1971b: Structural history of the Mariana Island arc system; Geol. Soc. Amer., Bull, 82, 323-344.

Kay, M. 1967: Stratigraphy and structure of northeastern Newfoundland bearing on drift in North Atlantic; Amer. Assoc. Petrol. Geol., Mem. 12, 579-600.

Kay, R., Hubbard, N. J., and Gast, P. W. 1970: Chemical characteristics and origin of oceanic ridge volcanic rocks; J. Geoph. Res., 75, 1585-1613.

Kean, B. F. 1973: Stratigraphy, petrology and geochemistry of volcanic rocks of Long Island, Newfoundland; unpublished M.Sc. thesis, Memorial University of Newfoundland, 155 p.

Kennedy, M. J. 1971: Structure and stratigraphy of the Fleur de Lys Supergroup in the Fleur de Lys area, Burlington Peninsula, Newfoundland; Proc. Geol. Assoc. Can., 24, 59-72.

_____ 1973: Pre-Ordovician polyphase structure in the Burlington Peninsula of the Newfoundland Appalachians; Nature, Phys. Sci., 241, 114-116.

Kennedy, M. J. and McGonigal, M. H. 1972: The Gander Lake and Davidsville Groups of northeastern Newfoundland: new data and geotectonic implications; Can. J. Earth Sci., 9, 452-459.

_____ and Phillips, W. E. 1971: Ultramafic rocks of the Burlington Peninsula, Newfoundland; Proc. Geol. Assoc. Can., 24, 35-46.

Kitahara, S., Takenouchi, S., and Kennedy, G. C. 1966: Phase relations in the system $MgO-SiO_2-H_2O$ at high temperatures and pressures; Amer. J. Sci., 264, 223-233.

Kominskaya, I. P., Belyaevsky, N. A., and Volvovsky, I. S. 1969: Explosion seismology in the USSR; in The Earth's Crust and Upper Mantle edited by P. J. Hart, Amer. Geoph. Union, monograph 13, 195-208.

Lamarche, R. Y. 1972: Ophiolites of southern Quebec; Publ. Earth Physics Branch, Dept. Energy, Mines and Resources, Ottawa, 42(3), 65-70.

Le Pichon, X. 1969: Models and structure of the oceanic crust; Tectonophysics, 7, 385-401.

Lotti, B. 1886: Descrizione Geologica dell' Isola d'Elba; Mem. Descrit. Carta Geol. Ital., 2.

MacDonald, G. A., and Katsura, T. 1964: Chemical composition of Hawaiian lava; J. Petrol., 5, 82-133.

MacLean, H. J. 1947: Geology and mineral deposits of the Little Bay area, Newfoundland; Newfoundland Geol. Surv., Bull. 22.

MacKenzie, D. B. 1960: High temperature alpine-type peridotite from Venezuela; Geol. Soc. Amer., Bull., 71, 303-318.

Malpas, J. G. 1973: A restored section of oceanic crust and mantle in western Newfoundland (abs.); Geol. Soc. Amer. Northeastern Sec. 8th Ann. Mtg. Abstracts 5, p. 1025.

_____ Stevens, R. K., and Strong, D. F. 1973: Amphibolites associated with Newfoundland ophiolites: their classification and tectonic significance; Geol. Soc. Amer., Bull., in press.

Marten, B. E. 1971: Stratigraphy of volcanic rocks in the Western Arm area of the central Newfoundland Appalachians; Proc. Geol. Assoc. Can., 24, 73-84.

Matsuda, T., and Uyeda, S. 1971: On the Pacific-type orogeny and its model-extension of the paired belts concept and possible origin of marginal seas; Tectonophysics, 11, 5-27.

Maxwell, J. C. 1969: "Alpine" mafic and ultramafic rocks--the ophiolite suite: a contribution to the discussion of the paper "The origin of ultramafic and ultrabasic rocks" by P. J. Wyllie; Tectonophysics, 7, 489-494.

_____ and Azzaroli, A. 1962: Submarine extrusion of ultramafic magma (abs.); Geol. Soc. Amer., 1962 ann. met., p. 103A.

McTaggart, K. C. 1971: On the origin of ultramafic rocks; Geol. Soc. Amer., Bull., 82, 23-42.

Melson, W. G. 1969: Ultramafic rocks from the Mid-Atlantic fracture zones: implications on the make-up of the lower oceanic crust (abs.); in Royal Society: abstracts for meeting on petrology of igneous and metamorphic rocks from the ocean floor, 12-14 Nov. 1969, London.

_____ Jarosewich, G., Bowen, V. T., and Thompson, G. 1967: St. Peter and St. Paul rocks: A high-temperature, mantle-derived intrusion; Science, 155, 1532-1535.

_____ and Van Andel, T. H. 1966: Metamorphism in the Mid-Atlantic Ridge, 22°N latitude; Marine Geol., 4, 165-186.

Menard, H. W. 1967: Transitional types of crust under small ocean basins; J. Geoph. Res., 72, 3061-3073.

Milliard, Y. 1959: Les massifs metamorphiques et ultrabasique de la zone palaeozoique interne du Rif; Notes Serv. Geol. Maroc., No. 147, 125-160.

Mitchell, A. H. G. 1970: Facies of an early Miocene volcanic arc, Malekua Island, New Hebrides; Sedimentology, 14, 201-243.

_____ and Reading, H. G. 1969: Continental margins, geosynclines and ocean floor spreading; J. Geol., 77, 629-646.

_____ and Reading, H. G. 1971: Evolution of island arcs; J. Geol., 79, 253-284.

Miyashiro, A., Shido, F., and Ewing, M. 1969: Composition and origin of serpentinites from the Mid-Atlantic Ridge near 24° and 30° N latitudes; Contrib. Mineral. Petrol., 23, 117-127.

Moores, E. M. 1969: Petrology and structure of the Vourinos ophiolitic complex of northern Greece; Geol. Soc. Amer., Spec. Pap. 118, 74 p.

_____ 1970: Ultramafics and orogeny, with models of the U.S. Cordillera and the Tethys; Nature, 228, 837-842.

Moores, E. M., and Vine, F. J. 1971: The Troodos massif, Cyprus and other ophiolites as oceanic crust: evaluation and implications; Phil. Trans. Roy. Soc. Lond. A, 268, 443-466.

Neale, E. R. W. 1957: Ambiguous intrusive relationship of the Betts Cove - Tilt Cove serpentinite belt, Newfoundland; Proc. Geol. Assoc. Can., 9, 95-107.

_____ 1958: Nippers Harbour, Newfoundland; Geol. Surv. Can., Map 22.

_____ and Kennedy, M. J. 1967: Relationship of the Fleur de Lys Group to younger groups of the Burlington Peninsula, Newfoundland; Geol. Assoc. Can., Spec. Pap. 4, 139-169.

_____ and Nash, W. A. 1963: Sandy Lake, (east half) map area, Newfoundland; Geol. Surv. Can. Pap. 62-28, 40 p.

Nicholls, G. D., Nalwalk, A. J., and Hayes, E. E. 1964: The nature and composition of rock samples dredged from the Mid-Atlantic Ridge between 22°N and 52°N; Marine Geol. 1, 333-343.

Nicolas, A. 1966: Etude pétrochimique des roches vertes et de leurs mineraux entre Dora Maira et Grand Paradis (Alpes piémontaises); Doctoral thesis, Nantes, Vol. 1.

_____ Bouchez, J. L., Boudier, F., and Mercier, J. C. 1971: Textures, structures and fabrics due to solid state flow in some European lherzolites; Tectonophysics, 12, 56-86.

Nockolds, S. R. 1954: Average chemical compositions of some igneous rocks; Geol. Soc. Amer., Bull., 65, 1007-1032.

O'Hara, M. J. 1963: Melting of garnet peridotite at 30 kilobars; Carnegie Inst. Wash. Yr. Book, 62, 71-76.

_____ 1967: Mineral facies in ultrabasic rocks; in Ultramafic and Related Rocks, edited by P. J. Wyllie, 7-18, John Wiley, New York.

Page, N. J. 1966: Mineralogy and chemistry of the serpentine group minerals and the serpentinization process; Ph.D. thesis, University California, Berkley.

Pakham, G. H., and Falvey, D. A. 1971: An hypothesis for the formation of marginal seas in the western Pacific; Tectonophysics, 11, 79-109.

Papezik, V. S. 1964: Nickel minerals at Tilt Cove, Notre Dame Bay, Newfoundland; Proc. Geol. Assoc. Can., 15, 27-32.

Papezik, V. S., and Fleming, J. M. 1967: Basic volcanic rocks of the Whalesback area, Newfoundland; Geol. Assoc. Can., Spec. Pap. 4, 181-192.

Parsons, W. H. 1967: Criteria for the recognition of volcanic breccias: review; Geol. Soc. Amer., Mem. 115, 263-304.

Peters, H. R. 1967: Mineral deposits of the Halls Bay area, Newfoundland; Geol. Assoc. Can., Spec. Pap. 4, 171-180.

Peterson, D. W. 1960: Descriptive modal classification of igneous rocks; Geotimes, 5, 30-36.

Poldervaart, A., and Taunebeck, W. H. 1960: Layered intrusions; 21st Int. Geol. Cong., Stockholm, 13, 239-246.

Pyke, D. R., Naldrett, A. J., and Eckstrand, O. R. 1973: Archean ultramafic flows in Munro Township, Ontario; Geol. Soc. Amer., Bull., 84, 955-978.

Reinhardt, B. M. 1969: On the genesis and emplacement of ophiolites in the Oman Mountains Geosyncline; Schweiz. Min. Petrog. Mitt., 49, 1-30.

Reynolds, D. L. 1954: Fluidization as a geological process, and its bearing on the problem of intrusive granites; Amer. J. Sci., 252, 577-613.

Rodgers, J. 1965: Long Point and Clam Bank formations, western Newfoundland; Proc. Geol. Assoc. Can., 16, 83-94.

_____ and Neale, E. R. W. 1963: Possible "taconic" klippe in western Newfoundland; Amer. J. Sci., 261, 713-730.

Sampson, E. 1923: The ferruginous chert formations of Notre Dame Bay, Newfoundland; J. Geol., 31, 571-598.

Searle, D. L. 1972: Mode of occurrence of the cupriferous pyrite deposits of Cyprus; Inst. Mining Met., Trans., Sec. B, 81, B189-B197.

_____ and Vokes, F. M. 1969: Layered ultrabasic lavas from Cyprus; Geol. Mag., 106, 515-530.

Shor, G. G. 1964: Structure of the Bering Sea and the Aleutian ridge; Marine Geol., 1, 213-219.

Sillitoe, R. H. 1972a: Relation of metal provinces in western America to subduction of oceanic lithosphere; Geol. Soc. Amer., Bull., 83, 813-817.

Sillitoe, R. H. 1972b: Formation of certain massive sulfide deposits at sites of sea-floor spreading; Inst. Mining Met., Trans., Sec. B, 81, B141-B148.

Sleep, N., and Toksöz, M. N. 1971: Evolution of marginal basins; Nature, 233, 548-550.

Smith, C. H. 1958: Bay of Islands igneous complex, western Newfoundland; Geol. Surv. Can., Mem. 290, 132 p.

_____ and MacGregor, I. D. 1960: Ultrabasic intrusive conditions illustrated by the Mount Albert ultrabasic pluton, Gaspe, Quebec (abs.); Geol. Soc. Amer., Bull. 71, p. 1978.

Smitheringale, W. G. 1972: Low-potash Lush's Bight tholeiites: ancient oceanic crust in Newfoundland?; Can. J. Earth Sci., 9, 574-588.

Snelgrove, A. K. 1928: The geology of the central mineral belt of Newfoundland; Can. Min. Met., Bull. 197, 1057-1127.

_____ 1931: Geology and ore deposits of Betts Cove - Tilt Cove area, Newfoundland; Can. Min. Met., Bull. 228, 477-519.

Steinmann, G. 1927: Die ophilothischen zonen in dem mediterranean Kettengebirgen; 14th Int. Geol. Cong., Madrid, 2, 638-667.

Stevens, R. K. 1970: Cambro-Ordovician flysch sedimentation and tectonics in west Newfoundland and their possible bearing on a Proto-Atlantic ocean; Geol. Assoc. Can., Spec. Pap. 7, 165-177.

Strong, D. F. 1972: Sheeted diabases of central Newfoundland: new evidence for Ordovician sea-floor spreading; Nature, 235, 102-104.

_____ 1973a: Plate tectonic setting of Newfoundland mineral deposits; Min. Deposita, in press.

_____ 1973b: A geological and geochemical comparison of the Lush's Bight and Roberts Arm Groups of central Newfoundland: possible juxtaposed oceanic and island arc volcanic suites; Geol. Soc. Amer., Bull., in press.

_____ and Payne, J. G. 1973: Early Paleozoic volcanism and metamorphism of the Moretons Harbour - Twillingate area, Newfoundland; Can. J. Earth Sci., in press.

_____ and Peters, H. R. 1973: The importance of volcanic setting for base metal exploration in central Newfoundland; Can. Inst. Min. Met., Trans., in press.

Suffel, G. S., Hutchinson, R. W., and Ridler, R. H. 1971: Metamorphism of massive sulfides at Manitouwadge, Ontario, Canada; Soc. Mining Geol. Japan, Spec. Issue 3, 235-240.

Szekely, J. 1971: Dike filling by magma intrusion and by explosive entrainment of fragments; J. Geoph. Res., 76, 2602-2609.

Thayer, T. P. 1960: Some critical differences between alpine-type and stratiform peridotite-gabbro complexes; 21st Int. Geol. Cong., Stockholm, 13, 247-259.

_____ 1966: Serpentinization considered as a constant volume metasomatic process; Amer. Mineral., 51, 685-710.

_____ 1967: Chemical and structural relations of ultramafic and felspathic rocks in alpine intrusive complexes; in Ultramafic and Related Rocks, edited by P. J. Wyllie, 222-239, John Wiley, New York.

_____ 1969: Alpine-type sensu stricto (ophiolitic) peridotites: refractory residues from partial melting or igneous sediments?: a contribution to the discussion of the paper "The origin of ultramafic and ultrabasic rocks" by P. J. Wyllie; Tectonophysics, 7, 511-516.

_____ and Himmelberg, G. R. 1968: Rock succession in the alpine-type mafic complex at Canyon Mountain, Oregon; 23rd Int. Geol. Cong., Prague, 1, 175-186.

Thompson, R. B. 1968: Serpentinization accompanied by volume changes; Pacific Geol., Pt. 1, 167-174.

Turner, F., and Verhoogan, J. 1960: Igneous and metamorphic petrology 2nd ed., McGraw-Hill, New York, 694 p.

Udintsev, G. B. 1969: The tectonics of the Mid-Indian Ridge and the petrography of the solid rocks of its rift zones; in Royal Society, abstracts for meeting on igneous and metamorphic petrology of rocks dredged from the ocean floor, 12-14 Nov. 1969, London.

Upadhyay, H. D. 1973: Volcanic rocks of the Snook s Arm Group and its basal ophiolite suite, northeast Newfoundland (abs.); Can. Geoph. Bull., in press.

_____ Dewey, J. F., and Neale, E. R. W. 1971: The Betts Cove ophiolite complex, Newfoundland: Appalachian oceanic crust and mantle; Proc. Geol. Assoc. Can., 24, 27-34.

_____ and Smitheringale, W. G. 1972: Geology of the Gullbridge copper deposit, Newfoundland: volcanogenic sulfides in cordierite-anthophyllite rocks; Can. J. Earth Sci., 9, 1061-1073.

Upadhyay, H. D., and Strong, D. F. 1973: Geological setting of the Betts Cove copper deposits, Newfoundland: an example of ophiolite sulfide mineralization; Econ. Geol., 68, 161-167.

Van der Linden, W. J. M. 1969: Extinct mid-ocean ridges in the Tasman Sea and in the western Pacific; Earth Planet. Sci. Lett., 8, 483-490.

Varne, R., Gee, R. D., and Guilty, P. G. J. 1969: Macquarie Island and the cause of oceanic linear magnetic anomalies; Science, 166, 230-232.

Viljoen, M. J., and Viljoen, R. P. 1969a: Evidence for the existence of a mobile extrusive peridotite magma from the Komati Formation of the Onverwacht Group; in Geol. Soc. S. Africa, Spec. Publ. 2, 87-112.

_____ and Viljoen, R. P. 1969b: The geology and geochemistry of the lower ultramafic unit of the Onverwacht Group and a proposed new class of igneous rock; in Geol. Soc. S. Africa, Spec. Publ. 2, 55-86.

Vine, F. J. 1966: Spreading of the ocean floor: new evidence; Science, 154, 1405-1415.

Vuagnat, M. 1963: Remarques sur las trilogie serpentinites-gabbros-diabases dans le bassin de la mediterranée occidentale; Geol. Rdsch., 53, 336-358.

_____ and Cogulu, E. 1967: Quelques réflexions sur le massif basique-ultrabasique due Kizel Dagh, Hatay, Turquie; Soc. Phys. Hist. Natu. Genéve, C.R. 2, 210-216.

Wadsworth, M. E. 1884: Notes on the rocks and ore deposits in the vicinity of Notre Dame Bay, Newfoundland; Amer. J. Sci., 28, 94-104.

Wager, L. R., and Brown, G. M. 1968: Layered igneous rocks; Freeman Press, San Francisco.

_____ Brown, G. M., and Wadsworth, W. J. 1960: Types of igneous cumulates; J. Petrol., 1, 73-85.

Watson, K. de P. 1947: Geology and mineral deposits of Baie Verte-Mings Bight area; Newfoundland Geol. Surv., Bull. 21.

Wenner, D. B., and Taylor, H. P., Jr. 1969: δD and δO^{18} studies of serpentinization of ultramafic rocks (abs.); Geol. Soc. Amer., Abstr. Programs for 1969, Pt. 7, 234-235.

Williams, H. 1962: Botwood (west half) map area, Newfoundland; Geol. Surv. Can., Pap. 62-9, 16 p.

_____ 1963: Relationship between base metal mineralization and volcanic rocks in northeastern Newfoundland; Can. Mining Jour., 84, 39-42.

_____ 1964: The Appalachians in northeast Newfoundland--a two-sided symmetrical system; Amer. J. Sci., 262, 1137-1158.

_____ 1967a: Silurian rocks of Newfoundland; Geol. Assoc. Can., Spec. Pap. 4, 93-137.

_____ 1967b: Island of Newfoundland; Geol. Surv. Can., Map 1231A.

_____ 1971: Mafic ultramafic complexes in western Newfoundland Appalachians and evidence for their transportation: a review and interim report; Proc. Geol. Assoc. Can., 24, 9-25.

_____ Kennedy, M. J., and Neale, E. R. W. 1972: The Appalachian structural province; in Variations in Tectonic Styles in Canada, edited by R. A. Price and R. J. W. Douglas, Geol. Assoc. Can., Spec. Pap. 11, 181-261.

_____ and Malpas, J. G. 1972: Sheeted dykes and brecciated dyke rocks within transported igneous complexes, Bay of Islands, western Newfoundland; Can. J. Earth Sci., 9, 1216-1229.

_____ and Smyth, W. R. 1973: Metamorphic aureoles beneath ophiolite suites and alpine peridotites: tectonic implications with west Newfoundland examples; Amer. J. Sci., in press.

Wilson, J. T. 1966: Did the Atlantic close and then re-open?; Nature, 211, 676-681.

Winkler, H. G. F. 1967: Petrogenesis of metamorphic rocks; Springer-Verlag New York Inc., 237 p.

Wyllie P. J. 1967 (editor): Ultramafic and related rocks; John Wiley, New York, 464 p.

APPENDIX

PETROGRAPHICAL DESCRIPTIONS OF ANALYSED ROCK SAMPLES

Sample Number	Description
1	Serpentinized lherzolite from the layered ultramafic sequence; shows cumulate texture; consists of highly altered olivine and orthopyroxene with minor fresh clinopyroxene.
2	Serpentinite from Partridge Pond; consists almost entirely of mesh-structure serpentine with minor bastite; accessory chromite.
3	Greenish grey clinopyroxenite from the transition zone, west of Joey's Pond; consists of clinopyroxene which is only slightly uralitized.
4	Serpentinized brown-weathering ultramafic rock with pillow-like shape, from near Partridge Pond; consists of a very finely-intergrown carbonate, serpentine and minor talc with some distinct patches of mesh-structure serpentine; accessory chromite present.
5	As above, except less carbonate.
6	Same as sample 4.
7	Light red weathering ultramafic (perknite) pillow; consists entirely of tiny laths of actinolite and tremolite that are secondary after clinopyroxene.
8	As above.
9	As above, but light coloured on fresh surfaces, coarser texture; accessory chromite present.
10	As above.
11	Light red weathering felsic perknite pillow lava; consists of tiny laths of actinolite and tremolite, with minor quartz.
12	As above.

Sample Number	Description
13	Same as sample 11, but minor carbonate and epidote present.
14	Same as sample 11.
15	Same as sample 11, except the presence of minor epidote and chromite.
16	Same as sample 11, but contains isolated patches of chlorite and quartz.
17	Same as sample 11, except the occurrence of more distinct patches of quartz, chlorite and epidote.
18	Same as sample 11, but contains minor epidote and chlorite.
19	Mafic pillow lava consisting of tremolite-actinolite and quartz with minor epidote.
20	Mafic pillow lava comprising actinolitic amphibole, sodic plagioclase and minor quartz and epidote.
21	Mafic pillow lava consisting of sodic plagioclase and amphibole; plagioclase moderately saussuritized and amphibole completely altered to chlorite.
22	Mafic pillow lava showing a fine intergrowth of amphibole, sodic plagioclase, quartz and epidote with patches of chlorite; alteration causes a turbid appearance of the rock in thin section.
23	Same as sample 20.
24	Same as sample 20.
25.	Same as sample 20, except a higher proportion of chlorite.
26	Mafic pillow lava consisting of granular, relatively fresh clinopyroxene and saussuritized, turbid plagioclase; large patches of chlorite common.
27	Mafic pillow lava consisting of tiny laths of plagioclase and completely chloritized pyriboles; fresh plagioclase grains occur as phenocrysts; abundant opaque material present.

Sample Number	Description
28	Mafic pillow lava consisting of uralitized clinopyroxene, plagioclase, and chlorite.
29	Gabbro screen within sheeted dykes; consists of uralite, subordinate plagioclase and quartz and accessory potash felspar.
30	Gabbro from the upper levels of Gabbroic Member; same as sample 29, but clinopyroxene is less uralitized; accessory potash felspar.
31	Coarse-grained diabase dyke cutting layered ultramafic sequence; comprises saussuritized plagioclase and uralite, with minor quartz.
32	Red-weathering felsic perknite dyke; consists predominantly of actinolitic amphibole and minor quartz; epidote occurs as an accessory mineral.
33	Fine-grained diabase dyke; constituent minerals are uralitic amphibole, minor plagioclase and quartz; chlorite, carbonate and epidote are also present.
34	Diabase dyke consisting of actinolitic amphibole, sodic plagioclase and minor quartz.
35	Quartz-rich felsic perknite dyke consisting of turbid-looking uralite and subordinate quartz; uralite partly altered to chlorite.
36	Fine-grained pillow lava; consists of highly turbid matte of uralitized clinopyroxene and plagioclase; carbonate, chlorite and epidote present; trachytic texture well developed.
37	As above, but shows subophitic, coarser texture.
38	Same as sample 36, except coarser texture and less altered plagioclase.
39	Same as sample 36.
40	Same as sample 36, but less turbid and contains isolated saussuritized plagioclase phenocrysts.

Sample Number	Description
41	Pillow lava sample consisting of a nearly opaque mixture containing phenocrysts of fresh plagioclase and clinopyroxene.
42	Same as sample 36, but much less turbid and shows subophitic texture; cut by a vein that carries quartz, chlorite and epidote.
43	Pillow lava sample comprising highly chloritized and uralitized clinopyroxene, saussuritized plagioclase, and minor quartz; abundant granules of an opaque mineral.
44	Core of a pillow; consists of partly saussuritized plagioclase and chloritized/uralitized clinopyroxene; both also occur as phenocrysts; magnetite-like opaque mineral in abundance.
45	Core of a pillow; contains moderately saussuritized plagioclase and uralitized clinopyroxene; few isolated phenocrysts of plagioclase; subophitic texture; chlorite and carbonate occur as accessory minerals.
46	Very dark pillow lava sample; consists of a turbid and semi-opaque mixture of indeterminate material that contains abundant submicroscopic granules of opaque material; plagioclase phenocrysts of various sizes present.
47	Porphyritic pillow lava containing plagioclase phenocrysts in a plagioclase-uralite matrix; the phenocrysts are more heavily saussuritized than the matrix plagioclase.
48	Porphyritic pillow lava consisting of fresh clinopyroxene and plagioclase; subophitic texture; plagioclase phenocrysts altered to semi-opaque saussurite.
49	Pillow lava containing highly chloritized clinopyroxene and saussuritized plagioclase with tiny granules of opaques; patches of carbonate and massive chlorite present.
50	Pillow lava comprising fresh clinopyroxene and plagioclase, with subophitic texture.

Sample Number	Description
51	As above, except clinopyroxene more altered.
52	Dark pillow lava consisting of slender plagioclase laths in a turbid uralitic groundmass; trachytic and porphyritic textures present; abundant magnetite.
53	Pillow lava consisting of moderately saussuritized plagioclase and completely uralitized/chloritized clinopyroxene, minor opaques; cut by a few epidote-bearing veins.

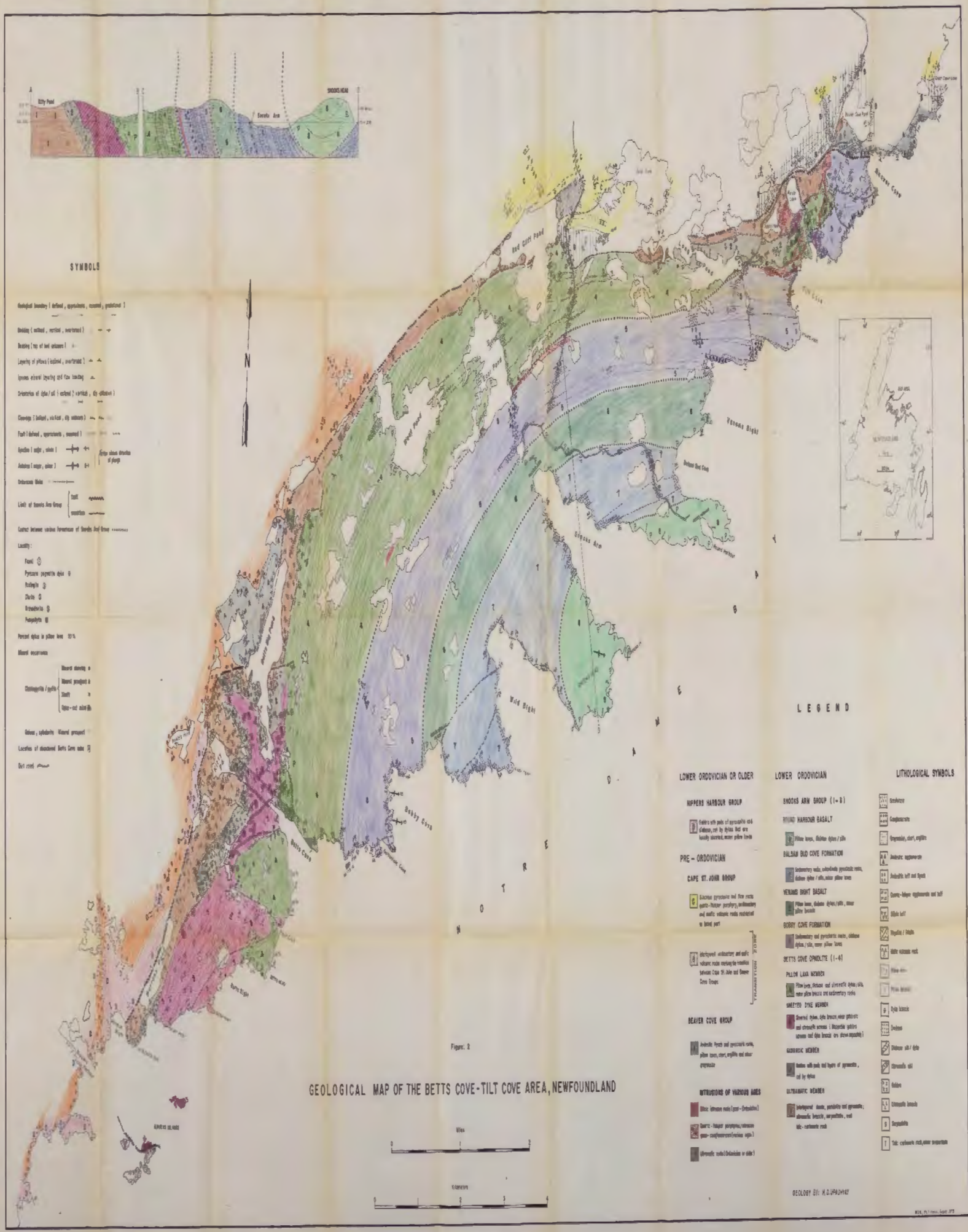
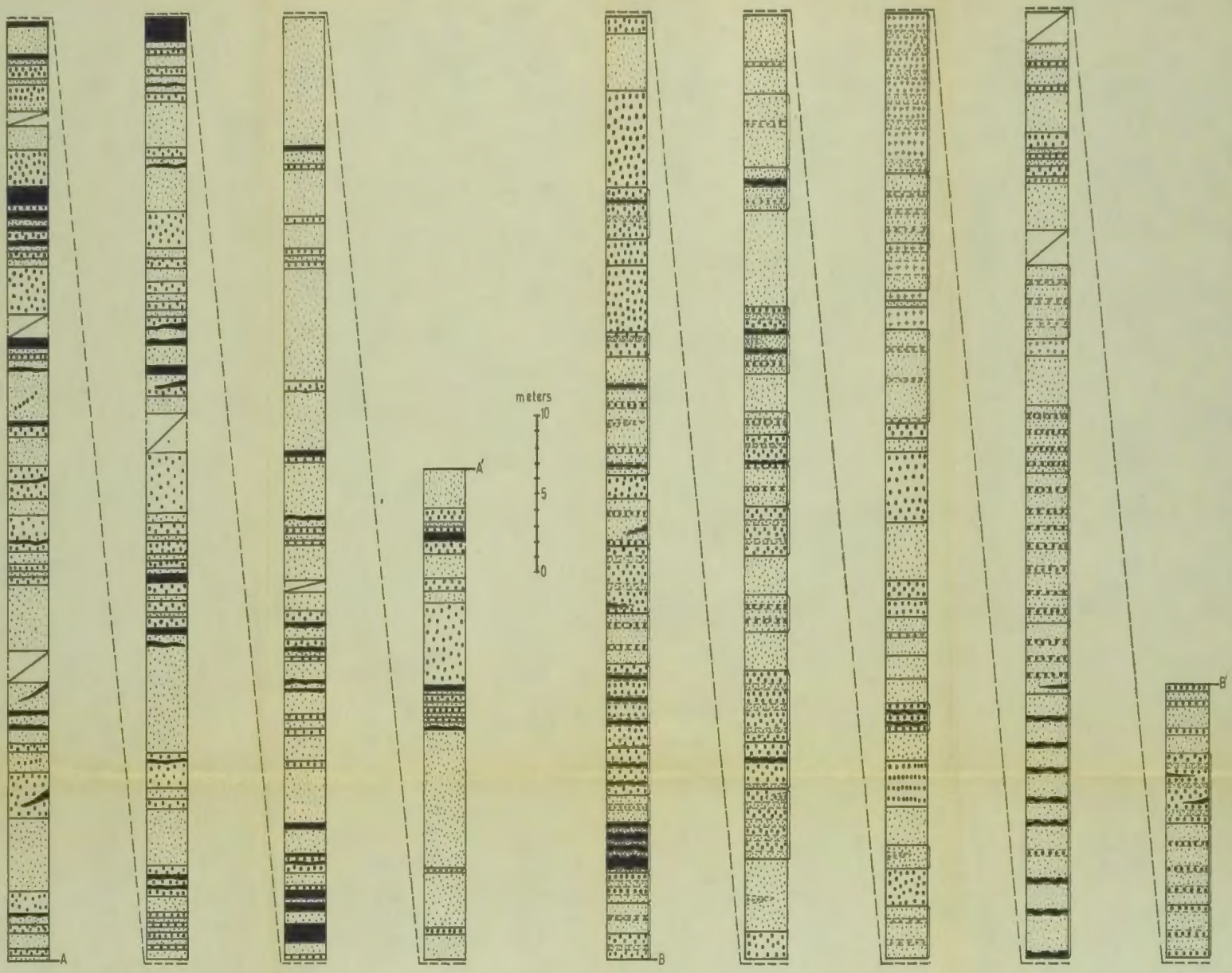
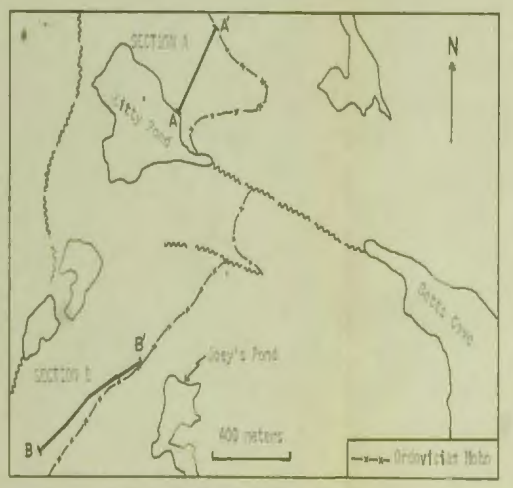


Figure: 2

GEOLOGICAL MAP OF THE BETTS COVE-TILT COVE AREA, NEWFOUNDLAND





LOCATION OF SECTION LINES

LEGEND

- Gabbro
- Pyroxenite
- Peridotite
- Dunite
- Percentage of rock types visually estimated; position of layers schematic.
- Covered interval

Figure 9: Columnar sections of the layered ultramafic rocks of the Betts Cove ophiolite, Newfoundland.